Kansas Geological Survey

THE HYDROGEOLOGY OF CRYSTAL SPRING WITH AND DELINEATION OF ITS SOURCE WATER ASSESSMENT AREA

By

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Kansas Geological Survey Open File Report 2003-35
October 2003

GEOHYDROLOGY

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EXECUTIVE SUMMARY

Introduction

The Safe Drinking Water Act (SDWA 1996), as amended by Congress in 1996, established the Source Water Protection Program (SWPP). This program requires all states to submit individual public water supply Source Water Assessment Plans (SWAPs) to the US Environmental Protection Agency for approval and subsequent implementation by the state. The SWPP of the Kansas Department of Health and Environment (KDHE) closely follows the program structure laid out in the SDWA 1996 and is based on a three-way partnership among KDHE, the water supply, and technical assistance providers (KDHE, 1999). An integral part of the SWPP is the requirement for a delineation of each public water supply's source water assessment area (SWAA). The Kansas Geological Survey (KGS) conducts research on the water resources of Kansas and has the expertise to provide technical assistance to public water suppliers should it be needed in the process of SWAA delineation.

Crystal spring is the sole source of water for the city of Florence, located in the central part of the Flint Hills region of Kansas. The spring is fed by ground water that discharges from a limestone aquifer. Prior to this study there was little information available on which to define its catchment area. To fill this gap, the KGS proposed conducting a two-year study to delineate the spring catchment (source water assessment) area to the Kansas Source Water Protection Program of KDHE. The two goals for this project were to:

- Complete a hydrogeologic study of the Crystal spring catchment and its surrounding area, and
- Delineate the SWAA for Crystal spring based on the results of the hydrogeologic investigation.

The methods used to complete the hydrogeologic study of the project area included field mapping of the surface geology, inventory of springs and wells, water-level surveys, water sample collection and chemical analysis, surveys of the streambed in Martin and Bruno creeks, dye trace experiments, and automated high frequency water-level data collection in a monitoring well east of Martin Creek.

Catchment Hydrogeology and Water Quality

Alternating limestones and shales of the Permian age Chase Group crop out at the surface in the project area. The limestones are considered to be aquifers because they are more permeable than the shales and yield water to wells in the area. Within the project area, the primary aquifer zones are in the Barneston Limestone, which is approximately 80 ft (24.4 m) thick. The Barneston consists primarily of limestone, which is soluble in water moving downward from the surface. Acting over long periods of time, the continued movement of ground water through the limestones has created an integrated network of solution-widened fractures and conduits of varying size. The scale of these features ranges widely from fractures with apertures that are tenths of an inch (several millimeters) wide up to master conduits up to several feet (1 m) in diameter.

During this study, measured spring discharge ranged from slightly less than 28.3 L/s up to more than 510 L/s (1 ft^3/s up to more than 18 ft^3/s) and at least half of the time the discharge was 73.6 L/s (2.6 ft^3/s). Spring discharge and turbidity are lower but dissolved solids concentrations are higher during dry than during wet periods.

Within its catchment ground water is transported through an integrated network of fractures and conduits to Crystal spring. Because of the variable size of the fractures and conduits, and the degree to which these features are connected to each other, rates of ground-water movement are highly variable. In the diffuse-flow part of the aquifer where the limestones are dominated by smaller-scale solution features, rates of ground water movement will be slower than in areas where larger scale conduits dominate (the conduit-flow part of the aquifer). Rapid entry and movement of ground water toward the spring is facilitated by sinkhole-like openings in the streambed of a reach of Martin Creek, approximately 2.5 mi to 3 mi (4 km to 4.8 km) north of the spring. Dye-trace experiments conducted during a relatively dry period indicated travel times through the ground water system of approximately 60 hours from sinkholes to the spring, indicating dye travel velocities of approximately 1 mi per day (1,500 m per day).

Analysis of the water-level data collected from wells shows that ground water moves in a southerly direction from the upland area toward Crystal spring and the Cottonwood River. As used here in this report, a catchment is defined as the area that provides recharge or is the source of water for a spring. In this case, the main source of water is infiltrated precipitation that falls on the upland surface north of the spring in the Martin Creek drainage. A well was installed 700 ft (213 m) east of Martin Creek to continuously monitor water levels in the Barneston Limestone. Water levels in the Barneston Limestone fluctuate over several 10s of feet (up to 10 m) depending on the occurrence and duration of wet and dry periods. Martin Creek was dry for much of the project period. Periods of high water levels in the aquifer coincide with streamflow events. Water levels are generally low during extended dry periods, such as occur during the summer, fall, and winter seasons and are high during the spring.

Spring discharge water quality reflects geochemical interactions between the ground water and a limestone aquifer. Calcium and bicarbonate dominate the dissolved ionic constituents. Dissolved solids concentrations ranged from 311 mg/L to 410 mg/L in samples of Crystal spring collected during the Winter and Spring 2002 water sampling events. To monitor changes in water quality, samples of water were collected from Crystal spring monthly from April 2002 to

June 2003 and analyzed for sulfate, chloride, and nitrate concentrations. Sulfate, chloride, and nitrate ranged from 12 mg/L to 42.4 mg/L, 4.5 mg/L to 8.5 mg/L, and 2 mg/L to 9 mg/L, respectively.

The relationships between precipitation events and spring discharge, turbidity, and chemistry, precipitation and water levels in the aquifer near Martin Creek and the short travel time between the sinkholes and the spring indicate that the creek is a major contributor of water to Crystal spring during wet periods. The other major contributor is the drainage from the part of the aquifer where the solution openings are small scale and less well connected. This source sustains spring discharge during drier periods.

Source Water Assessment Area

Two SWAAs were delineated based on the results of this study: one associated with the reach of Martin Creek that traverses the Barneston Limestone outcrop belt, and the other associated with the Barneston aquifer outcrop where the spring is located. The factors considered in defining the SWAAs for Crystal spring include the following: soils, adsorption potential of the aquifer materials, direct vertical pathways to the ground-water system, overlying confining layers, and time-of-travel from sources of recharge to the spring.

SWAA 1 includes the Barneston outcrop belt near Martin Creek bounded on the southeast by the estimated limit of saturation in the aquifer. Most of the attention of this study has focused on the Martin Creek SWAA (SWAA 1). The results of the field investigation indicate that this area is an important source of recharge to Crystal spring.

The contribution of recharge to the Barneston aquifer outcrop area (SWAA 2) in the unnamed drainage and its consequent discharge at Crystal spring is unclear. Conservatively, it would seem reasonable to conclude that this part of the outcrop belt is a source area because of its proximity to the spring and the generally southward direction of ground-water flow in the Barneston. However, it is possible that the local ground-water flow system in this part of the drainage is entirely separate from the flow system that involves the Crystal spring and SWAA 1. Additional research needs to be carried out in this area to determine the level of significance of recharge from SWAA 2 to Crystal spring.

Part 1: Introduction

1.0 Project Description

Section 1453 of the Safe Drinking Water Act (SDWA 1996), as amended by Congress in 1996, established the Source Water Protection Program (SWPP). This program requires all states to submit individual public water supply Source Water Assessment Plans (SWAPs) to the US Environmental Protection Agency for approval and subsequent implementation by the state. SDWA 1996 created this program to provide the tools and opportunities to build a prevention barrier to the contamination of public water supplies. Each SWAP contains a delineation of the source water assessment area (SWAA) for water supply, an inventory of contaminants within the SWAA, an assessment of water system susceptibility to contamination, and a plan for informing the public of the results of SWAP activities.

The SWPP of the Kansas Department of Health and Environment (KDHE) closely follows the program structure laid out in the SDWA 1996 (KDHE, 1999). The state program is based on a three-way partnership among KDHE, the water supply, and technical assistance providers. Each water utility in the state has the legal responsibility to produce and deliver water to consumers that complies with the standards and requirements of SDWA. At the state level, KDHE has the responsibility for assuring that water suppliers comply with these requirements. Technical assistance providers have an interest in protecting water supplies from contamination and the expertise to assist water suppliers in this endeavor.

KDHE (1999) recognized that most public water supplies would require some technical assistance in order to complete their assessments and envisioned that this assistance would come from consultants, the groundwater management districts, the Kansas Geological Survey (KGS), the local environmental protection programs, or professional organizations, such as the Kansas Rural Water Association. KDHE administers many federally mandated programs for environmental protection in addition to the drinking water regulatory program. As a natural outgrowth of these activities the agency declared its intention to share in the responsibility for the completion of the SWAPs with individual water supplies.

1.1 Kansas Geological Survey Involvement in the Technical Assistance Provided to the City of Florence

Near the end of 1999, a staff member of the Kansas Rural Water Association consulted with the KGS on the delineation of the SWAA for Crystal Spring, the sole source for the Florence public water supply in Marion County (Figure 1). Crystal spring is fed by ground water that discharges from a limestone aquifer, and prior to this study there was little information available on which to define its catchment area. To fill this gap, the KGS proposed conducting a two-year study to delineate the spring catchment (source water assessment) area to the Kansas Source Water Protection Program of KDHE. After meeting with the city and the Kansas Rural Water Association this proposal was accepted by KDHE in early 2000.

1.2 SWAA Delineation in a Karst Terrain

Karst aquifers are highly vulnerable to contamination. Ginsburg and Palmer (2002) state that this vulnerability results from (1) the presence of point-source recharge features, such as sinkholes; (2) solutionally widened flow paths; and (3) rapid ground-water velocities and transport of contaminants. Point source recharge features and solutionally widened pathways through the limestone allow rapid movement of contaminants from the surface into the aquifer, especially where the overlying soils are thin or absent, and limit any natural attentuation or filtration of contaminants. Rapid ground-water velocity minimizes the effectiveness of soil-rock-water chemical and biological interactions for in-situ water quality improvement.

Ground-water flow systems are difficult to delineate in karst terrain (Milanovic, 1981). Karst aquifers are extremely heterogeneous and anisotropic because they consist of integrated networks of fractures variously affected by solution widening, solution channels, microcaves, and master conduits. Changes in ground-water flow patterns also arise because of changes in river stage or precipitation events. Ground-water flow paths, divides, and basin boundaries can shift in response to rising ground-water levels during and after major precipitation events. Similarly, backflow from surface water bodies into a karst aquifer can result from rising stream levels, and changes in the ground-water flow can result from point sources of recharge in normally dry streambeds. Conduits in the unsaturated zone can behave independently, causing ground-water basins to interfinger and overlap at any given time (Smart, 1988; Ginsburg and Palmer, 2002). Several detailed studies describing these characteristics of karst aquifer systems can be found in Ford and Cullingford (1976) and White and White (1989).

In most hydrogeologic settings, SWAAs are delineated using an arbitrary or a fixed radius approach. This approach meets the minimal requirements of the SWAP to delineation, but is designed assuming the aquifer consists of a homogeneous and isotropic porous medium. In karst settings this method of delineation can significantly underestimate capture zones for pumping wells and spring catchment areas because of their hydrogeologic complexity. In recognition of this flaw, Ginsburg and Palmer (2002) published delineation guidelines that can be used for springs and wells in karst terrain.

1.3 Goals of the Project

Very little is known concerning the hydrogeology of the karst aquifers of the Flint Hills in Marion County. Section 4.0 is a literature review of the hydrogeology and stratigraphy of the central part of the Flint Hills region in Kansas. From the outset of the study, it was apparent that more information would be required than was available to delineate the SWAA for Crystal spring. Thus, the two goals for this project were to

- Complete a hydrogeologic study of the Crystal spring catchment and its surrounding area, and
- Delineate the SWAA for Crystal spring based on the results of the hydrogeologic investigation.

1.4 Project Objectives

The first objective of the project was to acquire all of readily available hydrogeologic data for the study area. This included:

- Water quality data on spring discharge
- Flow and turbidity data for Crystal Spring
- Locations of area springs from the literature
- The Marion County Soil Survey
- The Lincolnville SW and Cedar Point 7.5-minute topographic maps
- Aerial photos of the area
- Precipitation data from weather stations in the vicinity of Crystal Spring
- WWC-5 records of wells in the southern half of T. 19 S, R. 5-4 E., all of T. 20 S., R. 5-4 E. and Sec. 1-6, T. 21 S, R. 5-4 E.
- Surface geology maps at the county and smaller scale

The second project objective was to develop new information about the study area through field investigation. This included the following activities:

- Field mapping of the bedrock geology using a preliminary county geologic map from Kansas Department of Transportation and aerial photography in the Lincolnville SW and eastern Cedar Point quadrangle maps,
- Conducting an inventory of springs in the Lincolnville SW and eastern Cedar Point quadrangle maps,
- Mapping of the distribution of sinkhole features in Martin and Bruno Creek drainages in the Lincolnville SW and eastern Cedar Point quadrangle maps,
- Dry winter season and a wet, spring season data collection activities to gather ground-water level data and obtain ground and surface water samples for chemical analysis,
- Water sample collection and measurement of discharge from Crystal spring on a monthly basis,
- Installation of a monitoring well near Martin Creek,
- Continuous monitoring of water levels in the monitoring well, and
- Dye trace studies involving the Crystal spring and the sinkholes in Martin Creek.

The third project objective was to use all of the hydrogeologic information to delineate the SWAA for Crystal spring.

2.0 This report

2.1 Purpose

The purposes of this report are to summarize the hydrogeology of Crystal spring catchment and adjacent areas and to present the Crystal spring SWAA based on all of the information collected during this study.

2.2 Report Organization

The report is organized into three parts. Part 1 provides background information on the overall project. Part 2 contains the results from the hydrogeologic investigation, including presentation of a conceptual hydrogeologic model of Barneston aquifer and the Crystal spring catchment. Part 3 presents the delineation of the SWAA for Crystal spring.

Part 2: Hydrogeology and Hydrogeochemistry of the Study Area

3.0 Setting

3.1 Study Area Location and Extent

To begin the project, an estimate of the Crystal spring catchment extent had to be made in order to establish a study area. Crystal spring is located at the foot of the north valley wall of the Cottonwood River, approximately one mile north of the town of Florence in southeastern Marion County (Figure 1). The Cottonwood River was assumed to be a regional discharge area for the alluvial and the bedrock aquifers in the drainage, and the regional ground-water flow direction was assumed to be southerly from the uplands toward the river and the spring.

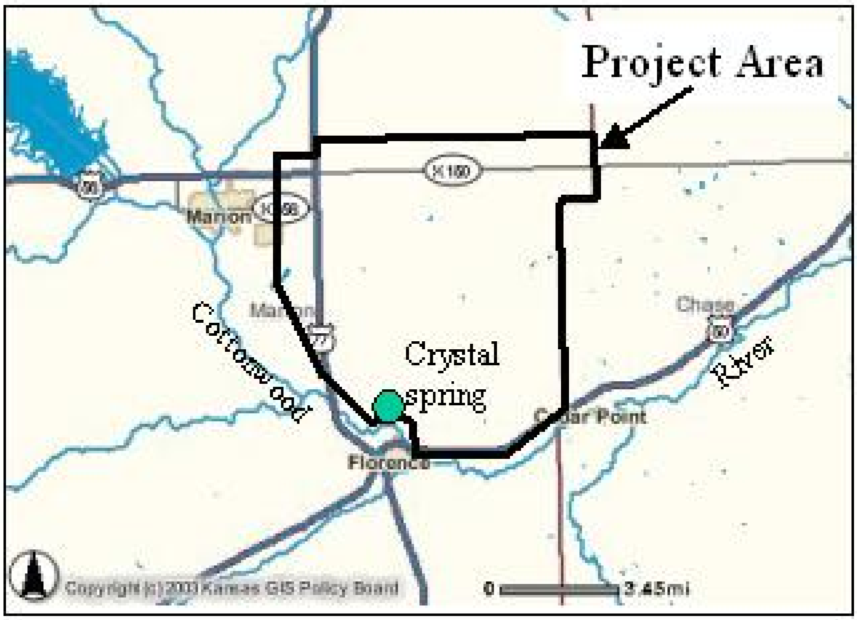
The estimated size of the Crystal spring catchment was derived from estimates of mean annual potential recharge from Hansen (1991) and two Crystal spring discharge estimates from Sawin et al. (1999). Mean annual potential recharge was converted to a mean annual volume per unit area of land surface and the estimates of discharge rate were converted to the annual volume of water discharged from the spring. These estimates yielded a catchment size of approximately 34.5 mi^2 with Crystal spring situated along the southern catchment boundary.

To insure that all of the Crystal spring catchment was included in the study area, the study area was expanded northward up to Kansas State Highway 150, and eastward to the Marion-Chase County boundary to include much of the Bruno Creek drainage. The north boundary of the study area nearly coincides with the drainage divide separating Martin and Bruno creeks from Middle Creek. The study area is approximately 57 mi^2 in size and covers most of the Lincolnville SW and the western part of the Cedar Point 7.5-min topographic quadrangle maps.

3.2 Physiography

The study area is located in the central part of the Flint Hills physiographic province (Figure 1; Schoewe, 1949). The contrasting resistance to erosion among the several bedrock units and variations in soils and vegetation produce considerable diversity in the region. Three characteristic types of terrain exist in the study area. Low-relief, dissected bluestem upland prairie is underlain by thick massive limestones. Steep, wooded or grassy, stair-step valley sides underlain by thick to thin shales and limestones occupy lower landscape positions adjacent to downcutting intermittent minor stream courses. In the southern part of the study area, the Cottonwood River and Martin and Bruno creek flow through valley bottoms underlain by alluvial sediments or alternating thick to thin limestones and shales and covered by food-plain forest or savanna vegetation. The highest elevations are in the northern part of the study area where the maximum elevation is 1,450 ft to 1,470 ft above mean sea level (amsl). The lowest elevation is in the Cottonwood River valley in the southeastern corner of the study area and is 1,250 ft amsl.

The Cottonwood River is a meandering stream that drains all of the study area. Martin and Bruno Creeks are tributary to the Cottonwood and flow from northwest to southeast across the



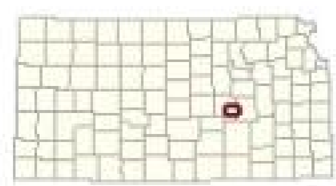


Figure 1. Extent of the project area in Marion County, Kansas.

study area. For much of the year, the lower reach of Martin Creek is dry and flow in the upper reach is fed by ground-water discharge. Bruno Creek is not an intermittent stream and is fed by ground-water discharge.

3.3 Geologic Structure

The study area is located on the west flank of the Nemaha uplift near the intersection of the north-northeast and south-southwest trending Nemaha tectonic zone and the northwest-southeast trending Fall River tectonic zone (Berendsen and Blair, 1986). The structural development of southeastern Marion County is characterized by repeated adjustment of basement blocks along pre-existing zones of weakness throughout post-Precambrian time. Locally important structures that have expression in Paleozoic rocks include: (1) the Cedar Creek Fault that is coincident with the northwest-southeast trending reach of the Cottonwood River, (2) an unnamed doubly plunging north-northeast – south-southeast trending anticline (3) a south-southwest trending anticline in the extreme southeastern part of the study area, and (4) an unnamed structural dome in the northeast part of T. 20 S., R. 5 E. (Figure 2). The regional dip of the bedrock units is in a westerly direction at approximately 3.2 m/km. Both the direction and the amount of dip may vary locally due to the influence of local geologic structures.

3.4 Soils

Soil series in the study area belong to three soil associations (Figure 3; Horsch and McFall, 1983). Soils of the Irwin-Ladysmith association can be found in the highest part of landscape in the northeastern part of the study area and consist of deep, near level to moderately sloping, moderately well drained to somewhat poorly drained soils that have a clayey subsoil. Irwin soils are moderately well drained with an upper silty clay loam overlying thick, silty clay subsoil. Ladysmith soils are somewhat poorly drained and possess an upper silty clay loam texture with a silty clay subsoil. The minor soils in this association are Clime, Goessel, and Rosehill soils. The moderately deep, calcareous Clime soils are on lower side slopes and on the sides of upland drainageways. The clayey Goessel soils are in slight depressions in the topography and moderately deep Rosehill soils are on side slopes.

Most of the study area is underlain by the soils of the Labette-Tully-Sogn association. This association consists of deep to shallow, nearly level to strongly sloping, well drained to somewhat excessively drained soils that have clayey or silty subsoil. This association occupies an intermediate upland position in the landscape. Small, narrow, intermittent stream valleys dissect this part of the landcape and bedrock exposures are common near edges of sideslope breaks where slopes are steep. Moderately deep, well-drained Labette soils are formed in the residuum of limestone on ridge tops and upper side slopes. The upper part of the profile consists of silty clay loam and lower subsoil consists of silty clay containing limestone gravel. Tully soils are deep and well drained and formed in colluvium near the footslopes. The upper part of the Tully profile consists of silty clay loam and the lower subsoil possesses a silty clay texture. Sogn soils are shallow, somewhat excessively drained and formed in residuum from limestone on ridgetops and upper side slopes. Typically thin, the Sogn soil profile consists of silty clay

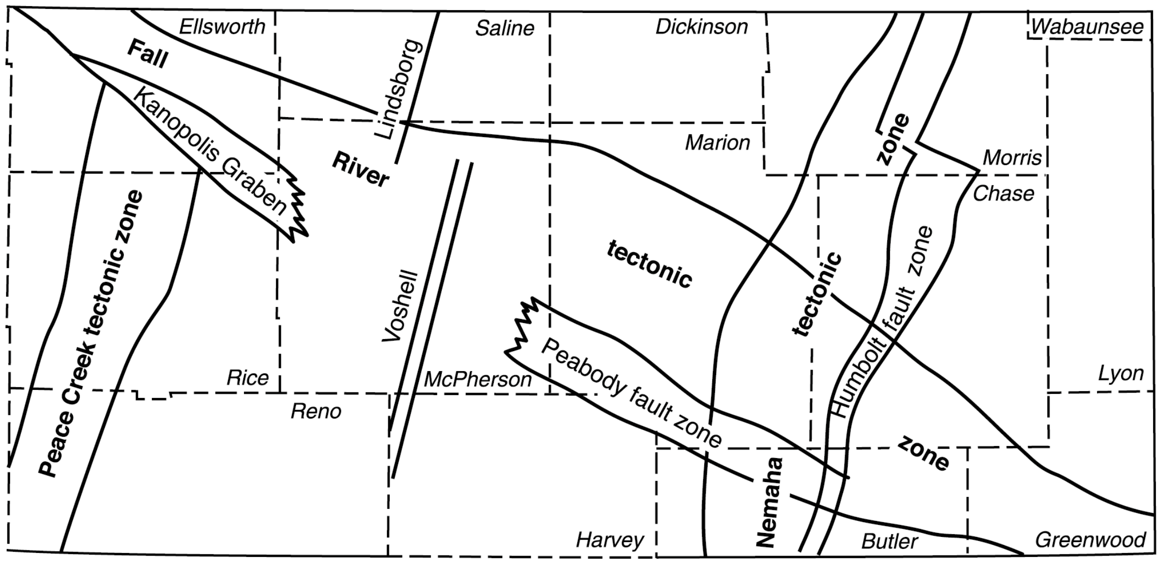
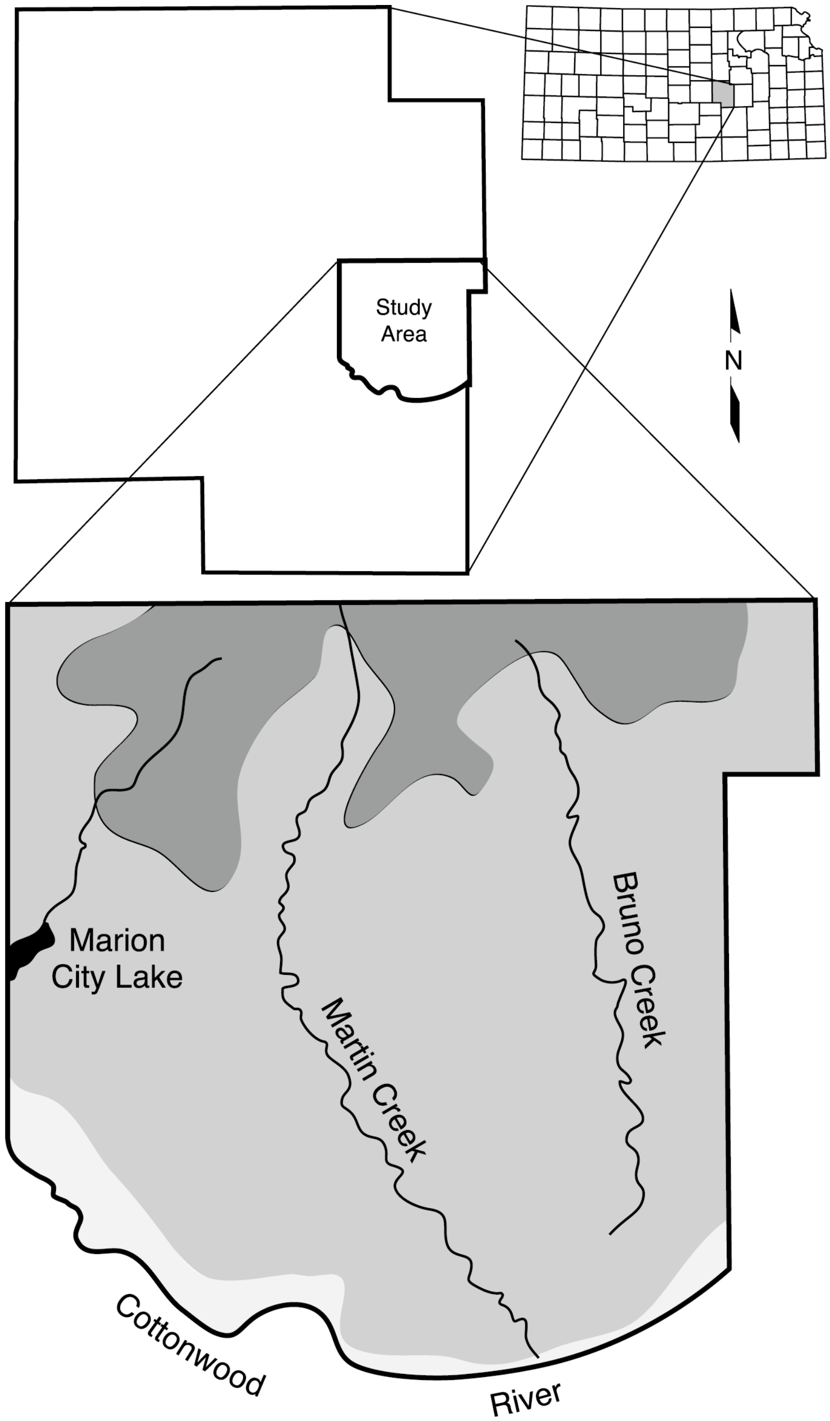


Figure 2. Major tectonic elements of study area. (From Berendson and Blair, 1986)



Irwin-Ladysnith Association

Labette-Tully-Sogn Association

Verdigris-Chase-Reading Association

Figure 3. Soil associations within the project area.

loam over limestone. The minor soils in this association are Clime, Dwight, Florence, Irwin, and Verdigris soils. The deep moderately well drained Dwight and Irwin soils are on ridgetops. The Florence soils have a cherty subsoil and are on ridgetops and side slopes. The deep, moderately well drained Verdigris soils are on narrow flood plains, such as in Martin and Bruno creeks.

The Verdigris-Chase-Reading soil association underlies the Cottonwood River valley. This association consists of deep, nearly level, well drained to somewhat poorly drained soils that have a silty or clayey subsoil. The moderately well drained Verdigris soils formed in silty alluvium on the flood plain and possess a silt loam texture. The somewhat poorly drained Chase soils formed in clayey alluvium on low stream terraces and vary in texture from a silty clay loam to silty clay in the lowermost part of the profile. Reading soils are well drained, formed in silty alluvium on stream terraces, and are a silty clay loam throughout its profile.

3.5 Climate

The study area has a continental climate characterized by large daily and annual variations in temperature. The average annual, mean temperature in the study area is approximately 12.8°C for the period 1961-1990 (Goodin et al., 1995). Winter conditions typically last from December through February, and the severity of these conditions depends on the frequency of southward moving polar air masses in the region. Mean winter temperature is –0.5°C for the 1961-1990 period (Goodin et al., 1995). Warm summer temperatures typically span the period from April to October. Both the spring and fall seasons are relatively short. Mean summer temperature is 25°C for the 1961-1990 period (Goodin et al., 1995).

Mean annual precipitation at Marion is 813 mm to 838 mm with greater than 70% of the precipitation occurring between April and September during the growing season (Horsch and McFall, 1983). During the study conditions were drier than normal. In 2000 the annual precipitation at the Florence and Marion Lake weather stations was 703 mm and 637 mm, respectively. The wettest months were March, June, July and October and the driest months were January, August, September, November, and December of 2000 (Figure 4a). In 2002 the annual precipitation recorded at the Florence and Marion stream gages on the Cottonwood River was 599 mm and 669 mm, respectively, and for the January-May 2003 period, 335 mm and 289 mm, respectively. The wettest months of the year were April through June 2002, October 2002, and March through May 2003 and the driest months were January through March 2002, July 2002, and November 2002 through February 2003 (Figure 4b).

3.6 Land Use

More than 97% of the study area is in pasture with less than 2% in cultivated croplands (Figure 5). Croplands under cultivation can be found in the Cottonwood River valley and in isolated areas of Martin and Bruno creek valleys. Urban uses take up approximately 1% of the land area in the vicinity of Marion City Lake.

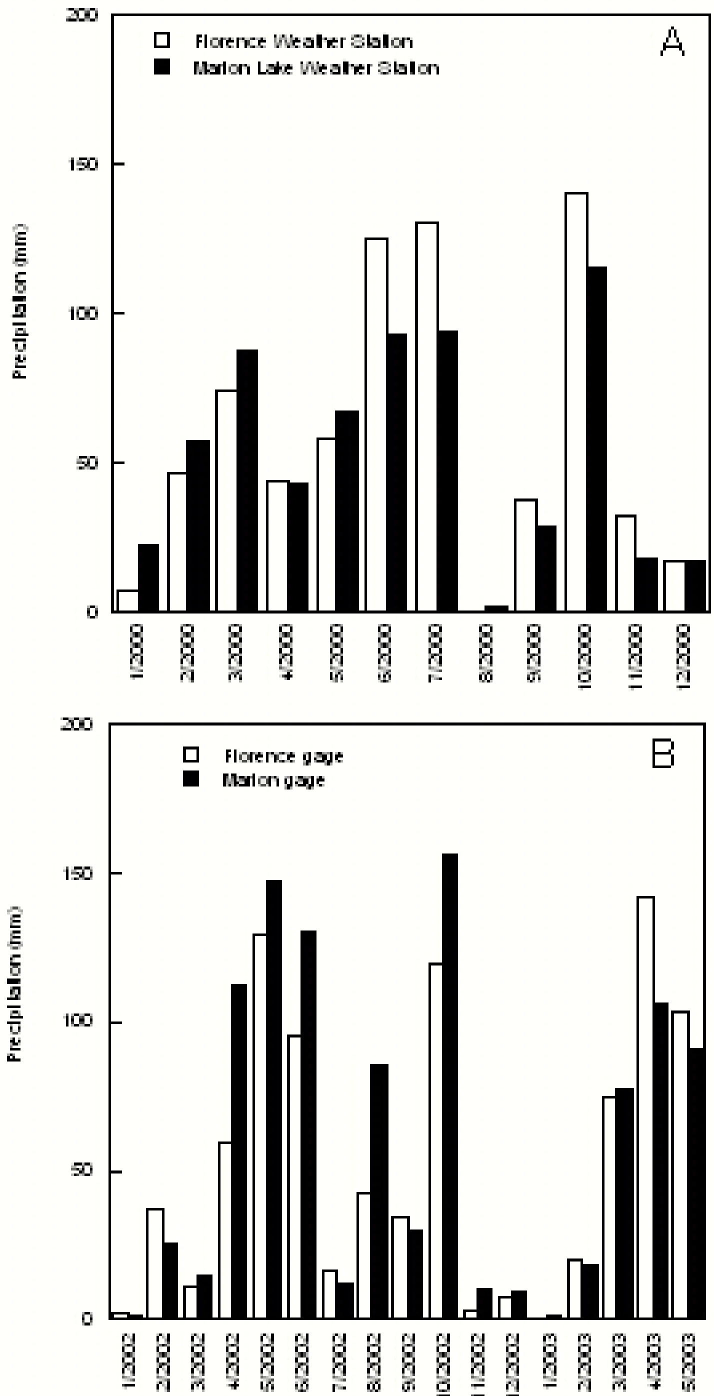


Figure 4. Monthly precipitation for 2000 using data from the Marion Lake and Florence weather stations (A) and for 2002-3 using data from the Marion and Florence stream gage sites on the Cottonwood River (B).

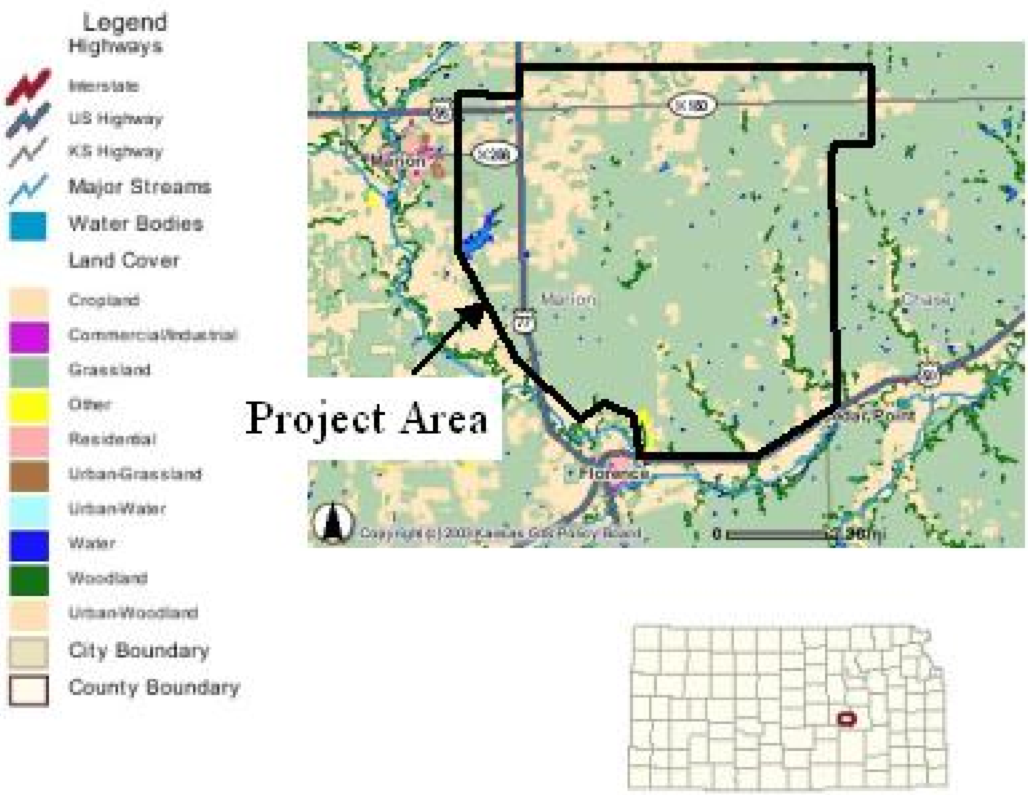


Figure 5. Land use in the project area.

4.0 Previous work

4.1 Hydrogeology

Neither the KGS nor the US Geological Survey conducted extensive investigations into the hydrogeology of the central Flint Hills region in the 20th century. In an early KGS volume, Bailey (1902) stated that the source of the discharge from Chingwassa springs in Marion County was from Permian rocks. Fath (1922) noted the formation of sinkholes at the surface in the middle part of the Ft. Riley Limestone Member of the Barneston Limestone in Butler County and attributed their formation to limestone dissolution by infiltration of precipitation from the land surface. Gordon (1938) reported on the recent formation of a sinkhole in the top of the Ft. Riley near Potwin, also in Butler County. He attributed the cause of the depression to the solubility of the limestones and the eventual caving in of the roof of a cavern in the upper part of the unit.

In the 1980s, a multi-year study focusing on the ground-water resources of Marion County was conducted by the Kansas Geological Survey. The results of this study are reported in O'Connor (1982, 1984), O'Connor and Chaffee (1983, 1984), Chaffee (1984, 1988), and Chaffee and O'Connor (1986). The objectives of this study were to (1) evaluate the ground- and surface-water systems of the county, (2) evaluate the extent of ground-water contamination by oil production activities, and (3) determine the causes of local depletion of fresh ground-water supplies. Approximately 600 wells and springs were inventoried and sampled for water quality and approximately 200 stream sites were evaluated for baseflow and 150 water samples were collected for determination of chloride and specific conductance in late fall and winter during the study. Estimates of stream and spring flows were made at the same time at the stream sites. In addition, test drilling was done and observation wells were installed to better understand the hydrogeology.

The results of this study revealed that the hydrogeology of eastern Marion County is dominated by karst aquifer systems. Many of the springs in the county discharge at relatively high rates [>1 ft^3/sec; 4th order or higher according to the Meinzer (1923) classification of springs based on discharge] from aquifers in the Winfield and Barneston Limestones (Figure 6). The majority of the streams in the county flow throughout the year because of the relatively steady discharge of ground water from shallow aquifers. The study also documented the vulnerability of the ground-water systems to contamination both as a result of poor well-construction and -abandonment practices, improper plugging of seismic shot holes, and improper disposal of oil brines and the depletion of ground-water sources from abandoned flowing wells.

In late 1980s and early 1990s, a study was conducted to investigate the hydrogeology of thin, sandwiched, limestone aquifers at the Long Term Ecological Research (LTER) site on the Konza Prairie Research Natural Area in Geary County, Kansas, and reported in Macpherson (1996). The part of the LTER selected for this detailed study is in the lower part of a 1.2-km^2 (0.46-mi^2) watershed in the Kings Creek drainage. The study site is underlain by alternating shales and thin limestones of the Council Grove Group with a thin cover of Quaternary alluvium adjacent to the drainage in the valley bottom. Well nests were installed along four traverses across the main

Figure 6. Fourth order or higher magnitude Springs discharging from the Winfield or the Barneston Limestone in Marion County according to the Meinzer (1923) classification of springs. From O'Connor and Chaffee (1983).

stem of the investigated drainage. Using water-level and water-chemistry data from the monitoring wells and the stream, Macpherson found confined and sometimes unconfined conditions in the limestone aquifers and documented the movement of water between the stream, the alluvial aquifer and at least one of the limestone aquifers in response to recharge events.

In the late 1990s a study was conducted by the KGS to inventory significant and representative springs of the Flint Hills region, including several from Marion County, and reported in Sawin et al., (1999). Spring locations were determined using a GPS (global positioning system) device. Spring discharge rates were estimated and measurements of discharge temperatures were also made. Water samples were collected and analyzed by the Analytical Services Section of the KGS for specific conductance and chloride, sulfate, and nitrate concentrations. Sawin et al. found that water from the springs of the Flint Hills region is of excellent chemical quality. They concluded that unlike the springs in some other areas of Kansas, spring discharge had not declined and spring chemical quality had not degraded over time as a result of human activities in the region.

4.2 Selected Published and Unpublished Reports Relating to Permian Stratigraphy and Surface Geology

Descriptions of the stratigraphy and lithology of the bedrock units that crop out in the study area can be found in Zeller (1968). Twiss (1991) described the Chase Group interval of a core collected from a well downdip of the outcrop belt in Riley County, Kansas, and noted the occurrence of bedded and disseminated evaporites in the Barneston Limestone. Later outcrop stratigraphic studies of the Chase Group in Kansas, Nebraska, and Oklahoma were conducted and reported in Mazzulo et al. (1997). Of interest to this study is the recognition of zones of enhanced secondary porosity resulting from evaporite dissolution in the Barneston Limestone.

County-scale mapping of the surface geology of Marion County was conducted by the US Geological Survey for the purpose of producing an inventory of materials suitable for highway construction (Byrne et al., 1959). Later, a partially completed surface geology map by Lynn Meyers, Kansas Department of Transportation, was made available to Howard O’Connor, KGS, to assist in the KGS study of Marion County.

5.0 Hydrostratigraphic units

5.1 Stratigraphy

5.1.1 Bedrock Units

Lower Permian bedrock units belonging to the Chase Group underlie the study area and consist of alternating limestones and shales (Figure 7). The stratigraphic nomenclature used in this report follows the 1994 Classification of Rocks in Kansas (Baars, 1994). The following geologic unit descriptions are taken from Zeller (1968) and Mazzulo et al. (1997). Where appropriate the lithology of the carbonate rocks is described using Dunham's (1962) classification.

Plate 1 is a map of the surface geology in the study area. Shown on the map are the traces of the tops of: (1) the Matfield Shale, (2) the Florence Limestone Member, Barneston Limestone, (3) the Ft. Riley Limestone Member of the Barneston Limestone, (4) the Holmesville Shale Member of the Doyle Shale, (5) the Gage Shale Member of the Doyle Shale, (6) the Grant Shale Member, Cresswell Limestone, (7) the Paddock Shale Member of the Nolans Limestone. The contact separating the alluvium from the bedrock units was also mapped. Plate 1 was derived using an unpublished, partially complete 1984 county geologic map of Lynn Meyers, Kansas Department of Transportation. The map was checked using air photographs and visual inspection in the field. The traces of contacts were then transferred to the Lincolnville SW and Cedar Point, 1:24,000 scale topographic maps that cover the project area extent.

5.1.1.1 Matfield Shale

The Matfield Shale consists of three members, which are, in ascending order, the Wymore Shale, the Kinney Limestone, and the Blue Springs Shale. The Matfield crops out at the surface in the lower reach of Bruno Creek and in the Cottonwood River, below the Barneston Limestone in the lower valley wall (Plate 1). The total thickness of the Matfield shale in the study area is approximately 15 m and is conformable beneath the Barneston Limestone. The Wymore Shale is poorly exposed in the study area and is a gray to yellowish-gray shale with beds of vivid red, green, and purple shale. The Kinney Limestone consists of interbedded shale, shaly limestone, and limestone capped by 0.75 m of limestone. The Kinney is conformable on the underlying Wymore, but is separated from the overlying Blue Springs Shale by an unconformity. Gray-green, yellow and brownish-green shale and mudrock dominate the Blue Springs. The uppermost part of the Blue Springs is marked by an interval of interbedded limestone and shale approximately 1 m in thickness.

5.1.1.2 Barneston Limestone

In the study area, the Barneston Limestone consists of three members, which are in ascending order: the Florence Limestone, the Oketo Shale, and the Ft. Riley Limestone. The type area for the Florence Limestone is the Sunflower rock quarry in the southern part of the study area (Secs. 5 and 6, T. 21 S., R. 5 E.). The top of the Barneston is conformable with the overlying Doyle

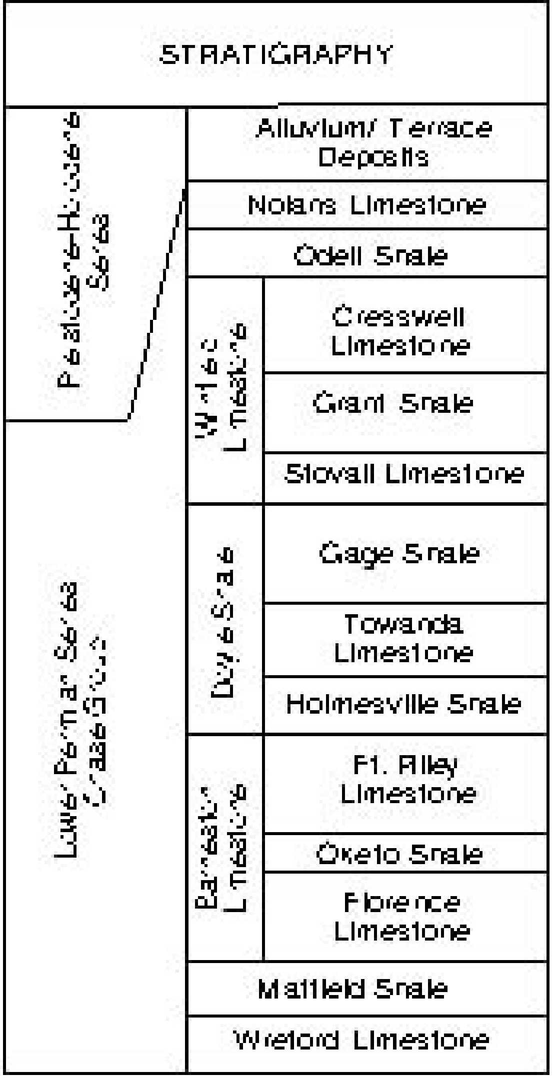


Figure 7. The stratigraphy of the shallow subsurface in the study area (From Zeller, 1968)

Shale. Mazzulo et al. (1997) proposed elevating the rank of the Florence to that of a formation. In addition they proposed revising its definition to include the Oketo Member at the top and non-cherty limestone at the bottom, the Cole Creek Member that belongs in the Blue Shale Member of the Matfield Shale under the currently accepted classification. The rocks between the Cole Creek and the Oketo members of the proposed Florence Formation were not grouped into a named member. Mazzulo et al. proposed retention of the type section at the quarry with a supplemental section at nearby Cedar Point in southeastern Chase County. In making this change to the stratigraphic classification, Mazzulo et al. also proposed elevating the Ft. Riley from the rank of member to that of a formation. None of these changes in nomenclature have been adopted by the KGS (Merriam, in review). Using the revised definition of the Florence, the total thickness of the Barneston is 25.7 m thick in the study area.

The Florence member consists of mostly light-yellow limestone, including fine-grained biowackestones and some biopackstones. About 1.5 m below the top of the Florence is a 0.3 m thick zone of interbedded fossiliferous mudrock and shaly lime mudstone to biowackestone. Chert occurs as nodules and layers. The top 0.5 m of the Florence directly beneath the Oketo is a non-cherty, light yellow, oncolitic biowackestone to packstone. The contact with the overlying Oketo is conformable and mostly sharp to gradational over distances of less than 10 cm.

In the study area, the Oketo Shale consists of highly fossiliferous, dark-gray to dark yellow-gray silty, bioturbated mudrock and is about 0.8 m thick in the Sunflower rock quarry. The Oketo-Ft. Riley contact is abrupt and both units are conformable across the contact.

Within the study area, the Ft. Riley Limestone member can be subdivided into a lower massive bedded, dominantly oncolitic limestone and an upper section of shaly limestone and mudrocks. The lower subdivision is expressed topographically as a prominent ledge-former (the rim-rock) in the Flint Hills and is approximately 3 m thick in the Sunflower rock quarry in the southern part of the study area. The upper subdivision is a recessive-weathering unit and is 11.4 m thick in the quarry. A 0.3-m thick lens of cross-stratified carbonate sands occurs in the lower part of the upper subdivision, north of the quarry in some of the road cuts.

5.1.1.3 Doyle Shale

The Doyle Shale consists of three members, which are in ascending order: the Holmesville Shale, the Towanda Limestone, and the Gage Shale. Total thickness of the Doyle is approximately 16.8 m. The Doyle/Winfield contact is abrupt and conformable.

In the Sunflower rock quarry the Holmesville shale consists of 3.8 m of dominantly light to dark green unfossiliferous shale containing a single bed of shaly lime mudstone. Calcite-lined geodes can be found in the shales. The Holmesville/Towanda contact is abrupt and both units are conformable across the contact.

The Towanda Limestone Member consists of 3 m of thin-bedded fossiliferous limestone, biowackestone-packstone to biograinstone, and thin interbeds of mudstone and shaly limestone. The upper contact with the overlying Gage Shale Member is abrupt and conformable.

The Gage Shale Member is poorly exposed in the study area and consists of 9.4 m of red to green to brown and yellow-brown largely unfossiliferous mudrocks with sporadic thin shaly limestone, biowackestone, in the upper third of the member.

5.1.1.4 Winfield Limestone

The Winfield Limestone consists of three members, which are, in ascending order: the Stovall Limestone, the Grant Shale, and the Cresswell Limestone. Total thickness of the Winfield in the study area ranges from 12.2 m to 13.1 m.

The Stovall Limestone consists of 0.7 m to 1.0 m of cherty, fossiliferous limestone, biowackestone that is easily traced in the field below the Cresswell Limestone.

The Grant Shale consists of 1.4 m of fossiliferous mudrock with lenses of shaly limestone. In the vicinity of Marion County Lake, the upper part of the unit contains abundant golfball-size chert nodules. Calcite-lined geodes are common in the lower part.

The Cresswell Limestone is a cherty, fossilferous, thin to medium-bedded limestone and shaly limestone, biowackestone to biopackstone-grainstone. Large irregularly-shaped chert nodules commonly litter outcrops of the Cresswell. The uppermost Cresswell may consist of medium bedded, porous limestone and cross-stratified biograinstone, partly oolitic and coarse-grained with some lime mud intraclasts. Anhydrite nodules and evaporite nodule molds are common throughout this unit. Total thickness of the Cresswell varies slightly across the study area, but is approximately 8.8 m.

5.1.1.5 Odell Shale

The Odell Shale is poorly exposed at a few locations in the study area and consists of approximately 3 m of unfossiliferous red and green mudrocks.

5.1.1.6 Nolans Limestone

The Nolans Limestone is poorly exposed near the northern boundary of the study area and consists of three members, which are, in ascending order: the Krider Limestone, the Paddock Shale, and the Herrington Limestone. Total thickness of the Nolans is approximately 5.8 m.

5.1.2 Unconsolidated Deposits

5.1.2.1 Quaternary Alluvium and Terrace Deposits

Quaternary alluvium and terrace deposits underlie the Cottonwood River valley and the lower reaches of the Martin and Bruno Creek. The maximum thickness of these deposits in the study area is estimated to be less than 10 m in the Cottonwood valley and less than 5 m in Martin and Bruno creek valleys. The unconsolidated sediments consist of silt and clay with lenses of sand and gravel. In Chase County, the gravel and sand deposits are most often in the lower part of the deposits and contain limestone fragments, and rounded pebbles of chert and quartz (Moore et al., 1951).

5.2 Surface Geology

The outcrop pattern in the bedrock units in the surface geology map in Plate 1 shows the westward dip of the bedrock units in the study area. Fluctuations in the westward dip are attributable to local geologic structure and thickness changes due to depositional processes or evaporite dissolution. The map also shows the incision of the drainage into the underlying bedrock units of the Chase Group. Most of the streambeds in the middle and upper reaches of the Martin and Bruno creek are formed in the bedrock units.

5.3 Regional Hydrostratigraphic Units

An essential part of any ground-water investigation is to define and map aquifer and confining units (hydrostratigraphic units) in the subsurface of the study area. Hydrostratigraphic units were originally defined by Maxey (1964) as bodies of rock with considerable lateral extent that act as a reasonably distinct hydrologic system. Delineation of these units subdivides the geologic framework on the basis of permeability and thus aids in flow system definition.

The shallow subsurface stratigraphy is best characterized as a sequence of alternating limestones and shales. In this study quantitative data are unavailable to discern permeability differences between the various lithologies that occur in the study area. However, limestones are usually considered to be at least several orders of magnitude more permeable than shaly rocks (see for example Table 3.2, p. 65, in Domenico and Schwartz, 1990). Fetter (1994) suggested that to be an aquifer, the rock unit should have an intrinsic permeability of at least 0.01 darcy or a hydraulic conductivity of about 0.03 ft/day (1×10^{-7} m/s). Most shales have hydraulic conductivities considerably less than 0.03 ft/day. This is largely because of post-diagenetic dissolution of carbonates by percolation fresh ground waters into the limestones after exhumation by geologic processes (see for example White, 1990; and Palmer, 1990). Joints resulting from tectonic stresses or unloading by erosional processes often facilitate development of secondary porosity and permeability within carbonate rocks. In contrast, the effect of fracturing on shales is considerably less due to the low solubility of the siliclastic components in these rocks.

Another line of evidence that can be helpful in discerning the aquifer units are the ground-water sources tapped by wells or springs in the study area. Lohman and others (1972, p. 2) define an aquifer as "a formation, group of formations, or a part of a formation that contains sufficient saturated permeable material to yield significant quantities of water to wells and springs." In eastern Marion County, the sources of ground water to wells are the Barneston Limestone, and Towanda Limestone Member of the Doyle Shale, and the Cresswell Member of the Winfield Limestone (O'Connor and Chaffee, 1983). All of the magnitude 3 or higher springs [Meinzer (1923) scale] in eastern Marion County have as their source ground water from the Barneston or the Winfield Limestone. During the course of this study, it was discovered that many perennial springs and seeps are fed by ground-water discharge from the Towanda Limestone (Figure 8). Most of these springs are magnitude 5 or lower on the Meinzer (1923) scale. None of the springs in eastern Marion County issue from the shale units and none of the wells are constructed in order to tap into ground-water supplies in the shales.

Thus, the hydrostratigraphic section consists of aquifers in the carbonate units separated by low permeability units dominated by shale (Figure 9). The primary shallow aquifer units in the study area are the Barneston Limestone and the Towanda Limestone Member of the Doyle Shale. The Cresswell Limestone is present only in the northern part of the study area. None of the wells in this part of the study area tap into ground-water supplies in the Cresswell. Hence, it is believed to contain only a very thin saturated thickness because of natural drainage through springs and seeps along the eastern edge of the outcrop belt.

5.4 Secondary Porosity and Permeability in the Barneston Limestone

Carbonate karst aquifers contain three types of porosity: primary and secondary fracture and solution conduit porosity (Smith et al., 1976). These porosity types and their relative proportion determine the distribution and ranges of permeability and storativity observed in these rocks and the prevalence of either laminar or turbulent flow for a given set of hydrologic conditions. Differences in flow rates and storativity between the different porosity types generally imply that most of the flow is in the conduits and most of the storage is in the matrix and fractures that have not been affected by solution enlargement (Atkinson, 1977; Martin and Dean, 2001). Schuster and White (1971) recognized a continuum of flow system types in carbonate rocks with two end members: (1) diffuse flow along joints, fractures, partings, bedding planes, and other small interconnected openings measured in centimeters and (2) conduit flow through integrated networks with the turbulent flow of water through passages measured in centimeters to meters.

Primary porosity of a rock mass is defined as the total volume of the intergranular pore space relative to the bulk volume of the rock and the effective porosity is the total volume of the interconnected pore space (Freeze and Cherry, 1979). Total and effective primary carbonate rock porosities are in the 5-15% range and 0.1-5% respectively (Domenico and Schwartz, 1990). The low effective intergranular porosity in carbonate rocks is due to the interlocking crystal structure of the rock mass, which promotes isolation of individual pores from each other (Bathurst, 1971). Primary permeability is the intergranular resistance to the flow of water through the rock mass or the efficiency with which a rock mass transmits ground water by

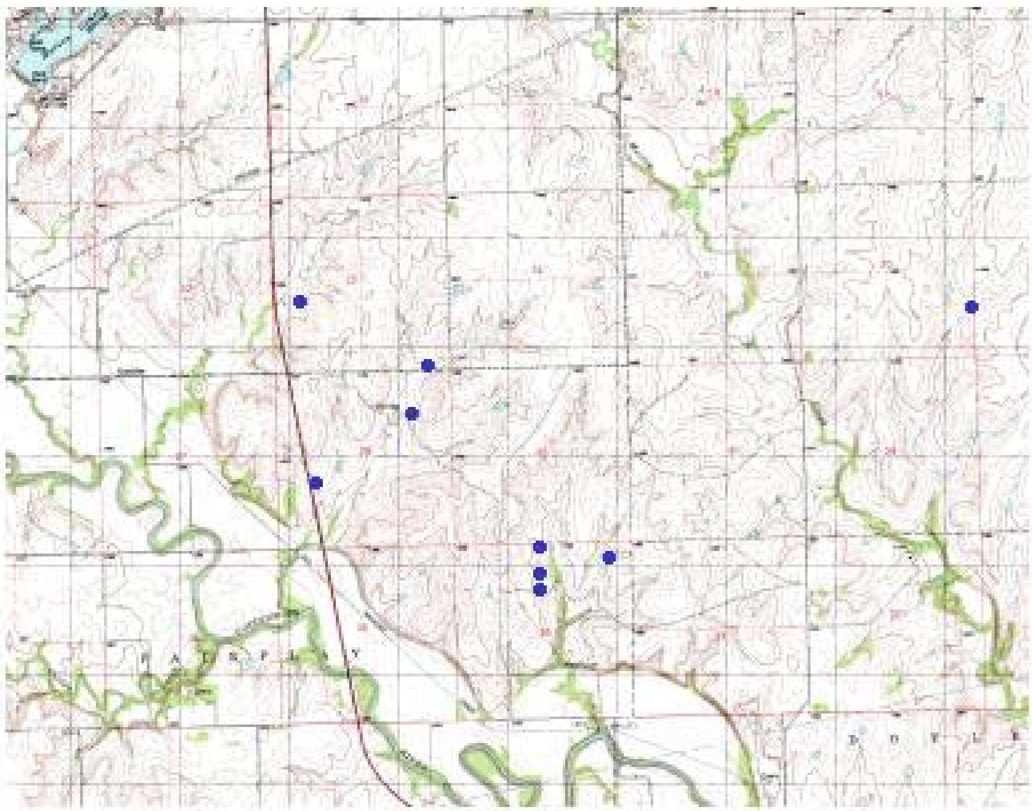


Figure 8. Springs and seeps fed by ground-water discharge from the Towanda Limestone Member of the Doyle Shale in the southern part of the Lincolnville SW 7.5-min topographic quadrangle map.

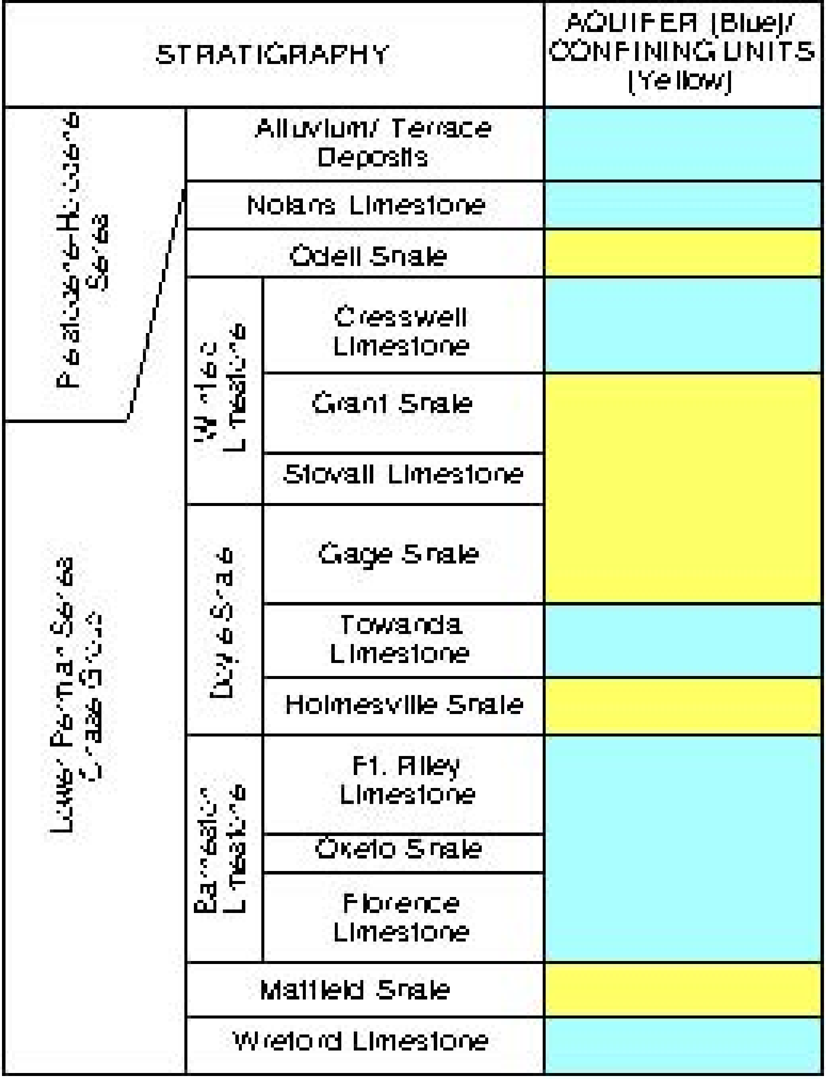


Figure 9. The hydrostratigraphic section of the shallow subsurface in the study area.

means of intergranular flow. Typical values of primary permeability are in the range of 1 x 10^{-12} to 6 x 10^{-9} cm^2 (0.1 to 624 md; Domenico and Schwartz, 1990).

Secondary porosity in carbonate rocks is defined as the relative volume of the rock mass occupied by primarily by fractures (joints or faults), bedding planes, and solution conduits relative to the bulk volume (Freeze and Cherry, 1979). Likewise, secondary permeability is the resistance to the flow of ground water through secondary porosity features in the bulk rock mass (Smith et al., 1976). This additional porosity and permeability result from post-diagenetic, karstic processes acting on the carbonate rock mass in near-surface continental environments to produce an increase in the interconnected pore space and in the efficiency of water flow (White, 1990; Palmer, 1990). In general carbonate dissolution depends on the following environmental factors: (1) pH of the interstitial fluid, (2) dissolved carbon dioxide content, (3) salts in solution, and (4) temperature (Brahanna et al., 1986). Karstified limestone porosity and permeability values are in the range of 5-50% and 1 x 10^{-9} to 0.1 cm^2 (120 to 1.2 x 10^{10} md; Brahana et al., 1986). Figure 10 shows the generalized porosity, pore size, and hydraulic conductivity of several types of carbonates. Bocker (1969) is quoted in Brahanna et al. (1986) as stating that a pore size of 0.01 mm is the lower threshold necessary for significant flow and solvent action by ground water. On the basis of this lower threshold, Brahanna et al. emphasize that carbonate dissolution is widely diffused wherever high primary porosity occurs and thus the pores grow more or less equally. As a result conduit systems cannot be produced and cannot capture additional surface water where rock dissolution is diffused on the scale of centimeters or less (Ford, 1980).

5.4.1 Observation of Joints and Fractures

Joints are caused by tectonic stresses or from erosional unloading and are structures that result from brittle behavior of the rock in which there is no relative motion of the rocks on either side of the structure (Hobbs et al., 1976). Joints usually occur as joint sets or families of fractures with more or less regular spacing in a given rock type and a common origin. Joint sets may be related to regional structures and stresses or may result from erosional unloading. In carbonate rocks, joint apertures may become enlarged through dissolution of soluble rock components can be important avenues of ground-water flow. The entire assemblage of joints present in an exposure or any region of interest is a joint system.

A total of 63 joint orientation (strike) measurements were taken in natural outcrops and quarry exposures of the Barneston Limestone using a Brunton compass as part of the field mapping for this project. Measurements of joint orientation were taken at exposures after careful observation of the fracture surfaces to eliminate human-induced fractures from consideration. Other characteristics of the joints were also noted including, the degree of weathering of the joint surfaces, evidence of carbonate dissolution, fracture fillings, solution channel development associated with the fractures and dip angle in relation to the bedding. In general joint sets were difficult to discern. Many of the fractures observed in natural outcrops were not continuous across vertically adjacent limestone beds. Quite often the joint surfaces were not planar and thus their orientation was variable. Only joints with consistent orientation were measured. The strike data were lumped into 5° categories from 0° to 90° to create polar frequency plots for the

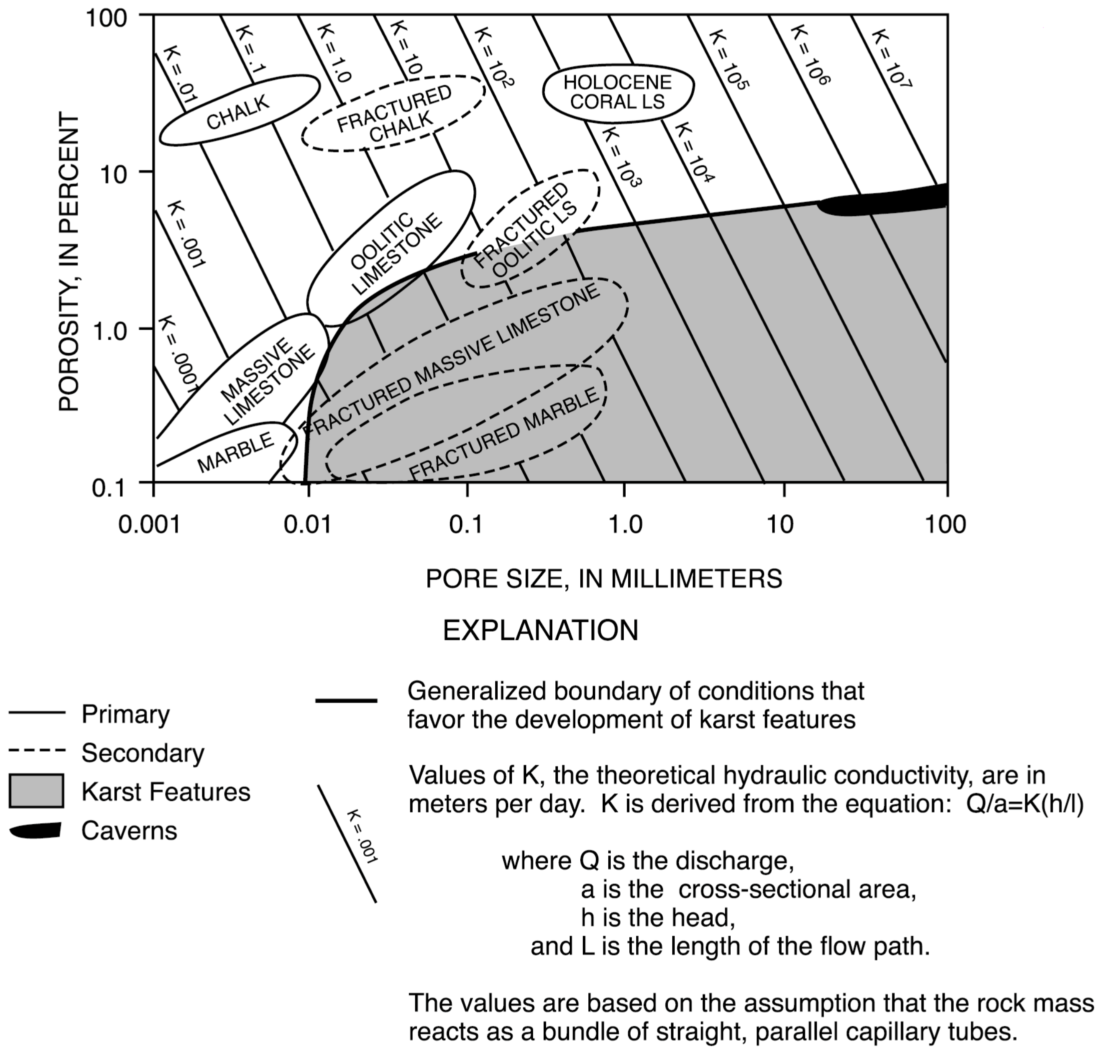


Figure 10. Primary and secondary porosity, pore size, and theoretical hydraulic conductivity of selected carbonate rocks, karst features, and caverns. (From Brahanna et al., 1988)

northeast and northwest quadrants of the compass analyzed by visual inspection because of the small number of readings taken within the study area.

Two joint sets are well developed in the limestone exposures of the study area. Jointing was poorly expressed in the shales due to weathering and slumping in natural outcrops. Set 1 strikes in a N. 60° E. direction and Set 2 strikes at N. 40° W. (Figure 11). Set 1 appears to be more consistent in direction than Set 2, as evidenced by the wider range of variation in orientation and the slightly higher frequency of readings about the apparent mean for the dominant northwest set. In an earlier study of joint patterns in the central part of the Flint Hills region, Ward (1968) took 107 readings of joint orientation in T. 21 S., R. 5 E. in Marion County. He found a dominant northeast joint set (his Set 1) striking at N. 69° E. mean direction and a dominant northwest set striking at N. 25° W. mean direction (his Set 2).

Joint set orientation is frequently but not necessarily related to geologic structure and stress fields (Hobbs et al., 1976). Within the study area, the dominant Set 1 joints do not appear to be directly related to the orientation of local or regional structural features in lower Paleozoic and Precambrian rocks shown in Figure 2. However, the mean strike orientation of Set 2 joints follows the trend of the Fall River tectonic zone of Berendsen and Blair (1986). Ward (1968) compared the joint systems of south-central Kansas with the joint system of the Central Plains of Oklahoma described in Melton (1929) and found similar strike orientations between the northeast and northwest dominant joint sets in both areas. Ward hypothesized that Set 1 and Set 2 joints resulted from the horizontal compressive stresses generated by the wrench-fault tectonics associated with the Ouachita uplift coupled with an opposite force created by wrench-fault tectonics elsewhere in the continent block, possibly in the Rocky Mountain region. Berendsen and Blair (1986) conclude that the structural characteristics in the deep subsurface imply a northward translation of stresses associated with the Ouachita uplift. However, they conclude that the alignment of a series of younger en echelon faults on top of the Elmdale dome (Figure 2) in nearby Chase County suggests formation in a right-lateral wrench zone. These structures may be related to the interaction of the Permian Marathon orogeny or the Late Cretaceous-Early Tertiary Laramide orogeny on the Nemaha uplift.

5.4.2 Observation of Solution Features in Natural Outcroppings

Field observation of solution features in natural outcroppings of the Barneston Limestone was limited to stream cutbanks and stream bottom exposures and locations between drainages where the overlying soil had been removed by erosion. Solution features observed in the study area include stream-bottom sinkholes in the Ft. Riley Limestone, micro-cave development in the Florence Limestone, solution-widened bedding planes and joints, and vuggy limestones in the upper Ft. Riley from weathering.

Four sinkholes, located in Sec. 19, T. 20 S., R. 5 E., were positively identified in the stream bottom of Martin Creek from field observation (Figure 12). Each sinkhole is elliptical in shape with the long axis oriented parallel to the stream and has been excavated into the bedrock surface as a result of chemical and physical weathering of the shaly limestone in the uppermost Ft. Riley Limestone and removal of the weathered rubble by fluvial processes. Sinkhole 1 has two

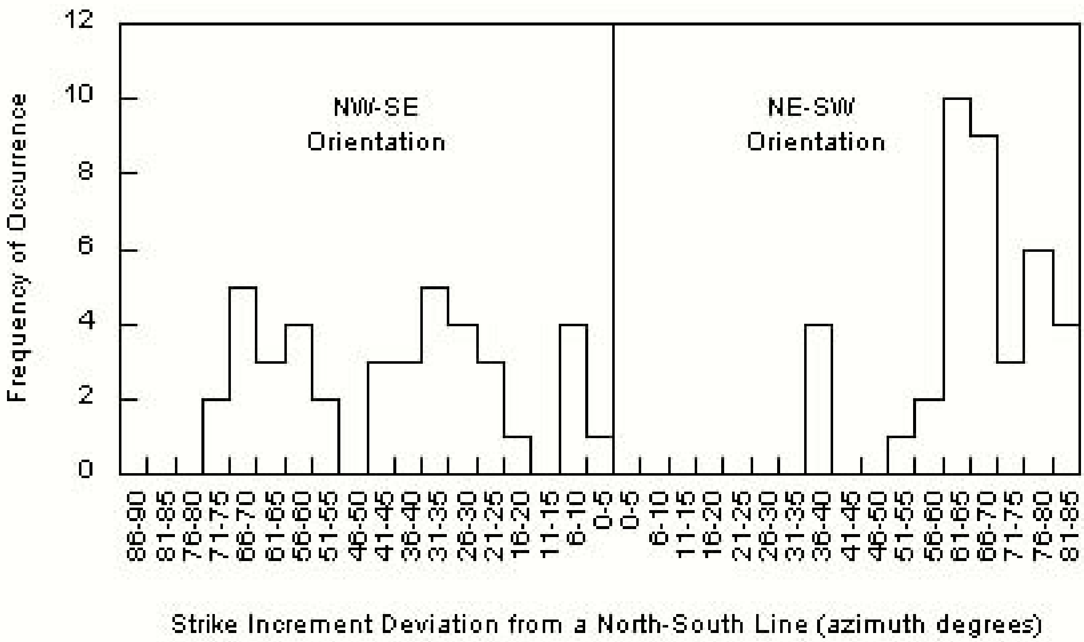


Figure 11. Joint strike measurements taken in the Barneston Limestone in the study area. Directional measurements are grouped into 5 degree increments and referenced to a true north-south line.

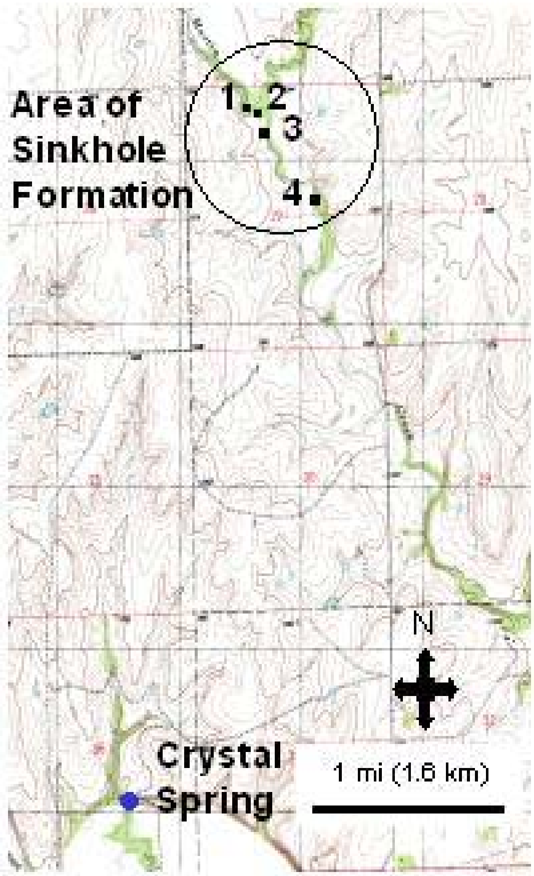


Figure 12. Distribution of known sinkholes in the streambed of Martin Creek with respect to Crystal spring.

openings into the underlying Ft. Riley (Figure 13). The best developed of these openings appears to be at the intersection of the joint surfaces at the bedrock surface. No single orifice could be found at the bottom of sinkhole 2 because the bottom was covered with sediment (Figure 14). At the bottom of sinkhole 3 is a single throat, which is the upper end of a solution-widened fracture or conduit in the underlying bedrock (Figure 15). Along the sides of this sinkhole, solution widened bedding planes and conduits following bedding planes are present and it appears that a block of the bedrock has been partially rotated in place as a result of slumping (Figure 15). Openings into the underlying bedrock through the bottom of sinkhole 4 could not be found because the bottom was covered with fine sediment (Figure 16). The trend of the line of sinkholes is parallel to joint set 2 in the study area.

Solution-widened fractures (includes joints) and bedding planes are also common in outcrops of the Ft. Riley Limestone (Figure 17). For most of the Martin Creek reach that traverses the Ft. Riley outcrop the stream flows directly over the bedrock surface. In stream-bottom exposures the bedrock surface is a pavement of irregularly shaped polygonal blocks of the Ft. Riley separated by solution-widened randomly oriented fractures (Figure 18). Fracture aperture widths near the top of the weathered bedrock surface are on the order of several centimeters or more.

Intense solutioning of the interbedded shaly limestone and mudrocks of the uppermost Ft. Riley was observed in many stream-bottom exposures. At one locality, in SW, SW, NW Sec. 20, T. 20 S., R. 5 E. the weathered limestone is extremely porous and vuggy (Figure 19). Calcite-filled geodes up to 12 cm or more in diameter are also commonly present. The geodes may have formed after the dissolution of evaporites known to be present in this part of the section westward of the outcrop belt (Twiss, 1991).

Solution-widened vertical fractures in the Florence Limestone are rare in natural outcrops and seem to be associated with springs in the study area. Bedding plane discontinuities are not prominent in Florence Limestone outcrops. Near the middle of Sec. 10, T. 20 S., R. 5 E., the upper Hett spring discharges from a well-developed master conduit in cherty limestone near the middle of the Florence Limestone in Bruno Creek (Figure 20). The master conduit opening to the surface is rectangular in shape, approximately 0.6 m high and 0.4 m wide, and appears to have formed by solution widening of a fracture that extends to the top of the stream cutbank. Examination from the surface suggests that the conduit shape and size are maintained for some distance back into the hillside. The floor of the conduit near its mouth is littered with angular pieces of chert and limestone rubble.

5.4.3 Observation of Solution Features in the Sunflower Rock Quarry

Sunflower rock quarry operations have been conducted in two adjacent locations in the southern part of the study area to obtain limestone from the Barneston Limestone for construction. The older, inactive part of the quarry is located in the E 1/2, E 1/2 Sec. 6, T. 21 S., R. 5 E., and the newer, active part is located across the township road in the adjacent W 1/2 Sec. 5 (Figure 1). The lowermost part of the Fort Riley Limestone and the Oketo Shale, and Florence Limestone are exposed in the older part of the quarry, and the Ft. Riley, the Oketo, and the upper part of the Florence are exposed in the newer part.

A:

B:

Figure 13. Sinkhole 1 in the stream bed of Martin Creek (A). Note the outcropping Ft. Riley Limestone near the upper left of the photo. Closeup of one of two points of entry of Martin Creek surface water into the upper part of the Ft. Riley Limestone (B). Entry point diameter is approximately 10 cm. In (A) this entry point is submerged and near the left central part scene in the photo. The other entry point is not visible in (A) and located approximately from where the photo was taken.

Figure 14. Suspected location of Sinkhole 2 in Martin Creek. Flow in the stream channel in the background is from right to left into a small pool in the upper left of the photo. At the time this photo was taken there was no stream flow beyond this pool.

A:

B:

C:

Figure 15. (A) Sinkhole 3 located in the streambed of Martin Creek and partially full of water. (B) Bedrock slumping near the edge of the sinkhole. (C) Main entry point for water into the Ft. Riley Limestone at the bottom of the sinkhole.

Figure 16. Sinkhole 4 located in the streambed of Martin Creek and partially full of water.

Figure 17. Outcrop of the Ft. Riley Limestone near an intermittent stream in NE, SW, SW, NE Sec. 24, T. 20 S., R. 5 E. showing pitting and solution-enlarged joints.

Figure 18. Outcrop of the lower Ft. Riley Limestone in the stream bottom of Martin Creek, NE, SW, NE Sec. 19, T. 20 S., R. 5 E. The stream bottom is best characterized as a pavement of irregularly shaped blocks of limestone. Solution-widened joints and fractures define the blocks and have aperture widths ranging from millimeters to centimeters.

Figure 19. Outcrop of the upper Ft. Riley Limestone in the bottom of an intermittent stream in SW, SW, NW Sec. 20, T. 20 S., R. 5 E. At this location the outcropping shaly limestone is riddled with cavities and geode fillings from the early dissolution of evaporates and the later dissolution of soft, soluble carbonate.

Figure 20. Upper Hett spring located near the middle of Sec. 10, T. 20 S., R. 5 E. The spring empties directly into Bruno Creek from the middle of the Florence Limestone. Spring exit point is approximately 0.6 m high and 0.3 m wide at the outcrop face. The conduit extends back into the outcrop for approximately 2 m before it diminishes in size. Note the sinuous trace of the fracture that extends from the surface to the top of the opening for the spring.

Quarry operations have induced fractures in the limestones. Thus, it was important to distinguish between fractures that resulted from natural causes (joints) from those that resulted from human activity. In most cases, differentiation between natural and human-induced was evident by inspection of the joint surfaces and joint aperture. Fractures resulting from quarry operations tend to anastomose, have very small apertures (on the order of 1 mm or less in width), and the fracture surfaces are typically the same color as the adjacent, freshly exposed limestone.

Solution-widened joints are prominent in the thick-bedded limestones of the lower Ft. Riley, but less so in the thick-bedded limestones within 0.7 m to 1 m base of the Ft. Riley and in the Florence. Joint surfaces are typically iron-stained, scalloped, and have a weathered appearance (Figure 21). Invertebrate fossils stand out in relief on the surfaces suggesting that infiltrating water has preferentially dissolved the fine-grained matrix. Joint aperture widths in the lower Ft. Riley are typically 3-10 mm and 1-3 mm in the Florence. Joint apertures in the Ft. Riley may be partially filled with a red clayey silt residuum from dissolution of the overlying limestone (Figure 22). In several instances this residuum was found in joints 6-7 m below the land surface that existed prior to the quarry.

Geodes are abundant in two zones of the upper Ft. Riley shaly limestone in the quarry and are believed to have resulted from the dissolution of evaporites from this part of the section (Mazzullo et al., 1997). Smaller diameter solution conduits are well developed in a 3.6-m thick zone 0.7 m to 1 m above the base of the Ft. Riley and within thick-bedded limestones. Conduits are vertically to near horizontally oriented, sinuous, smooth-walled and most commonly 2.5 cm to 5 cm in diameter. Several of these features contain a residuum of red clayey silt (Figure 23). Many of these smaller diameter conduits intersect joint surfaces, but do not appear to be genetically related to the solution-widened fractures (Figure 24). Conduits developed within the joint planes are up to 15 cm to 20 cm in diameter and up to 2 m in length where exposed in the joint surface. All of these larger carbonate dissolution features have scalloped surfaces (Figure 25). None of these features were observed in the upper interbedded shaly limestones and mudrocks in the quarry.

A single cavernous passageway was found along the north wall of the quarry in the non-cherty limestones at the top of the Florence Limestone (Figure 26). At this location within the quarry, the Oketo Shale appears not to be present. The exposed part of the cavern is rectangular in cross-section, 1 m to 1.2 m in height, and 0.4 m to 0.5 m in width. The lower part of the cavern is filled with stratified layers of reddish clayey silt. The cavern appears to have developed as a result of joint solution-widening. Extending upward from the cavern is a well-defined solution-widened joint. The joint is vertical in the lowermost thick-bedded basal limestones of the Ft. Riley, but dips at a high angle to east and is curvilinear in the overlying thick-bedded limestones. The joint does not appear to extend into the interbedded shaly limestones and mudrocks higher up in the Ft. Riley.

Near the bottom of the cherty limestone interval in the lower part of the Florence, two vuggy zones each slightly >0.3 m thick and spaced approximately a meter apart. (Figure 27). Vug size ranges from <2.5 cm to approximately 0.3 m in diameter. These zones were encountered during the cable-tool drilling for a monitoring well 4 km north of the quarry (See Chapter 7 of this

Figure 21. Solution-widened joint surfaces in the lower Ft. Riley Limestone in the Sunflower rock quarry in SW Sec. 5, T. 21 S., R. 5 E.

Figure 22. Red clay residuum in a solution-widened joint in the lower Ft. Riley Limestone in the Sunflower rock quarry in SW Sec. 5, T. 21 S., R. 5 E. Aperture width ranges from 5-15 mm.

Figure 23. Red-clayey residuum partially fills a small-diameter conduit near the handle of the hammer.

Figure 24. Small-scale conduit features in the lower Ft. Riley Limestone in the Sunflower rock quarry in SW Sec. 5, T. 21 S., R. 5 E. Conduit diameters are up to 5 cm.

Figure 25. Small-scale conduit features exposed in solution-widened joint face in the lower Ft. Riley Limestone in the Sunflower rock quarry in SW Sec. 5, T. 21 S., R. 5 E. Conduit diameters range from several up to 20 cm. The large conduit length is approximately 2 m in the exposed joint face.

A:

B:

Figure 26. Large-scale conduit feature exposed in the uppermost Florence Limestone in the Sunflower rock quarry in SW Sec. 5, T. 21 S., R. 5 E. Approximate conduit size is 1.2 m high and 0.4 m wide. Lower part of the conduit is filled with 0.4 m of stratified red clayey residuum. Note the scalloped appearance of the limestone just to the right of the opening.

A:

B:

Figure 27. (A) Vuggy limestones near the base of the Florence Limestone and (B) large vug developed in the vuggy limestone horizon in the abandoned part of the Sunflower rock quarry in NE Sec. 6, T. 21 S., R. 5 E. and the SE Sec. 31, T. 20 S., R. 5 E. Approximate vug size is 30 cm high and 45 cm wide.

report). Little ground water inflow from the Barneston to the borehole was observed during drilling until it penetrated these zones.

6.0 Crystal Spring

6.1 Description

Crystal spring is located along the north wall of the Cottonwood River valley, just to the east of where an unnamed tributary drainage enters the valley from the surrounding upland area (Figure 1). The exposed bedrock around the spring site consists of the thick-bedded limestones of the lower Ft. Riley Limestone. A spring house was constructed over the discharge point and thus the point of discharge from the bedrock cannot be observed from outside the building. However, from the surface geology it appears that the ground-water source for the spring is the upper part of the Florence at the same stratigraphic level as the cave in the Sunflower rock quarry.

Prior to construction of the spring house, two 7.6-m (25-ft) deep wells were drilled side-by-side into the underlying Florence Limestone and pumps were installed to more easily produce water for the municipal supply. Water is pumped periodically from the spring to replenish water in the storage tower at a maximum rate of 7.6 L/s (0.27 ft^3/s). Otherwise, spring discharge exits from the spring house through two 0.3-m (1-ft) diameter pipes into a discharge pool before it enters the unnamed tributary drainage. The discharge pool is approximately 10 m (33 ft) in length and 6.5 m (21.3 ft) at its widest point.

6.2 Spring Discharge

6.2.1 Data collection

Continuous monitoring of spring discharge could not be set up during the study. To do so would have permanently altered the discharge pool, which would have been unacceptable to the local residents. Measurements of spring discharge were made monthly and at other times as part of this study in order to determine discharge variability and to assess the influence of individual precipitation events and wet/dry periods on discharge. Spring discharge time series and hydraulic heads in the catchment reflects the complexity of ground-water flow through integrated conduit networks and diffuse flow through karstic aquifers (Smith et al., 1976; Milanovic, 1981).

An alternative approach was adopted by establishing a set cross section across the discharge pool and taking velocity measurements of the flow through the discharge pool at set locations along the cross-section for each calculation of spring discharge. Before the cross section was selected, care was taken to identify all of the sources of water to and all of the exit points from the discharge pool. Stakes were set in the ground at either end of the selected cross section and a u-shaped section of PVC pipe was marked with the locations for exact flow velocity measurement points (Figure 28). Each end of the u-shaped pipe was placed over the appropriate stake prior to making the velocity measurements. Spring discharge velocity measurements were made at each marked station on the PVC pipe along the cross section using a Teledyne Gurley No. 625F pygmy current meter. Spring discharge measurements were made more frequently in the early part of the study. From August through December 2000, 48 discharge measurements were taken

Figure 28. Spring discharge measurements at Crystal spring. Flow velocities and discharge pool depths were measured at points marked on the PVC plastic pipe and used to calculate spring discharge. Measurements were made using a Teledyne Gurley No. 625F pygmy current meter.

at Crystal spring. However in the 2002-2003 project years discharge measurements were made only on a monthly basis. A total of 69 discharge measurements were taken during the course of the study.

6.2.2 Range and distribution of discharge values

Spring discharge ranged from 0.86 ft^3/s (24.36 L/s) on March 3, 2003, up to 18.36 ft^3/s (519.96 L/s) on March 21, 2003 (Figure 29). The distribution of spring discharge values collected during the study is log-normal with a geometric mean value of 2.96 ft^3/s (83.83 L/s). Figure 30 shows the discharge data plotted as a flow duration curve with a best-fit line through the data. The plot shows that spring discharge exceeds 10 ft^3/s (280.32 L/s) less than 4% of the time and is greater than 1 ft^3/s (28.32 L/s) 93% of the time. Spring discharge is generally in the range of the 2 ft^3/s (56.64 L/s) to 4 ft^3/s (113.28 L/s) between 70% and 30% of the time, respectively.

6.2.3 Spring discharge in relation to precipitation

Spring discharge measurement times did not coincide with rainfall events in the catchment but rather were taken as time and opportunity afforded. Overall, measured discharge was generally higher during wet periods and lower during dry periods (Figures 29). Sustained higher discharge seems to coincide with wet periods rather than high yielding individual storm events. Discharge was generally higher during May through mid-July 2002 and late March through early June 2003 and spring discharge was generally lower during the winter months of 2002 and 2003 and in the late summer and fall 2002.

The apparent lack of a discharge response to precipitation events during the drier months could be a reflection of the seasonal change in precipitation patterns (Horsch and McFall, 1983). Typically more than 70% of the annual precipitation received in this part of Kansas occurs between April and September with the highest mean monthly values in April through June. Precipitation events in the spring and early summer months are more strongly influenced by interactions between cold and warm air masses along frontal boundaries, which results in more widespread and more evenly distributed precipitation in the Flint Hills. In contrast, the later summer and fall precipitation events are more often associated with convective storms. These storms may produce locally significant precipitation in some but not all parts of the region. Precipitation is generally lower throughout Kansas in the winter months. Other factors that may have impacted the infiltration of precipitation include evapotranspiration and antecedent soil moisture conditions. During the warmer months there is more evaporation and transpiration of water by plants during the growing season.

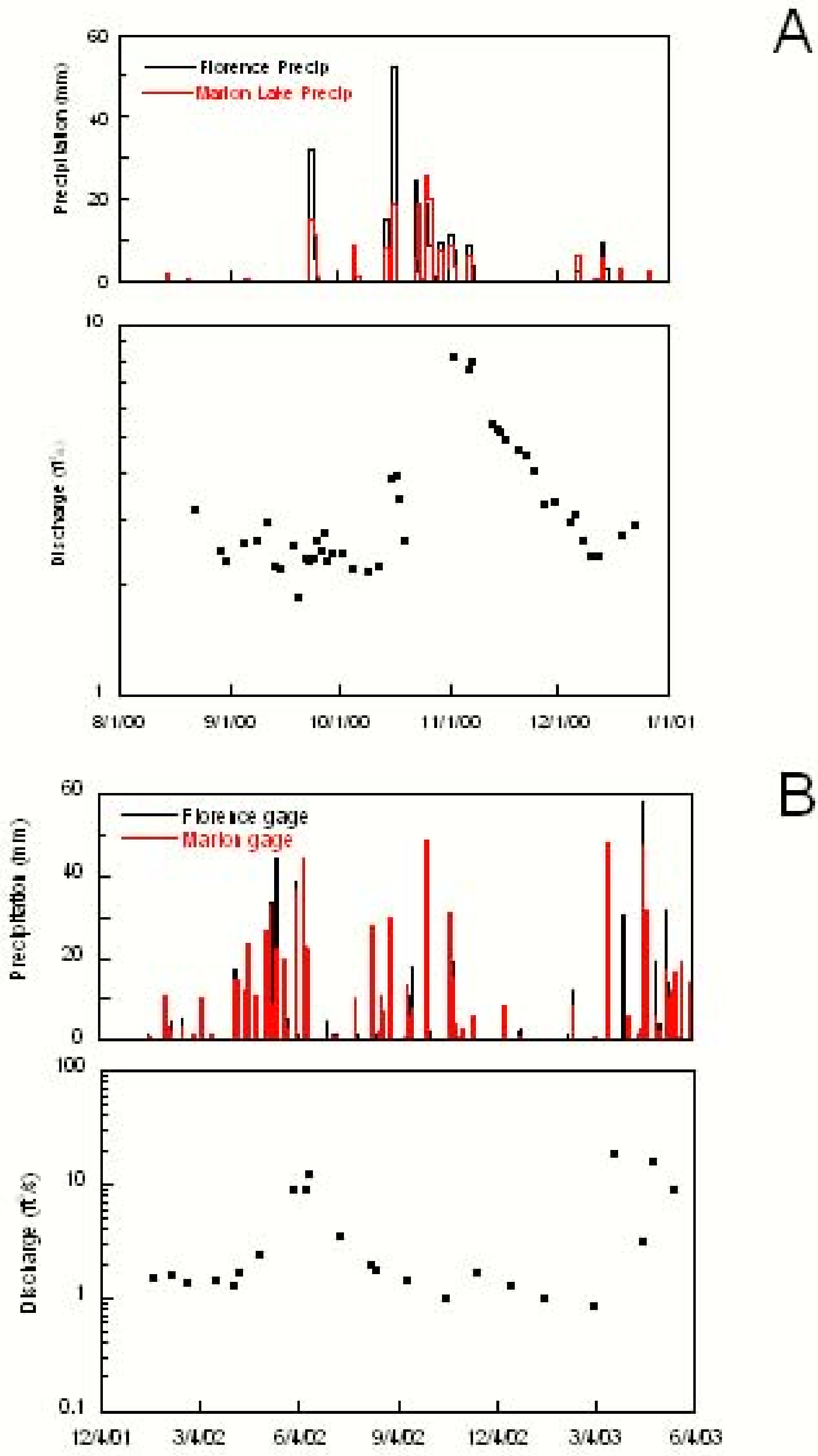


Figure 29. Time vs spring discharge and precipitation at the Florence and Marion Lake weather stations (2000; A) and stream gage sites (2002-3;B).

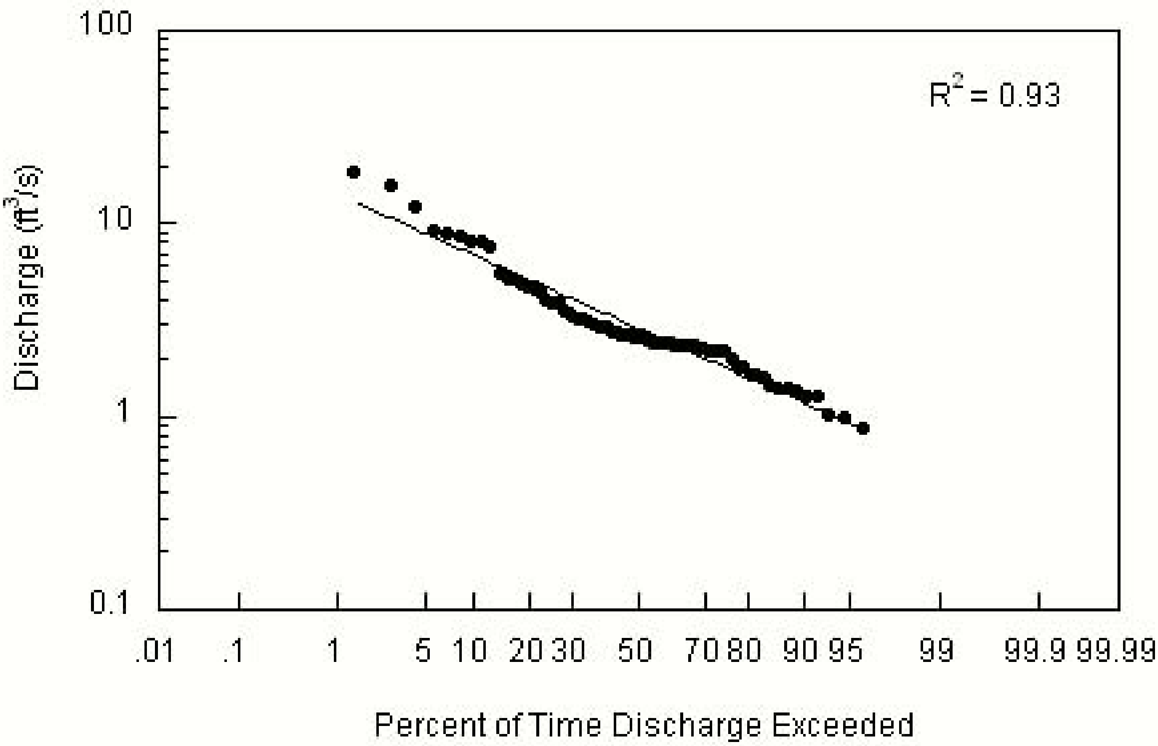


Figure 30. Flow duration curve for Crystal spring based on the discharge measurements made during this study.

7.0 The Monitoring Well

7.1 Location, Purpose

A monitoring well was installed near the line of sinkholes in Martin Creek in SE, SW, NE Sec. 19, T. 20 S., R. 5 E., approximately 210 m east of Martin Creek (Figure 31). Land surface elevation at the monitoring site is 420 m a.m.s.l.

The monitoring well was installed at this location to establish continuous monitoring of water levels near the sinkhole-entry points into the Barneston aquifer. Water-level histories in combination with other data may provide useful information related to karst aquifer structure (Milanovic, 1981). Analysis of water-level data collected from monitoring during the recession from flood-wave passage can provide information on aquifer recharge because of its proximity of the well to the intermittent stream (Moench and Kisiel, 1970).

7.2 Borehole Drilling

The monitoring well was installed by Zinn Water Well Drilling of Lost Springs, Kansas, during November 27-29, 2001. Drilling of the borehole for the well was done using a cable-tool rig with a 20.3 cm diameter bit. Samples of the cuttings were collected for examination every 1.5 m or when a change occurred in the rate of advance of the bit through the rock being drilled. Figure 30 is a sample log of the borehole drilled for the monitoring well. Rapid water loss from the borehole occurred from 8.2-9.1 m, 10.7-12.2 m, and 16.8-18.3 m below surface all within the Ft. Riley Limestone. Vug-fills of calcite crystals were abundant in the cuttings from 4.3-9.1 m in the upper Ft. Riley. This interval is correlated to two zones of shaly limestone with an abundance of geodes described by Mazzullo et al. (1997) in the Sunflower rock quarry, from 1.2-2.4 m and 2.6-5.3 m below the Ft. Riley top. These zones crop out in the bottom of unnamed intermittent tributary of Martin Creek east of the monitoring site in the SW, SW, NW Sec. 20, T. 20 S., R. 5 E. No significant water loss was noted during the drilling of the Florence Limestone. The borehole was drilled to a depth of 30.5 m and ended near the bottom of the Florence Limestone. Ground-water inflow occurred when the borehole penetrated two porous zones near the bottom of the Florence at 25-25.6 m and 26.5-27.4 m.

7.3 Well Installation

The well was constructed of 12.7 cm diameter PVC pipe and the well depth was measured to be 30.42 m (Figure 32). The lower 6.1 m of the pipe was perforated with sawcuts to allow entry of water from the aquifer into the well. A packer was set in the borehole 7.3 m below surface and the annular space from the packer to the surface was filled with cement grout. Bailing tests of the completed well conducted by the driller indicated a potential sustained well yield of 0.15 L/s to 0.38 L/s. The top of the casing is 0.2 m above land surface.

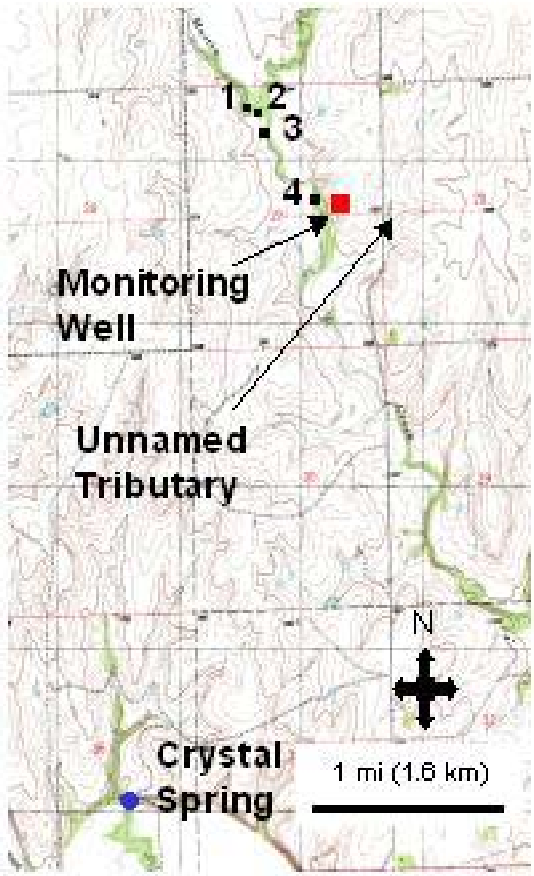


Figure 31. Location of the monitoring well with respect to the sinkholes in Martin Creek, an unnamed tributary east of the creek, and Crystal spring. The monitoring well is approximately 215 m east of Martin Creek and 400 m west of the unnamed tributary.

Depth (ft)	Stratigraphy/ Drill Cuttings Description
	Surface
0-8	Towanda Limestone Member, Doyle Shale Limestone
8-14	Holmesville Shale Member, Doyle Shale Green shale and limestone
14-22	Ft. Riley Limestone Member, Barneston Limestone Soft, white, shaly limestone with manganese coatings on fracture surfaces and vug fillings of crystalline calcite.
22-30	-do- with white, calcareous shale
30-35	Gray, shaly, limestone
35-40	Tan iron-stained medium crystalline limestone and gray shaly limestone
40-45	-do- with dark gray shale
45-50	-do- with tan siltstone
50-55	Tan, shaly limestone
55-63	Gray to tan, iron-stained, fossiliferous limestone
63-66	Oketo Shale Member, Barneston Limestone Tan, fossiliferous limestone: gray fossiliferous shale and shaly limestone
66-70	Florence Member, Barneston Limestone Tan fossiliferous limestone
70-75	Gray fossiliferous limestone and chert with some iron staining
75-80	Gray, fossiliferous, cherty limestone with masses of quartz vug (?) fillings
80-85	Light gray, fossiliferous, porous limestone
85-90	Iron-stained, light tan, very fossiliferous porous limestone with crystalline masses of calcite vug (?) fillings
90-95	-do- with gray chert and gray limestone
95-100	Gray, fossiliferous limestone and chert

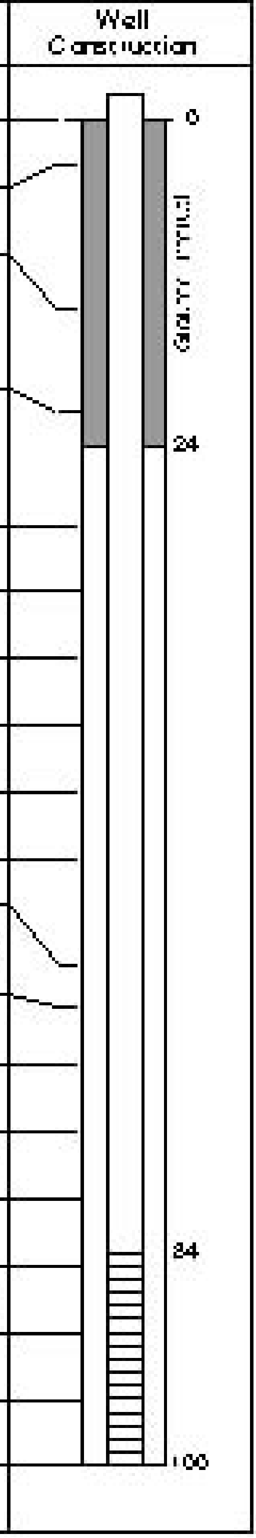


Figure 32. Drill cuttings description and construction information for the monitoring well drilled near Martin Creek in SE, SW, NE Sec. 19, T. 20 S., R. 5 E.

7.4 Equipment for Continuous Measurement of Water Levels

Soon after well installation was completed an In Situ, Inc., MiniTroll water-level monitoring system was lowered into the well and set up to continuously record water-level fluctuations. The MiniTroll used in this study is battery-powered and consists of a 30 psia (2.0685 x 10^5 N/m^2) pressure transducer and data logger contained in a 25.4 cm long stainless-steel sonde. The pressure transducer measures the pressure of the water column above the bottom of the sonde. In the monitoring well, the bottom of the sonde was situated in the lower part of the screened interval, 28.63 m below the top of the well casing. The pressure transducer is vented to the atmosphere at the surface and records only the fluid pressure of the overlying water column in the well. In Situ claims a measurement accuracy of 0.05% of full scale or approximately 0.01 m.

The accuracy of the device was checked periodically during its deployment by comparing the transducer measured water levels with water-level measurements taken manually at the same time using an electric tape. Figure 33 is a comparison of the manual and pressure transducer water-level measurements. Under ideal conditions the measurements taken with the transducer should match the measurements taken manually. A best-fit line through the paired measurement points would have a slope of 1.0 and a coefficient of variation (R^2) value of 1.0. In this case the slope of the best-fit line is 0.993 and the R^2 value is 0.999, which indicates that the pressure transducer measurements are reasonably accurate in comparison to those taken manually with the electric tape.

Water levels were continuously monitored using the MiniTroll from January 11 to April 30, 2002 and from June 14, 2002 to June 30, 2003. The device was out of service from April 30 to June 13, 2002, because of hardware problems.

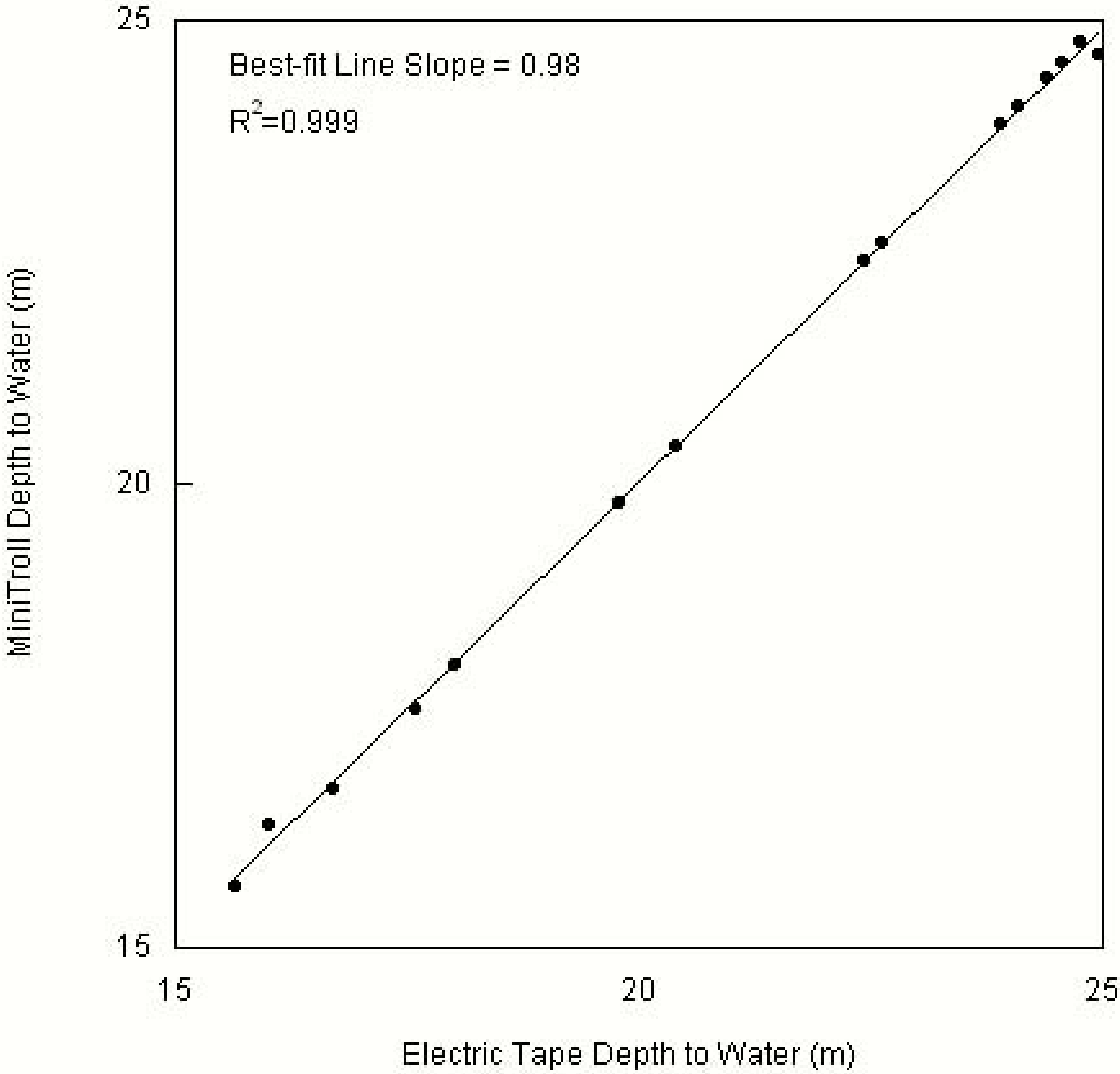


Figure 33. Depth-to-water measurements taken with the MiniTroll compared to measurements taken manually at the same time.

8.0 The Ground-water Flow System in the Barneston Limestone

8.1 Water-level Measurements

Static water-level measurements were taken in 17 non-pumping wells believed to tap ground-water supplies in the Barneston Limestone within the study area in late November and early December 2000 and 2002 (Figure 34). All of the measurements were taken in non-pumping windmill wells or wells with submersible pumps and were made with a steel tape to determine depth to water from a measurement point at the surface. The elevation of the measurement point was estimated from measurements of the height of the measurement point above or below land surface and the approximate ground-surface elevation from a topographic map. These data were supplemented by estimates of spring discharge point elevations, including Crystal spring. Additional manual water-level measurements could not taken in most of the wells during the wetter, late spring period because most of the wells were pumping. However, manual depth-to-water measurements were taken in some abandoned and unused domestic wells periodically during the early part of the study to collect data on water-level fluctuations. Over the course of the study, fluctuations in depth to water in three wells were on the order of 1.5 m or less with higher water levels occurring during the wetter part of the year.

8.2 Potentiometric Surface Map

Without exception ground water is constantly moving from points of recharge where it enters aquifer systems to points of discharge where it leaves these systems. In order to define the pattern of moving ground water in an aquifer, hydrogeologists use measurements of water-level elevation taken in wells scattered throughout the aquifer system's extent. In most cases, the water-level elevation is equivalent to the hydraulic head in a well screened in the aquifer. The hydraulic head is a measure of the potential energy per unit weight of ground water at that point (Freeze and Cherry, 1979). The movement of ground water involves a loss of energy and is similar to the flow of water in river systems. Water naturally flows from elevated regions of the continent (higher hydraulic head) to the ocean basins (lower hydraulic head). In the same way ground water moves from points of higher hydraulic head to points of lower hydraulic head in the aquifer.

The areal pattern of ground-water flow and the location of recharge and discharge areas can be determined from maps showing the variation of hydraulic head in the aquifer. These maps are prepared by plotting the measured water-level elevations (hydraulic head) of all of the wells penetrating the aquifer in the region and drawing lines of equal hydraulic head or equipotentials. Such a map is referred to as a potentiometric surface map in the case of a confined aquifer.

For most situations the direction of ground-water flow is normal to the isolines on the potentiometric surface or opposite the direction of the hydraulic gradient. In karst aquifer systems ground-water flow directions are difficult to determine with much precision for two reasons.

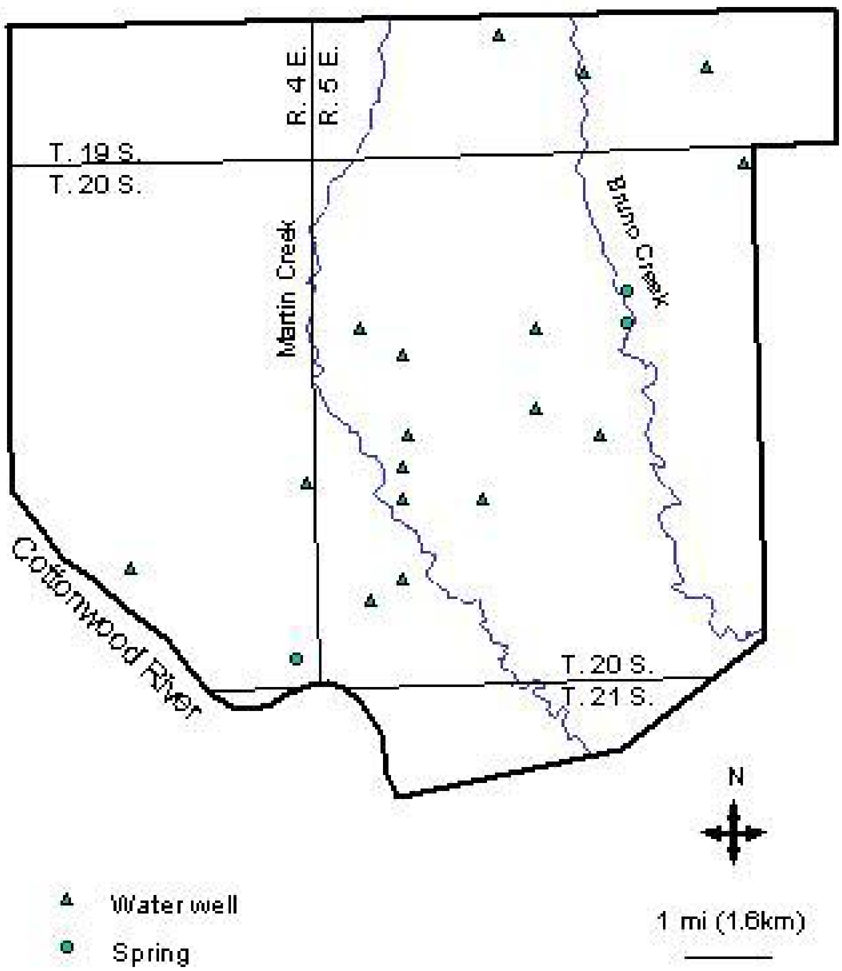


Figure 34. Distribution of hydraulic head measurement points in the Barneston aquifer.

Firstly, flow paths and rates can be readily determined if the flow is through a reasonably homogeneous and isotropic porous medium under laminar flow conditions (Domenico and Schwartz, 1990). In karst aquifers the ground water flows through a heterogeneous and anisotropic network of fractures and conduits (Smith et al., 1976). Most of the void space in karst aquifers is in fractures that have undergone little solution widening and to some extent, Darcy's Law can be applied within these parts (Palmer, 1990). In fractured rocks, a porous medium-equivalent hydraulic conductivity is determined by the fracture aperture width and spacing (Snow, 1968) and the relationship between flow rate and hydraulic gradient is controlled by this equivalent hydraulic conductivity (Domenico and Schwartz, 1990). In most settings, the fractures tend to be oriented around a preferred set of directions, which leads to a strong anisotropy with respect to flow. Palmer (1990) estimates that laminar flow conditions can persist below the water table at fracture aperture widths of up to 1-2 cm under low gradient conditions. Solutional widening of fractures enhances anisotropy and favors closed conduit and open-channel flow under turbulent conditions. Flow under these conditions is directly related to the cross-sectional area, the hydraulic radius, and the hydraulic gradient and inversely related to a friction factor (Palmer, 1990). Under laminar flow conditions, the flow rate is in linear relationship to the hydraulic gradient, but under turbulent flow conditions, the relationship is nonlinear. Thus, the relationship between hydraulic head gradients deduced from observation wells and ground-water flow can be problematic.

Secondly, the observed water-level elevation in observation wells may fluctuate over a wide range of values in response to: (1) seasonal changes from wet to dry conditions or from single precipitation events or (2) hydrologic conditions in the nearest water-filled conduit in the karst aquifer (Milanovic, 1981). Thus, ground-water flow patterns may change and local ground-water systems may interfinger and overlap spatially and temporally with different flow conditions. Water-level elevations may also be difficult to interpret because of monitoring well construction. The interval of the karst aquifer being monitored may hydraulically connect parts of the aquifer that previously were more hydraulically isolated from each other and thus may influence local hydrologic conditions (Milanovic, 1981).

Keeping these limitations in mind, a potentiometric surface map of the Barneston Limestone aquifer was prepared using the limited water-level elevation data from wells and the elevations of spring discharge points (Figure 34). The water level in many of the windmills in the study area was not accessible from the surface and thus a depth to water level measurement could not be made using a steel tape. For basin analysis, Quinlan (1989) suggests a control point density of 1 well per km^2 to adequately define the potentiometric surface of diffuse-flow systems in carbonate rocks.

An approximate limit of saturation in the Barneston Limestone was added to the map based on field observation of the Florence and discussions on finding available ground-water supplies in the Barneston with local residents in the southeastern part of the study area. This part of the aquifer is believed to be dry or possess a very thin saturated thickness. Residents have experienced difficulty in finding a source of shallow ground water in this part of the basin perhaps because of drainage along the nearby Florence outcrop belt.

The potentiometric surface slopes to the south and generally indicates the movement of ground water along the Barneston outcrop belt and from areas where the Barneston is covered by younger Permian rocks toward Crystal spring and the Cottonwood River valley (Figure 35). Water-level elevations are highest in the northeastern part of the study area and lowest at Crystal spring. Water levels also appear to be higher in the upland areas than in either Martin or Bruno creek valleys.

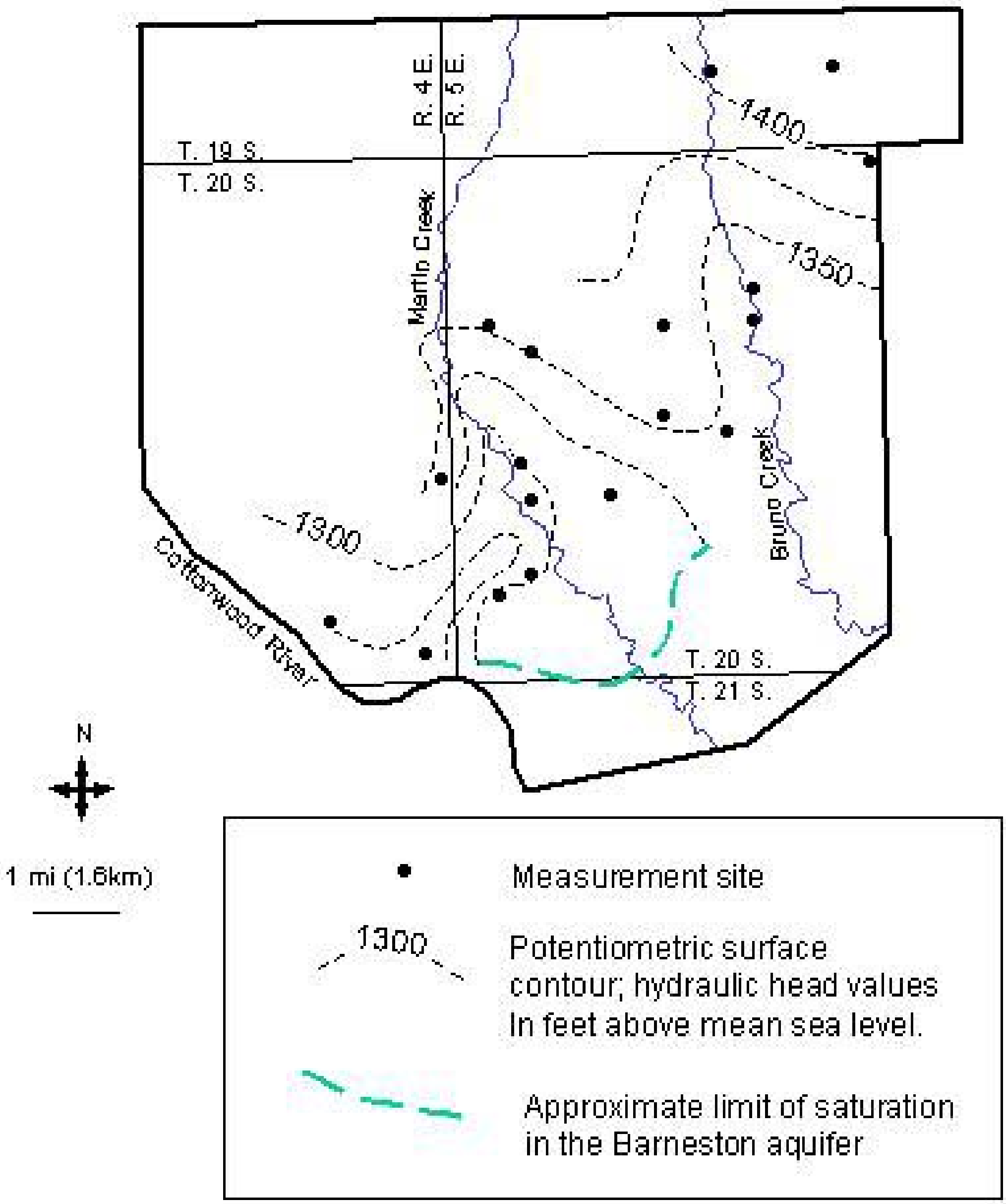


Figure 35. The Barneston aquifer potentiometric surface in the project area showing the approximate limit of saturation.

9.0 Dye Trace Studies

9.1 Rationale for Conducting These Studies

Uncertainty in the interpretation of water-level fluctuations in monitoring wells and potentiometric surface maps makes determination of ground-water flow direction and rate problematic in heterogeneous and anisotropic karst aquifers (Milanovic, 1981). Such information is important in delineating flow paths from recharge areas to points of discharge and flow system boundaries. Furthermore, dramatic temporal changes in the active part of the integrated conduit system can occur in response to precipitation input and cannot be understood from measurements of the hydraulic head only (Milanovic, 1981; Smart, 1988).

To develop a basic understanding of flow systems and the integrated conduit network in a karst terrane requires the use of water tracing techniques using dyes, other solutes, or nonsoluble materials (Davis et al., 1985). In most applications, dye traces are used to establish catchment or local flow paths and system boundaries (e.g. White and White, 1989). With continuous monitoring of the tracer arrival, dye traces can also provide information to develop structural models of the conduit system based on the tracer breakthrough curve at the down gradient end of a flow path (e.g. Smart, 1988).

Early in this study it was hypothesized that the sinkholes in the stream bottom of Martin Creek were hydraulically connected to Crystal spring through a well-developed conduit system. The spring is located approximately 4.25 km (2.64 mi) south of the reach of Martin Creek containing the sinkholes and along the strike of the westward dipping Barneston Limestone. The sinkholes are situated within the upper to middle part of the Ft. Riley Limestone and it is believed that ground water discharges from the Florence Limestone to Crystal spring. The top of the Florence Limestone is approximately 4 m below of the bottom the southernmost sinkhole 4 and approximately 12 m below the bottom of the northernmost Sinkhole 1 in Martin Creek. The potentiometric surface map indicates that the hydraulic head is higher near the sinkholes than it is near Crystal spring. Thus, there exists a horizontal and vertical hydraulic gradient between the sinkholes and the spring. Establishing this hydraulic relationship was considered important to developing a better understanding the hydrogeology of the catchment and also the delineation of the SWAA because the sinkholes are a potential direct entry point of not only recharge but also contamination to the spring.

9.2 Methodology

Dye trace experiments in karst settings must be carefully planned to insure the quality and interpretation of the data collected (Duley, 1997). It is also important to prevent alarming local residents in case the injected tracer does not follow the hypothesized flow path through the subsurface and is pumped from a well to the surface. In this study, a short paper was prepared for KDHE and the city of Florence that outlined the rationale for conducting the dye traces and described the methodology to be used, and described the expected results. The water supply operator was informed of the potential for discoloration of the spring discharge at the intake to city's water supply and a press release that described the dye trace study was prepared by KGS and distributed to the local media in advance of the field testing.

Rhodamine-WT and fluoroscein (uranine) were used in the September 2000 and rhodamine-WT in the October 2002 dye traces of Sinkholes 1 and 3. Both dyes are commonly used fluorescent tracers, easily detectable at low concentrations, and effective for most tracing applications. Fluorescein can lose much of its fluorescence in acidic environments and it is destroyed relatively quickly by exposure to sunlight (Duley, 1997). On the positive side, fluoroscein is not readily adsorbed onto most natural materials, including clays. Rhodamine WT has a lower detection limit than fluorescein and shows only moderate resistance to absorption on aquifer matrix materials (Smart and Laidlaw, 1977). In the laboratory, the spectrofluorophotometer can distinguish the emission peaks for each dye (517 nm for fluorescein and 568 nm for rhodamine-WT). Toxicity testing of both dyes indicates that they are safe in concentrations common to dye recovery sites (Smart and Laidlaw, 1977; Smart, 1984). Duley, (1997) indicates that all fluorescent dyes are destroyed by chlorination and many tracers are adsorbed or destroyed in water softeners. Raw water coming to the city from the spring passes through a fiberglass filter to remove some of the suspended solids and then is chlorinated before it enters the distribution system as drinking water.

The amount of dye (D_a)to be used (in pounds for dry powder or gallons for dyes in liquid form) depends on the type of dye (D_y), the flow dilution potential (F_d), the distance of travel in the subsurface (D_s),and the injection point retardation factor (I_r) (Duley , 1997; Appendix D) and is calculated using the formula:

$$D_a = D_y F_d D_s I_r \qquad \text{(Eqn. 1)}$$

The D_y factors for fluorescein and rhodamine-WT are 0.002 and 0.001, respectively. The flow dilution potential is the square root of the recovery point flow rate divided by the injection point flow rate. The distance factor is the square root of the distance between the injection and recovery points in feet. The injection point retardation factors are 2 for the Sinkhole 1 because of the flow into the entry point and 4 for Sinkhole 3, which in the September 2000 experiment had an artificial flush of approximately 3,700 L of water. According to Eqn. 1, approximately 4.2-4.5 L of 20% rhodamine-WT dye solution was needed for injection at Sinkhole 1, assuming that ratio of discharge at Crystal spring to injection point inflow is at least 25:1. For the other tracer experiment at Sinkhole 3, 1.1-1.2 kg of fluorescein dye powder should be needed according to the above formula. The amounts of dye to be used as calculated by this formula seemed excessive and may be more appropriate for tracer tests performed in much thicker carbonate aquifers and over longer distances, such as in the Ozarks of southern Missouri. On the advice of Jim Van Dyke of the Missouri Division of Geology and Land Survey, Missouri Department of Natural Resources (MDGLS), more conservative amounts of the dye were used in the initial testing to avoid any problems with city water users in Florence. Only 0.95 L of the 20% rhodamine-WT dye solution and 0.23 kg of the fluorescein dye powder were injected in the September 2000 tests. It was decided that if the dye did not appear at the Crystal spring two months after injection, the tracer experiment would be repeated using larger amounts of the dye, depending on the local rainfall pattern and the discharge at the spring.

In the September 2000 tests, Crystal spring was monitored passively using activated coconut charcoal bugs to absorb the dyes present in the discharge water. With this monitoring technique the activated charcoal absorbs dye continuously if it is present in the water. Each bug consists of a 5 cm by 7.6-cm fiberglass, screen-wire packet that contains about 15 cm^3 of 6-14 mesh, activated coconut charcoal. The advantage of using this monitoring technique is that it is inexpensive and requires only occasional site visits to recover the bug and install a fresh one for continued monitoring. The activated charcoal packets have several advantages over water samples. They absorb dye continuously, even below water-sample detection limits because the charcoal will effectively concentrate the dye. The packets can be changed at frequent intervals for accurate time-of-travel data if travel to and from the site is not an issue. Alternatively, the packets can be left in place for several weeks, if the purpose of the experiment is only to establish ground-water flow paths, as it was primarily in this series of experiments. The charcoal bug was placed in the discharge pool directly in the flow of water leading from the spring house. This insured constant water movement through the bug to facilitate absorption of the dye onto the charcoal. Plastic-coated steel wire was used to attach the packet to rocks in the stream bottom and secure each packet in the path of the flow from the spring. Prior to the start of the dye-tracing experiment, the background fluorescence in the spring was monitored for 2 weeks. It was not known if the spring discharge contained naturally fluorescent compounds. The background fluorescence data was also useful for interpretation of laboratory results.

The frequency of charcoal packet retrieval and replacement hinged on a good estimate of the arrival time of the tracer at the spring after injection into the sinkholes. It was also recognized that this estimate would be highly uncertain because the velocity and the path taken by the dye between entry and exit points were unknown. Also, the dyes used in tracer tests tend to travel slower than the ground water due to absorption on the aquifer matrix and on clays (Davis et al., 1985). On the basis of discussions with Jim Van Dyke (MDGLS) the travel time between the sinkholes and Crystal spring was estimated to be 10-20 days. Accordingly, it was decided that monitoring for background fluorescence would continue after the start of the dye trace experiment with retrieval/replacement daily for up to one month.

As each charcoal packet bug was retrieved from the spring discharge, it was logged in, assigned a sample number, and sent by overnight mail in light-tight containers to the Division of Geology and Land Survey, Missouri Department of Natural Resources, Rolla, Missouri. At the laboratory, standardized procedures were followed to measure for the presence of adsorbed dye on the charcoal (Duley, 1997). The packets were washed under a high-velocity water jet to remove sediment and extraneous material, and were opened. The charcoal was removed, was placed in plastic specimen containers. The charcoal was then elutriated with a 5% solution of ammonium hydroxide in ethyl alcohol to release the dye from the charcoal. After an hour, 4 ml of elutriant is pipetted from the liquid, placed in a sample holder, and analyzed using a spectrofluorophotometer. The samples were scanned and the results presented as spectrofluorograms which are interpreted by visual inspection.

The disadvantage of monitoring with either water samples or charcoal bugs is the difficulty of conducting high-frequency sampling. The true character of the breakthrough curve may also be misrepresented if the sampling rate at the monitoring point is not sufficiently high (Smart, 1988). According to Smart (1988) the processes that affect the form of the breakthrough curve include dispersion, divergence and convergence of dye-bearing flow paths, dilution caused by convergence with a dye-free flow path, and storage of tracer dye in branching conduits with low velocity flow.

For the October 2002 test, monitoring at Crystal spring was done using a YSI 6130 Rhodamine-WT optical sensor mounted on a model 6600 Extended Deployment System for onsite, continuous data collection on loan from YSI. The sensor has a detection limit of 0.5 μg/L, an accuracy of 1 μg/L, a range of 0-200 μg/L as true dye concentration, and is not affected by turbidity or chlorophyll interference. The instrument was calibrated using the 2-point method (0 and 200 μg/L) prior to deployment to the field and the calibration was rechecked on completion of monitoring to test for instrumental drift. The sampling rate for the sensor was set at 15 min and the sampling period length was 2 weeks.

9.3 September 2000 Dye Trace Results

Sinkhole 1 and Sinkhole 3 are located 4.4 km and 4.29 km north of Crystal spring, respectively (Figure 36). On September 19, 2000, fluorescein dye was injected into Sinkhole 3 at 1145 and rhodamine-WT dye into the Sinkhole 1 at 1230. On September 20 at approximately 0930, the city water supply operator noticed that the water in the spring discharge pool was reddish brown in color and reported this to KGS. The water supply operator noted that the water was green in color the following morning (September 21) and also reported this to KGS personnel. From these observations, the maximum travel times are estimated to be 45 hrs from Sinkhole 1 to the spring and 69 hrs from Sinkhole 3 to the spring. The longer travel time for the fluorescein dye may have resulted from retardation due to adsorption on aquifer materials (Davis et al., 1985) or from reduced flow rates through the aquifer as indicated by discharge measurements taken at the spring. During the dye trace experiments, spring discharge was 72 L/s (September 18), 52 L/s (September 20) and 66 L/s (September 22). Analysis of the dye packets retrieved after each visual sighting confirmed the arrival of both dyes at the spring. It was concluded that more than enough dye had been used in this test.

9.4 Results and Analysis of the Data from the October 2002 Dye Trace

At 1057 October 17, 2002, approximately 0.3 L of 20% rhodamine-WT dye solution was injected into Sinkhole 1 and the spring discharge was 28 L/s. Sinkhole 1 was being recharged by streamflow from Martin Creek at the rate of approximately 0.1 L/s when the trace was initiated. Figure 35 contains plots of rhodamine-WT concentration in the discharge from Crystal spring, precipitation, and depth to water in the monitoring well vs. time. A check of the calibration indicated that instrumental drift was less than 0.1 μg/L during the monitoring period. A total rainfall of 0.5 mm was recorded at the stream gage on the Cottonwood River between the time of dye injection and arrival at the spring (Figure 37). Breakthrough of the dye at the spring occurred at 2130, October 19 or 58.5 hrs after injection. The maximum concentration in this first pulse of dye to arrive at the spring was 1.9 μg/L at approximately 60 hrs after dye injection. The

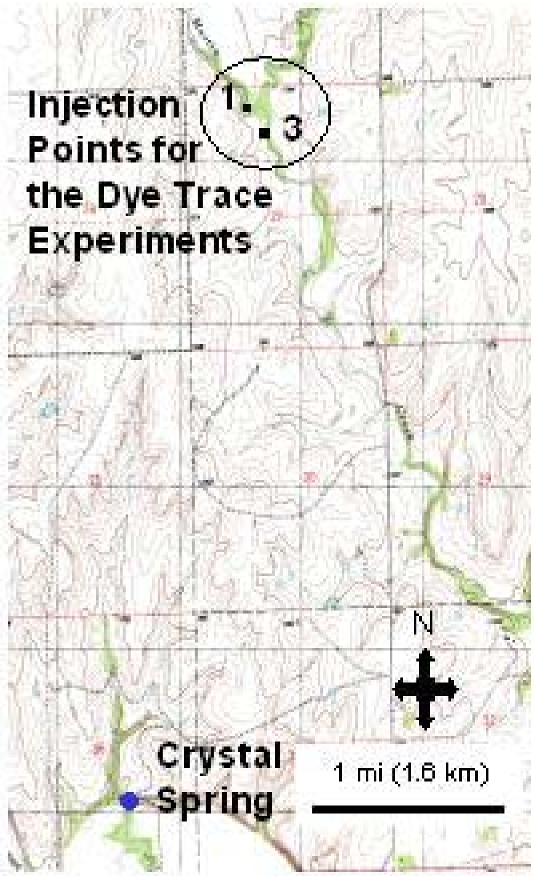


Figure 36. Location of sinkholes 1 and 3 with respect to Crystal spring and used as injection points in the dye trace experiments.

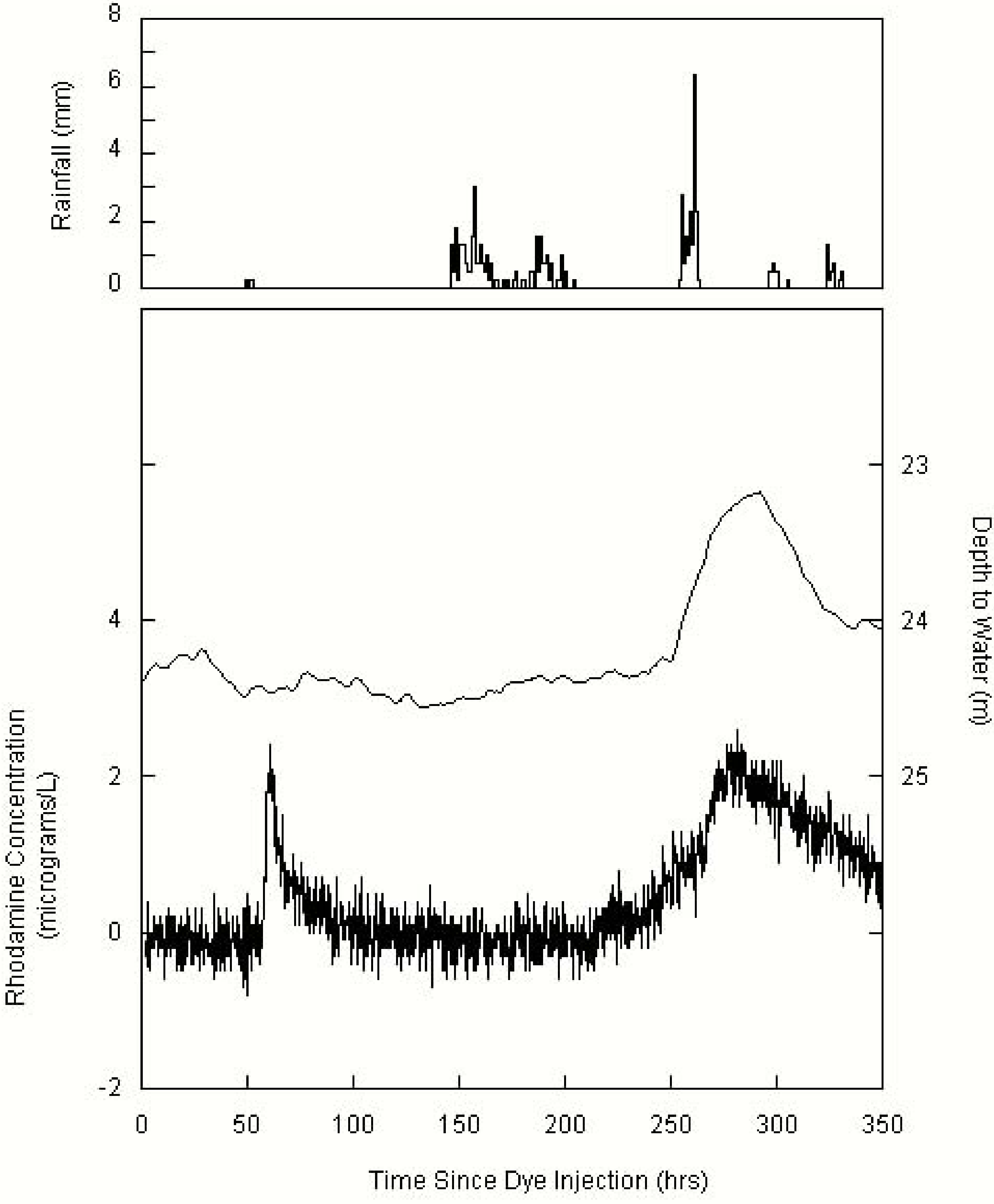


Figure 37. Time vs rhodamine-WT concentration in the discharge from Crystal spring and monitoring well depth to water.

first pulse of dye passed completely through the spring discharge at approximately 100 hrs after injection.

At approximately 124 hrs after injection, a series of rainfall events resulted in 56.9 mm of precipitation at the stream gage on the Cottonwood River. This series of rainfall events caused a rise in water levels in the monitoring well adjacent to Martin Creek and the passage of a second dye pulse through the spring from 239 hrs after injection to the end of the monitoring period at 360 hrs after injection. It is believed that the source of the second pulse is the dye that adsorbed onto the clay silt that had accumulated in the conduit system. The maximum concentration of the dye measured by the sensor in the spring was approximately 2 μg/L. Unfortunately, monitoring was discontinued before this pulse passed completely through the spring.

The solution that was injected into Sinkhole 1 to start the dye trace contained approximately 59 gms of rhodamine dye. Approximately 2.36 gms of the undiluted dye passed through the spring in the first pulse from 58.5 to 89.55 hrs after injection and 15.77 gms from 245 hrs after injection to the end of the monitoring period. Thus, a minimum of 17.69 gms or approximately 30% of the injected dye arrived at the spring after the 2-week period of monitoring. From the dye concentration vs. time plot it appears that most of the second pulse had passed through the spring by the end of the monitoring period (Figure 35) with a small additional amount of dye remaining to be discharged through the spring. This small recovery indicates that a large fraction of the injected dye has been adsorbed onto fine sediment, organic matter, and the aquifer materials. Repogle et al., (1966) reported a 28% loss in rhodamine-WT concentration onto bentonite clay. Smart and Laidlaw (1977) found significant adsorption of rhodamine-WT onto limestone sediment (up to 34% for sediment concentrations of 20 mg/L) and humic materials (up to 89% concentrations of 20 mg/L) in a series of laboratory experiments with different tracer dyes. Dye adsorption onto sediment surfaces is considered to be mostly an irreversible process (Smart and Laidlaw, 1977) and thus flushing of the dye from the aquifer depends on the removal of trapped sediments from the aquifer by high-flow events.

Using the dye concentration vs. time data for the first dye pulse, estimates of the average residence time and the average velocity of the dye in the aquifer can be made. The average residence time of the dye is the centroid of the area under a mass recovered vs. time curve with respect to time (Atkinson, 1982):

$$\text{Average Residence Time} = \frac{\int_0^\infty QCtdt}{\int_0^\infty QCdt} \qquad \text{(Eqn. 2)}$$

where Q is spring discharge, C is the dye concentration, and t is the time since injection. A residence time of 68.3 hrs in the aquifer was calculated for the dye in the first pulse. Using the average residence time and assuming a straight-line distance of 4.29 km (2.64 mi) between the injection point at the Sinkhole 1 and the spring, the average flow velocity of the tracer in this first pulse was approximately 1,500 m/day.

The Reynolds number (N_R) is a ratio of the inertial to the viscous forces during flow and is an index that is used to determine if ground-water flow is occurring under laminar or turbulent conditions. This parameter is calculated as (Freeze and Cherry, 1979):

$$N_R = \rho v d/\mu, \qquad \text{(Eqn. 3)}$$

Where, ρ and μ are the density and viscosity of the water, d is a representative length dimension usually taken as the average conduit diameter, and v is the ground-water velocity. Ground-water flow is considered to be laminar for Reynolds numbers in the range of 1 to 10 and turbulent for values greater than 10. From field observation of the Florence Limestone conduit dimensions range in diameter from 0.15 m to 1.22 m. Applying this range of conduit diameters and the average ground-water flow velocity yields Reynolds numbers of approximately 2,300 and 189,000, respectively. These high values indicate that ground-water flow is occurring under turbulent conditions (Freeze and Cherry, 1979). Note that the discharge, and by extension, the flow rate in the aquifer are near the low end of the values recorded in this study (Figures 29, 30), which implies that ground-water flow is well within the turbulent range at higher discharge rates.

9.5 Comparison of the Results with Those of Other Dye Trace Studies

The average ground-water flow velocity calculated from the October 2002 dye trace between Sinkhole 1 and Crystal spring was approximately 1.5 km/day. Dye trace studies have also been conducted on other karst aquifers in a wide variety of settings in other parts of the world for many years and have yielded a wide range of flow velocities. Only a few of these are mentioned here. Smith et al. (1978) report mean flow velocities ranging from 3.45 to 7.36 km/day for karst aquifers in the White Limestone, Jamaica, and the Carboniferous limestones of the Mendip region, United Kingdom. In the former Yugoslavia, Milanovic (1981) reported ground-water velocities ranging from 1.7×10^{-3} km/day to 47.7 km/day. In the Ozark region of United States, Vandike (1992) reported ground-water velocities ranging from 0.1 km/day to 4.44 km/day that derived from dye trace studies in Ordovician carbonates.

10.0 Water-level Fluctuations in the Monitoring Well

10.1 Monitoring Well Hydrograph

Water-level data were collected continuously from the monitoring well during two time periods using an In Situ MiniTroll. The first period of data collection began on January 11, 2002 and ended on March 20, 2002 due to equipment problems. The second period of monitoring began June 14, 2002 and ended June 11, 2003. For both monitoring periods water-level measurements were taken hourly and stored in the device for later downloading to a laptop computer. Figures 38, 39 present the record of water-level fluctuations from both monitoring periods. Depth to water measurements are referenced to the top of the well casing at the surface.

10.2 Factors Influencing Large Scale Fluctuations

In the January-March 2002 hydrograph, the depth to the water level in the well shows an obvious slight downward trend from approximately 24.84 m to 24.99 m (81.5 ft to 82 ft) from the top of the casing (Figure 38). Throughout this period the discharge from Crystal spring was also decreasing. Precipitation events were few and low yielding during this monitoring period and would have produced little if any recharge to the Barneston aquifer. Pumping from the aquifer is for domestic and stock use from a small number of low-yielding (less than 0.57 L/s [0.02 ft^3/s]) wells and the major discharge from the aquifer is through Crystal spring. Thus, the downward trend in water levels results from the discharge of water from the aquifer to Crystal spring.

In the June 2002 to June 2003 hydrograph, the depth to water in the well varied over a wide range from 15.35 m to 24.39 m (50.37 ft to 80.03 ft) from the top of the casing (Figure 39). In the early part of the hydrograph, the water level was high as a result of recharge to the aquifer from the preceding, wet spring period. Following this period up to late October 2002, the depth to water in the well increased over time to approximately 24.38 m (80 ft) below the top of the casing. During this time, high yielding storm events did occur in the basin, but did not significantly impact the water level in the monitoring well. In early November a series of storms passing through the study area resulted in a slight rise in the water level in the monitoring well. Following this series of rainfall events until late March 2003, precipitation events were few and low yielding and the hydrograph shows a slight decline in the water level during this period down to the levels observed during the January-March 2002 period (Figure 38). Periods of heavy rain in late March and April 2003 caused the water level in the monitoring well to rise up to the levels observed at the beginning of this monitoring period in mid-June 2003.

From field observation it appears that the high water levels in the monitoring well coincide with periods when there is streamflow in Martin Creek. Abrupt water-level rises seem to follow stormy periods of several days duration (Figure 39). The creek was not observed following the late October-early November 2002 rise in water level until the middle of November. At that time it was observed that Sinkhole 3 was dry and there was flow into a pool of water above Sinkhole 1 that had not been there in mid-October.

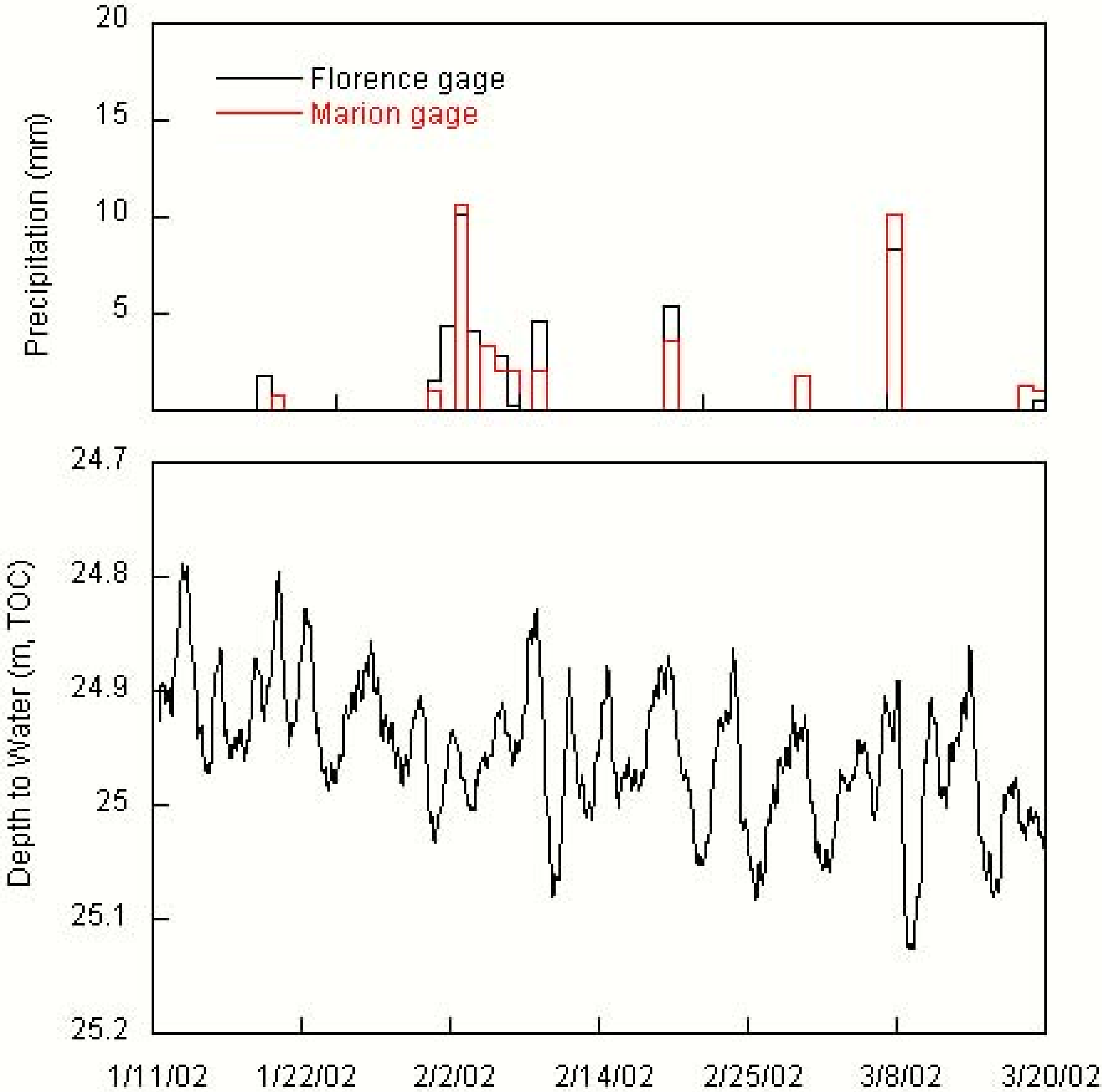


Figure 38. January-March 2002 monitoring well hydrograph with precipitation for the period.

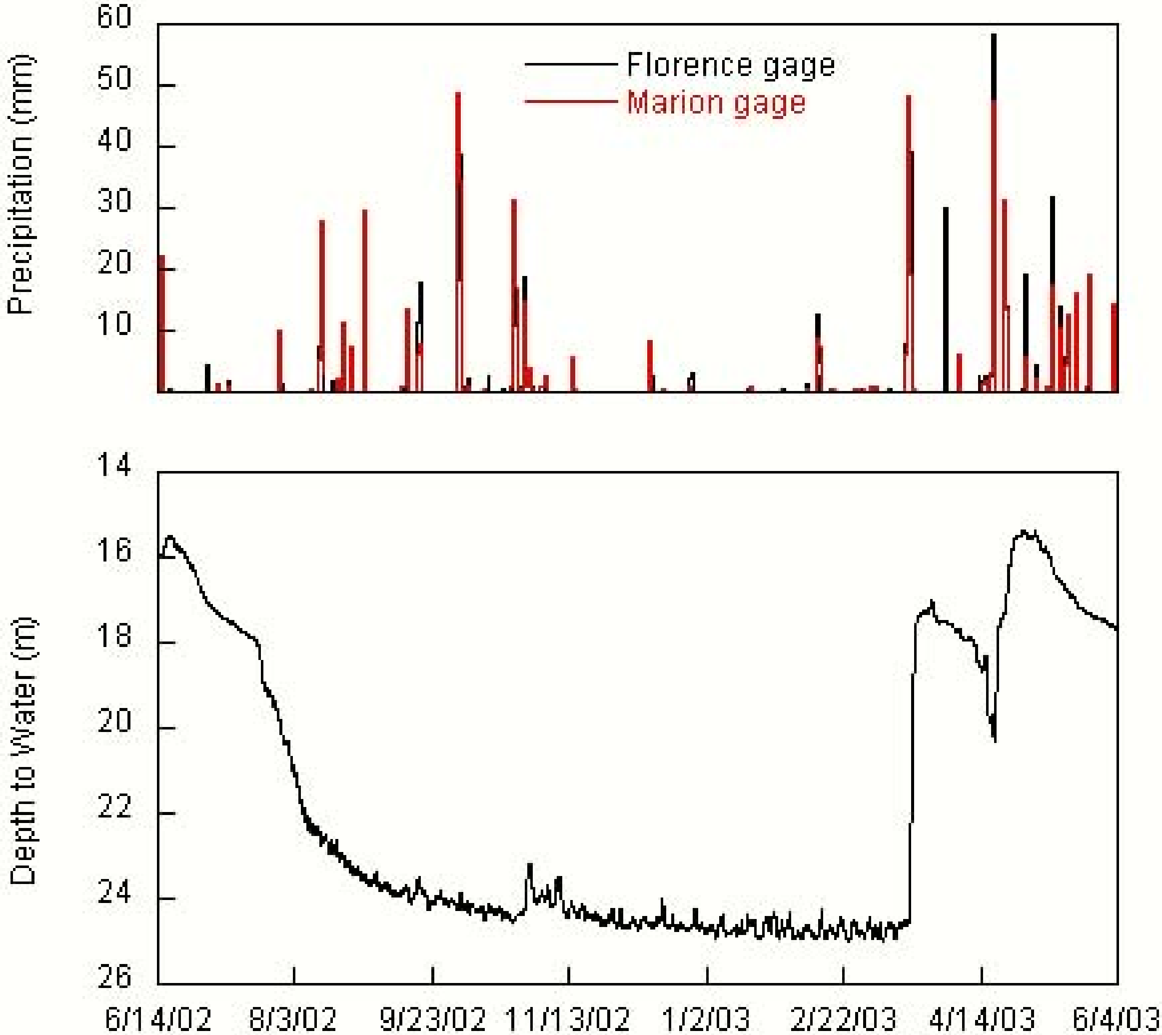


Figure 39. June 2002-June 2003 monitoring well hydrograph with precipitation for the period of record.

Water can be heard cascading into the monitoring well below the packer (7.3 m [24 ft] below surface) from the upper part of the Ft. Riley Limestone during periods of high water levels. Below the packer, the well is cased but the annular space above the well screen is open. In the recession following the Spring 2002 series of recharge events, water cascading into the well could still be heard at least until mid-August when the water level was below the top of the Florence Limestone at approximately 22.9 m (75 ft) below the top of the casing. Cascading water in the well also followed recharge events in late March 2003 and continuing until monitoring ended in June 2003.

The monitoring well is located approximately 213 m (700 ft) east of Martin Creek and approximately 427 m (1,400 ft) west of an intermittent stream, which is tributary to the creek (Figures 31, 40). The elevation of the streambed can only be estimated approximately from the 1:24,000 scale Lincolnville SW 7.5-min quadrangle, topographic map. From the topographic map, it appears that the highest water levels observed in the monitoring well are approximately 1.5 m (5 ft) higher than the streambed immediately to the west of the monitoring well. This elevation difference is approximately the height of the water surface in the stream at bankfull conditions. The distribution of flood debris on the adjacent floodplain following flood crests suggests that bankfull and flood conditions sometimes occur as a result of stormy periods. Inflow from the stream through upstream sinkholes could also impact the water level in the monitoring well. The entry points for flow are near the top of the Ft. Riley Limestone at the upstream Sinkholes 1 and 2 and near the middle of the unit at Sinkhole 3. The small rise of water level in the monitoring well associated with the late October-early November 2002 storm events and the development of a pool above Sinkhole 1 suggests that recharge to aquifer from this point source may influence water levels in the aquifer downstream. Thus, the rise of water level in the monitoring well is at least in part attributable to streamflow events in Martin Creek.

The tributary stream immediately to east of the well may also be a source of delayed recharge to the aquifer near the monitoring well. This source most likely contributes some of the water that cascades into the well when there is no longer streamflow and seepage from Martin Creek into the Ft. Riley has ceased. In this part of the drainage the shallow bedrock units dip slightly to the west-southwest. The exposed upper part of the Ft. Riley Limestone in the creek bank is weathered and highly porous from limestone dissolution with an abundance of calcite-lined geodes from evaporite dissolution. Seepage into the limestone in creek bank from storm events could move downdip from the tributary and drain into the well. Field observation indicates that flow in the intermittent tributary ceases within a few days following wet periods. The long period of time that water continues to cascade into the well following wet periods suggests that other sources also contribute water to the well.

10.3 Factors Influencing Small Scale Fluctuations

Small-scale water-level fluctuations correlate with changes in barometric pressure recorded hourly at the Wichita Midcontinent Airport, approximately 103 km (64 mi) south of the monitoring well (Figures 41, 42). The hydrograph from June 2002-June 2003 shows that the influence of barometric pressure changes is more pronounced when water levels are below than above the top of the Florence Limestone (Figure 42). The degree to which atmospheric pressure

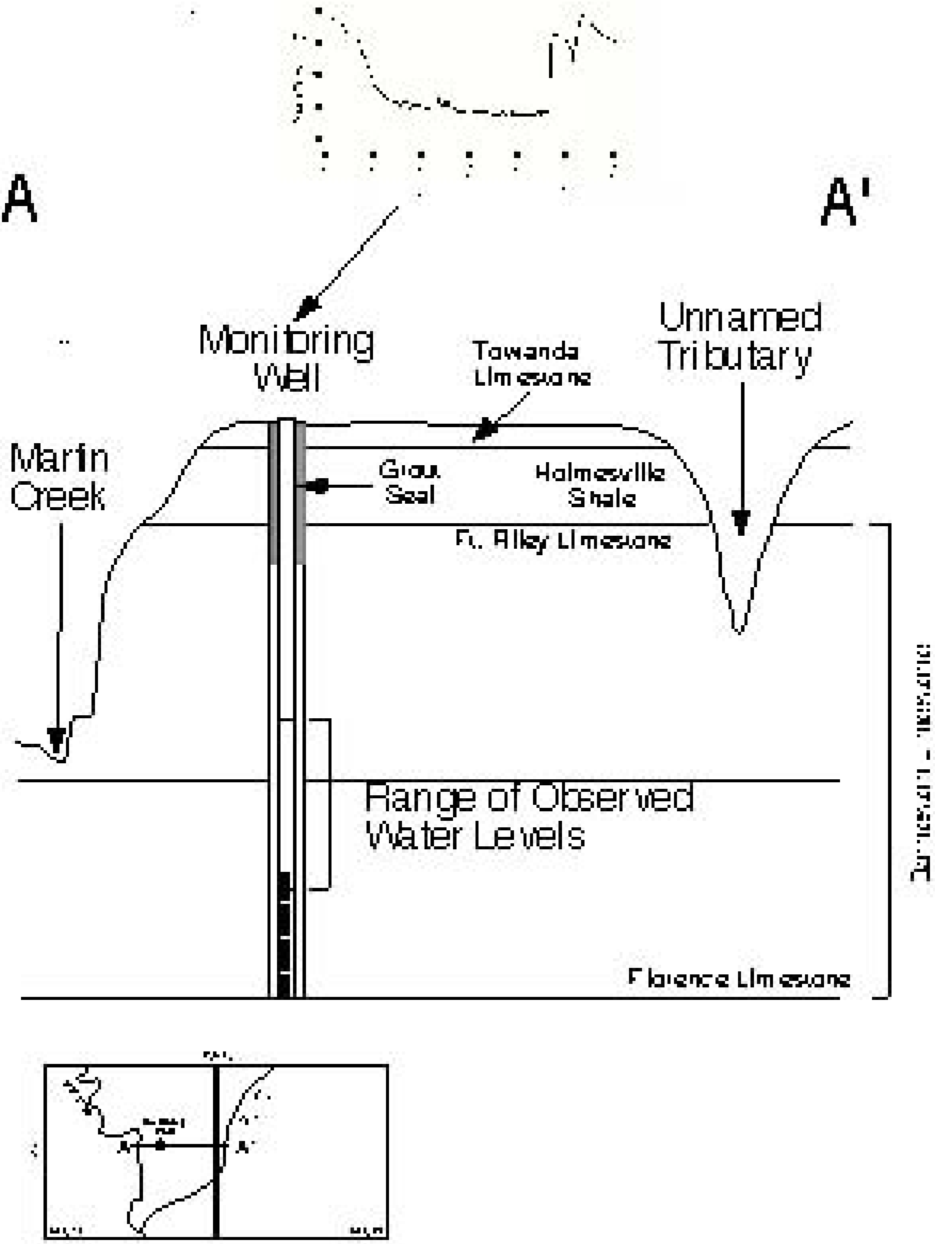


Figure 40. Vertical cross section from Martin Creek eastward through the monitoring well and extending to the unnamed Martin Creek tributary in SW, SW, NW Sec. 20, T. 20 S., R. 5 E.

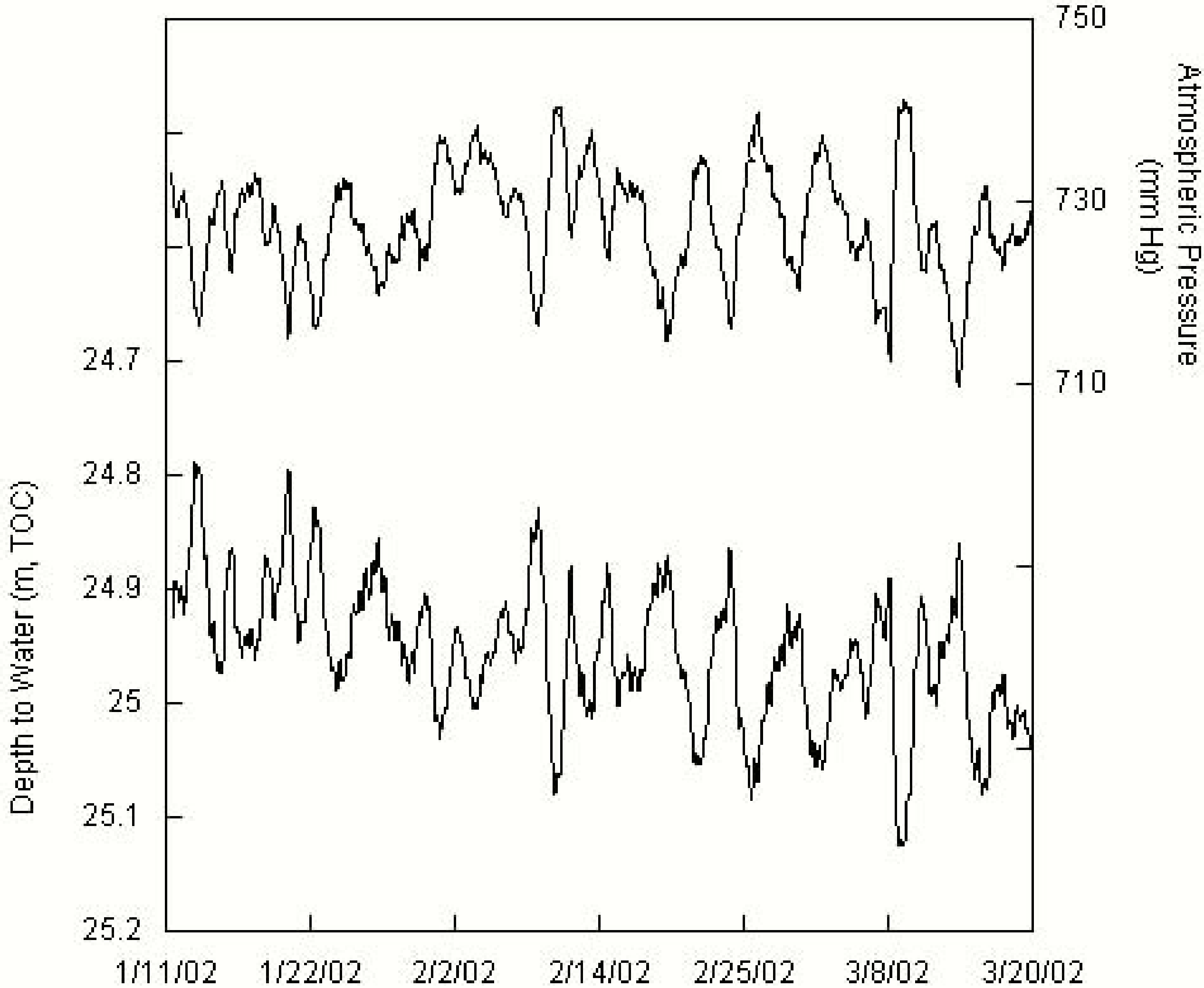


Figure 41. January-March 2002 monitoring well hydrograph with the record of atmospheric pressure fluctuations measured at the Mid-Continent Airport, Wichita, Kansas.

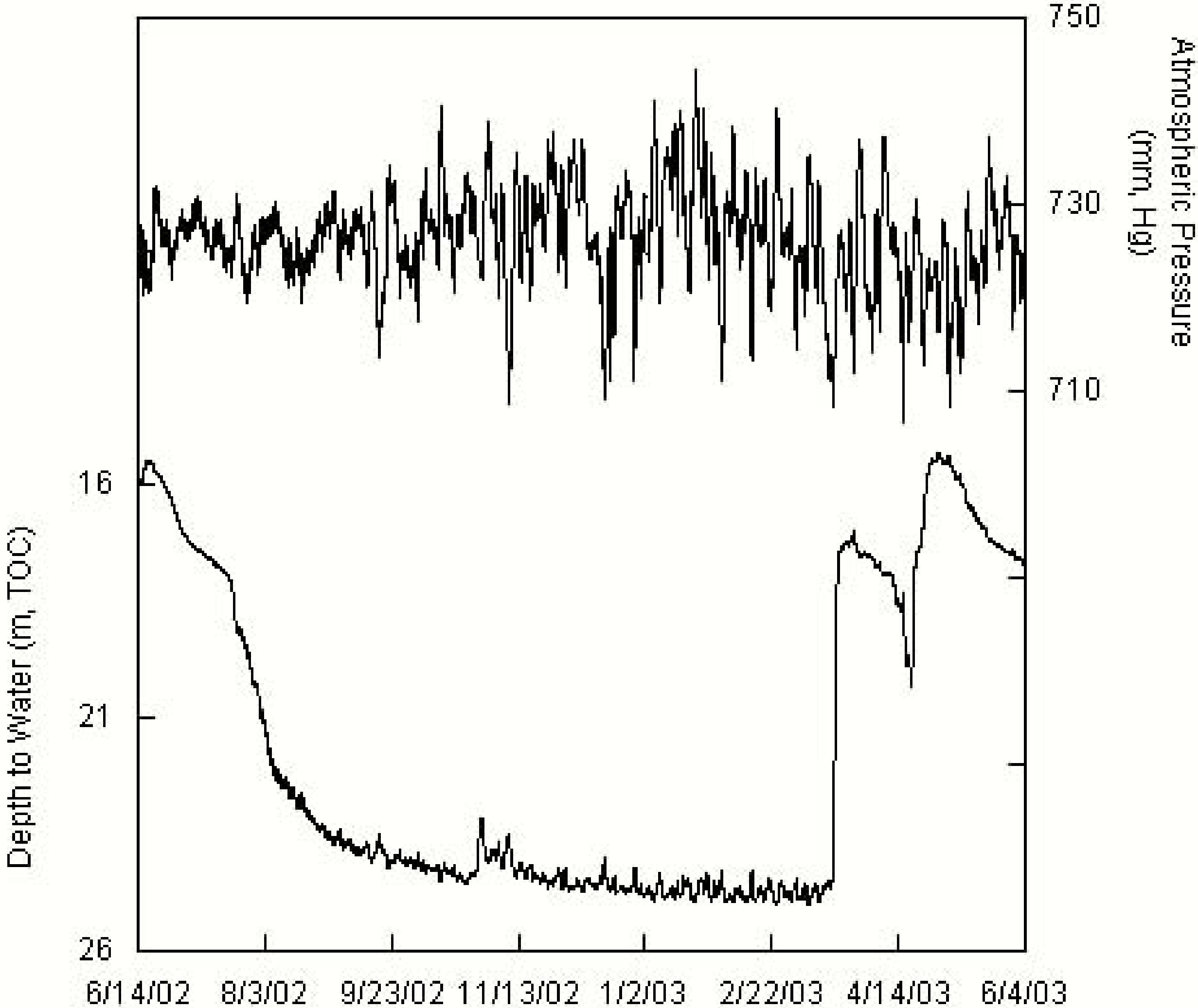


Figure 42. June 2002-June 2003 monitoring well hydrograph with the record of atmospheric pressure fluctuations measured at the Mid-Continent Airport, Wichita, Kansas. Depth to water is measured from the top of the well casing (TOC).

influences water levels is usually an indication of the level of confinement of the aquifer (Domenico and Schwartz, 1992). Under unconfined conditions, changes in atmospheric pressure are transmitted equally to the water table through the unsaturated zone and to column of water in the well, which results in the lack of a pressure differential between the well and the aquifer. This is not the case where confined conditions exist because atmospheric pressure changes result in a pressure differential between water in the well and the water in the aquifer. Water moves from the well to aquifer when atmospheric pressure rises causing a temporary water-level decline and moves back into the well from the aquifer when atmospheric pressure decreases causing a temporary water-level rise. The barometric efficiency of the aquifer is defined as:

$$\text{Barometric Efficiency} = \gamma_w(dh/dP_a) \qquad \text{(Eqn. 4)}$$

Where γ_w = the specific weight of water and the ratio dh/dP_a is the change in hydraulic head (h) with respect to the change in atmospheric pressure (P_a) (Domenico an Schwartz, 1992). The barometric efficiency cannot exceed 1 by definition and thus the change in atmospheric pressure must be greater than the induced change in fluid pressure in the well.

To estimate the barometric efficiency of the aquifer and illustrate the correlation of water-level changes with changes in atmospheric pressure, six intervals were selected for analysis from the June 2002 – June 2003 hydrograph and the January – March 2002 hydrograph (Table 1; Figures 41, 42).

Table 1. Selected time intervals of the January-March 2002 and the June 2002-June 2003 hydrographs and barographs used to estimate barometric efficiency.

Time Period	Number of Data Points Included	Percent of the Water-Level Residual Variation Explained by Atmospheric Pressure Residual Variation (R^2)	Barometric Efficiency
7/5/2002 – 7/19/2002	360	0.60	0.66
5/21/2003 – 6/11/2003	515	0.67	0.59
5/10/2003 – 5/19/2003	239	0.54	0.89
1/11/2002 – 3/20/2002	1,632	0.74	0.54
8/26/2002 – 10/22/2002	1,391	0.50	1.47
12/8/2002 – 3/13/2003	2,302	0.14	0.67

The first 3 time intervals in Table 1 follow significant recharge events in 2002 and 2003 during which the water-level in the monitoring well was in recession and in the Ft. Riley Limestone interval (Figure 42). Within each selected interval, the water-level recession trend appears to be linear with time and can be approximated by a best-fit line from linear regression (Figures 43a-c). A best-fit line through the barometric pressure vs. time data was also computed to remove the trend from the data. The residuals from regression were computed as the difference between the observed data and the trend predicted estimate. The atmospheric pressure residuals were computed originally in units of inches of mercury relative to sea level and were converted to units of meters of water. Plots of time vs. the converted atmospheric pressure and depth to water residuals show that these variables are well correlated (Figures 44a-c). A best-fit line through a scatter plot of atmospheric pressure residual vs depth to water residual shows that more than

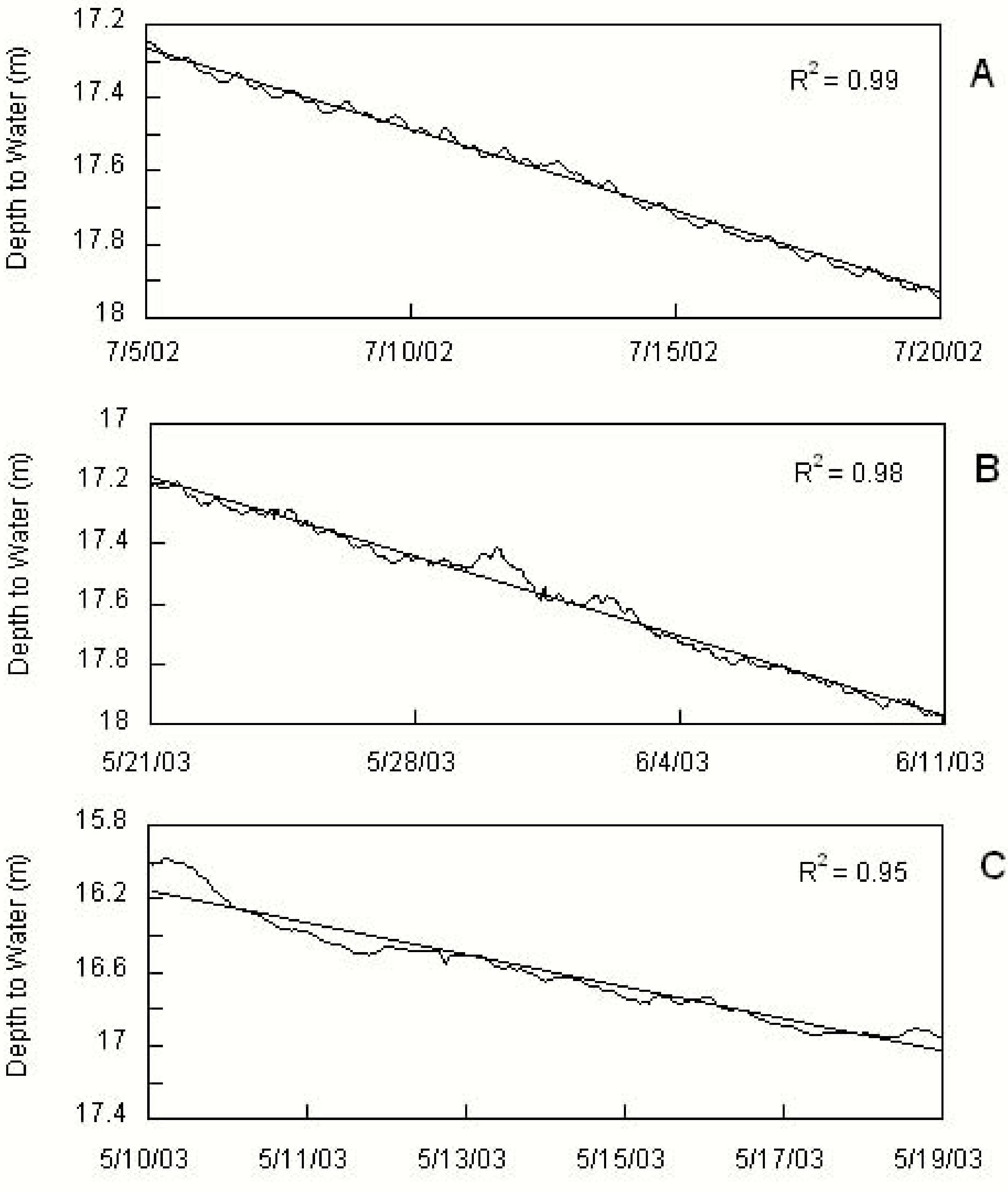


Figure 43. Trend lines from linear regression through the monitoring well hydrographs for 7/5-19/2002 (A), 5/21-6/11/2003 (B), and 5/10-19/2003 (C). All depth to water values are taken from the top of the well casing.

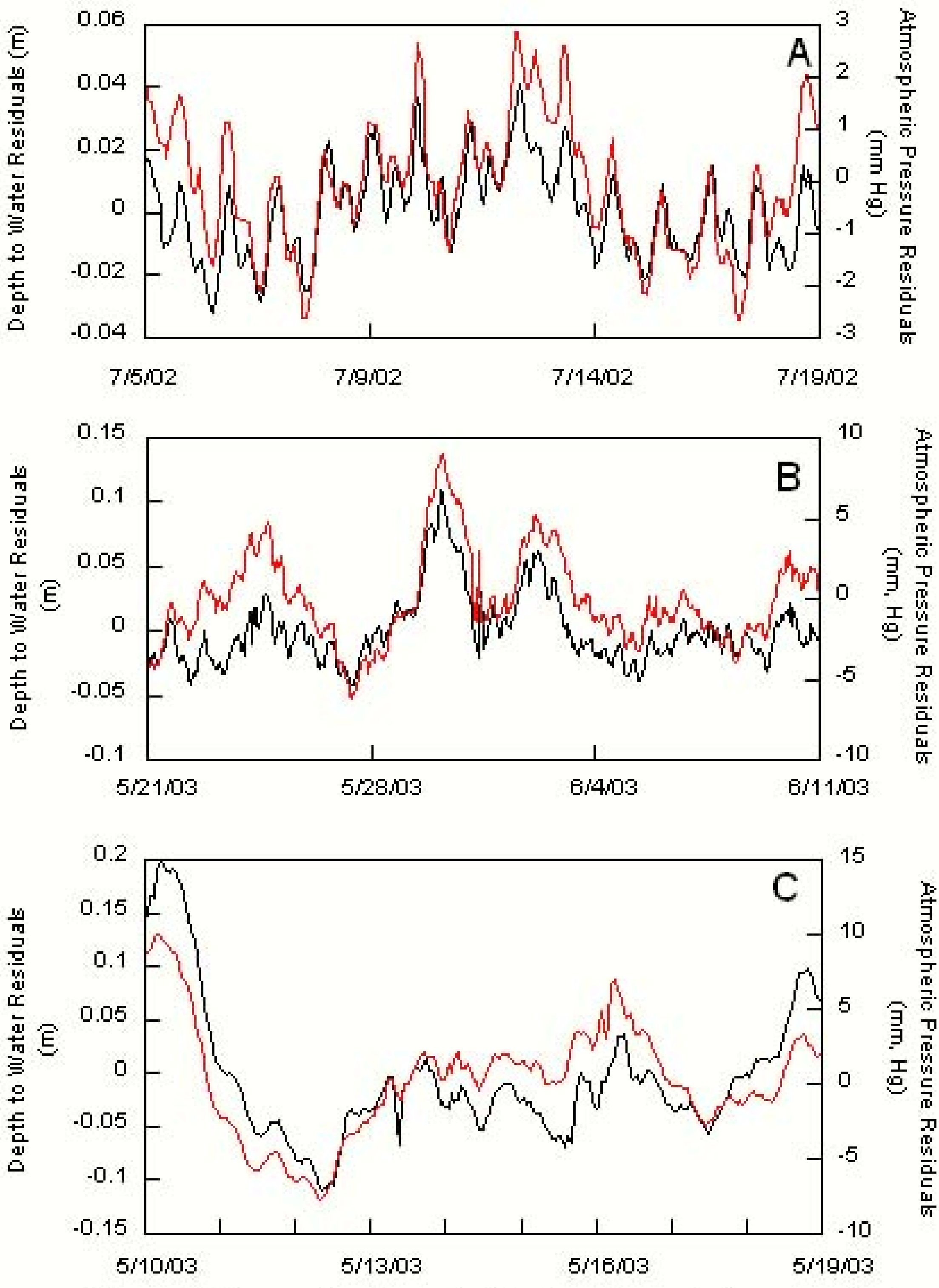


Figure 44. Time vs depth to water (black) and atmospheric pressure (red) residuals from linear regression for the periods 7/5-19/2002 (A), 5/21-6/11/2003 (B), and 5/10-19/2003 (C).

50% of the depth to water variation is explained by fluctuations in atmospheric pressure (Figures 45a-c). The slope of the best-fit line through the scatter plots is the barometric efficiency, which ranges from 0.59 to 0.89 for the 3 time periods.

In the last 3 time intervals in Table 1 water levels in the monitoring well were below the top of the Florence Limestone and were declining slightly with time (Figures 41, 42). The downward trends appeared to be linear with time and were approximated by a best-fit line from linear regression (Figure 46a-c). Plots of time vs. the transformed barometric pressure residuals and the depth to water residuals for each of the three time periods demonstrate a good correlation between water-level and atmospheric pressure change (Figures 47a-c). Overall, the time vs. atmospheric pressure and depth to water residuals plots show that water-level changes correlate very well with changes in barometric pressure for each of these time periods. The January – March 2002 time vs. residuals plot shows that the amplitude of the water-level change is less than the amplitude of the corresponding change in atmospheric pressure as predicted (Figure 47a). The best-fit line through a scatter plot of the residuals shows that most of the variation in the water-level residual can be explained by atmospheric pressure change (Figure 48a; Table 1). The barometric efficiency is slope of the best-fit line through the scatter plot and is calculated to be 0.54. The barometric efficiency calculated using the December 8, 2002 – March 13, 2003 data is in the same range as that calculated using the January – March 2002 data. However, the variation in depth to water explained by barometric pressure changes is much lower (14%) and thus the estimate has a low reliability even though it falls in the range of the other calculated values.

The time vs. residuals plot for the August 26 – October 22, 2002 period shows that the amplitude of the water-level change is significantly greater than the corresponding change in atmospheric pressure (Figures 47c). The best-fit lines through scatter plots of the residuals show that the R^2 value is somewhat less for both sets of residuals than the R^2 values for the January – March 2002 residuals. Visually, it is obvious that there is more scatter in these plots than in the plot for the January – March 2002 period. The slopes of the best-fit lines or the barometric efficiency are both greater than 1 (Figures 48b-c; Table 1).

No apparent reason can be discerned from the data to explain the change in the response of the transducer to atmospheric pressure changes. MiniTroll equipment malfunction does not appear to have contributed to the overestimation of the water level. Monthly checks were made during the monitoring period to insure that the water-level data collected using the MiniTroll reflected actual water-level changes by comparing the recorded data with measurements made using an electric tape (Figure 33). As part of the analysis for this report, a “dry run test” was conducted to determine if the pressure transducer vent tube might have become plugged during monitoring. The results showed only small random changes in pressure with a mean “out-of-water” pressure value of -0.081 psi and a 0.005 psi range of variation over a 91 hr (3.8 day) period (Figure 49). Comparison of the atmospheric pressure residuals with the pressure residuals demonstrates that is no relationship between change in atmospheric pressure and changes in the “out-of-water” pressure readings from the MiniTroll (Figure 50).

The atmospheric pressure residuals were also scrutinized to determine if barometer sensitivity at the Midcontinent Airport at Wichita had changed between the end of the January-March 2002

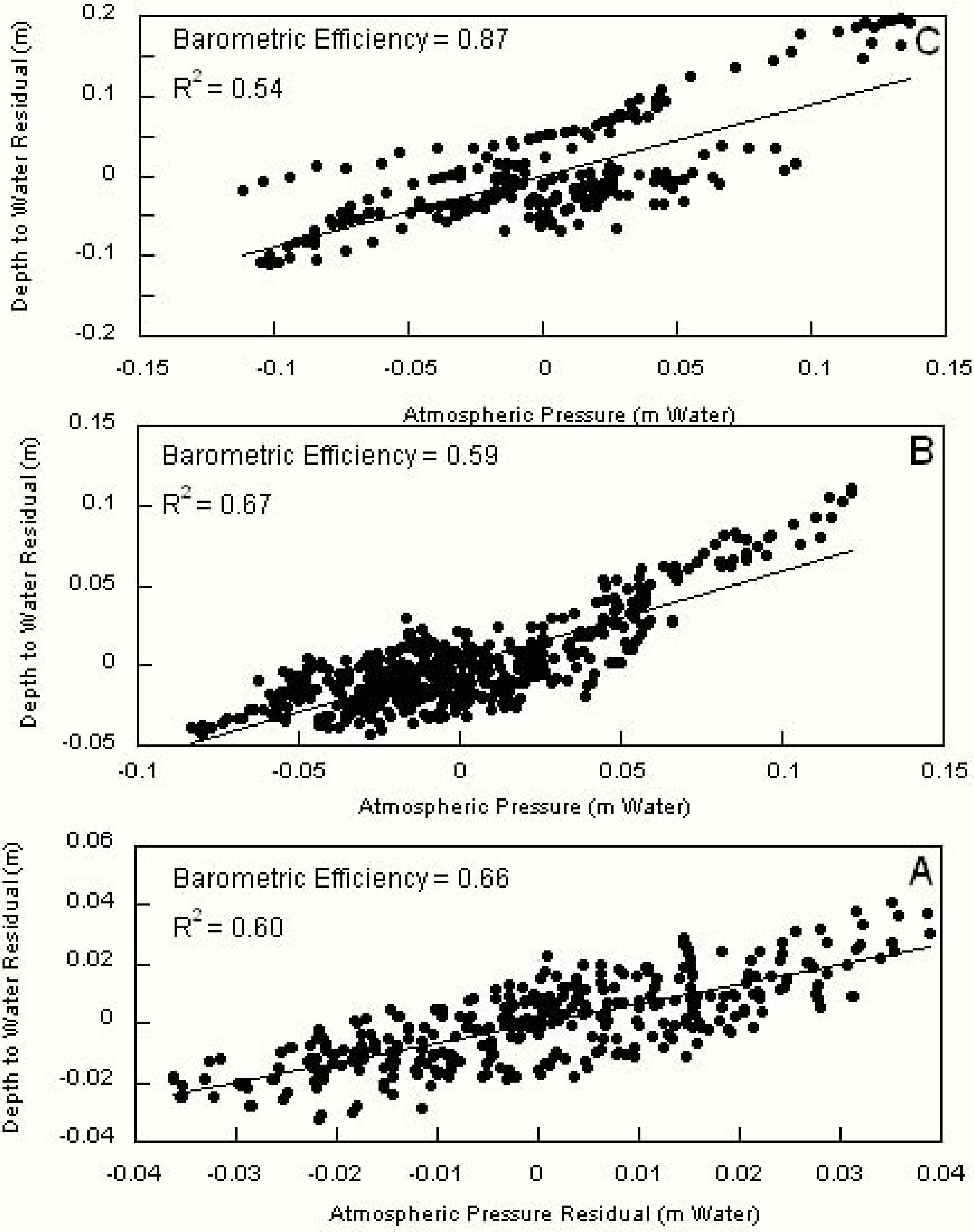


Figure 45. Scatterplots of the atmospheric pressure residuals in equivalent meters of water vs the depth to water residuals with a best-fit line through the points from linear regression for the periods 7/5 - 19/2002 (A), 5/21 - 6/11/2003 (B), and 5/10 - 19/2003 (C).

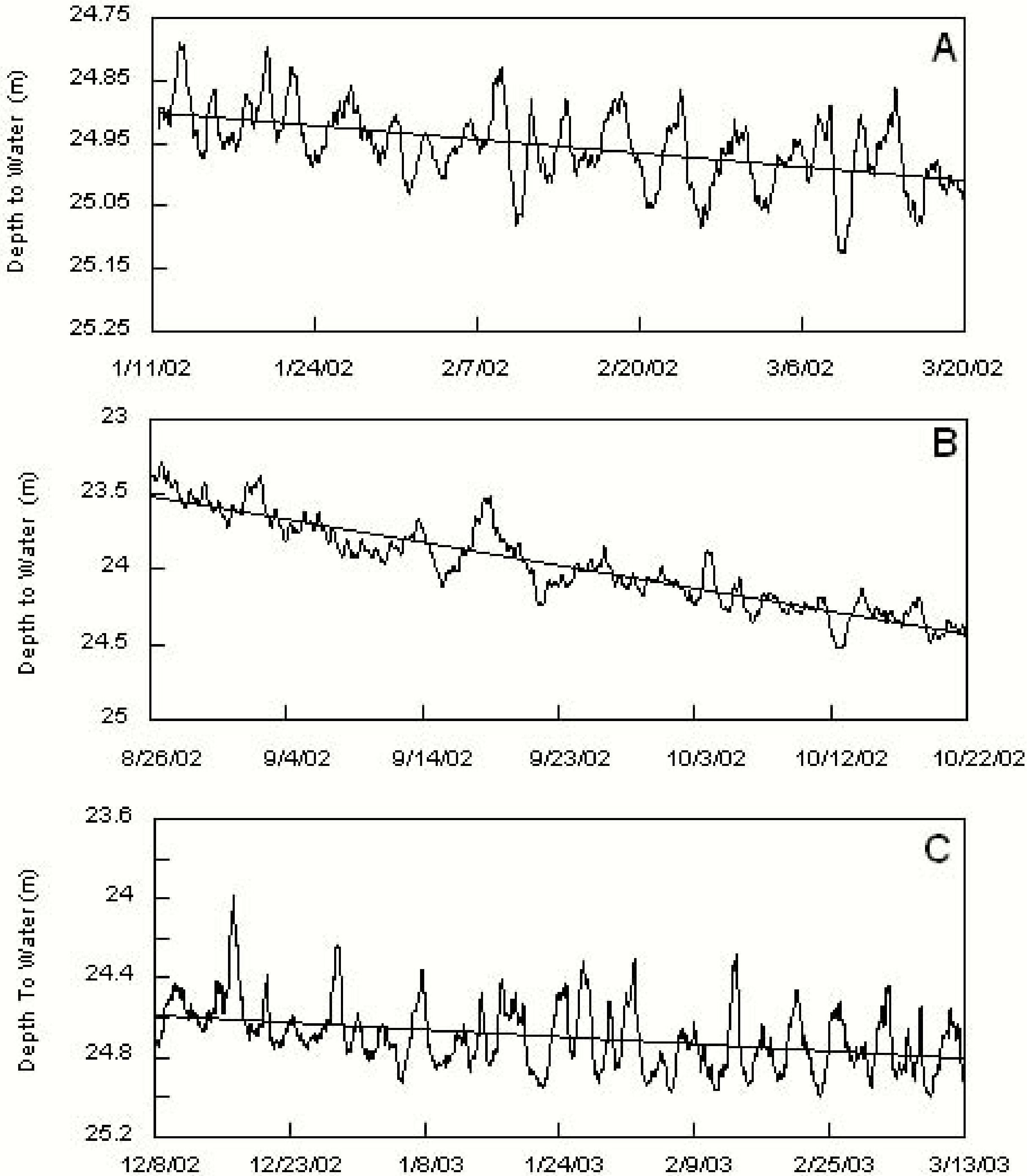


Figure 46. Time vs depth to water with trend lines from linear regression for the periods 1/11 - 3/20/2002 (A), 8/26 - 10/22/2003 (B), and 12/8/2002 - 3/13/2003 (C).

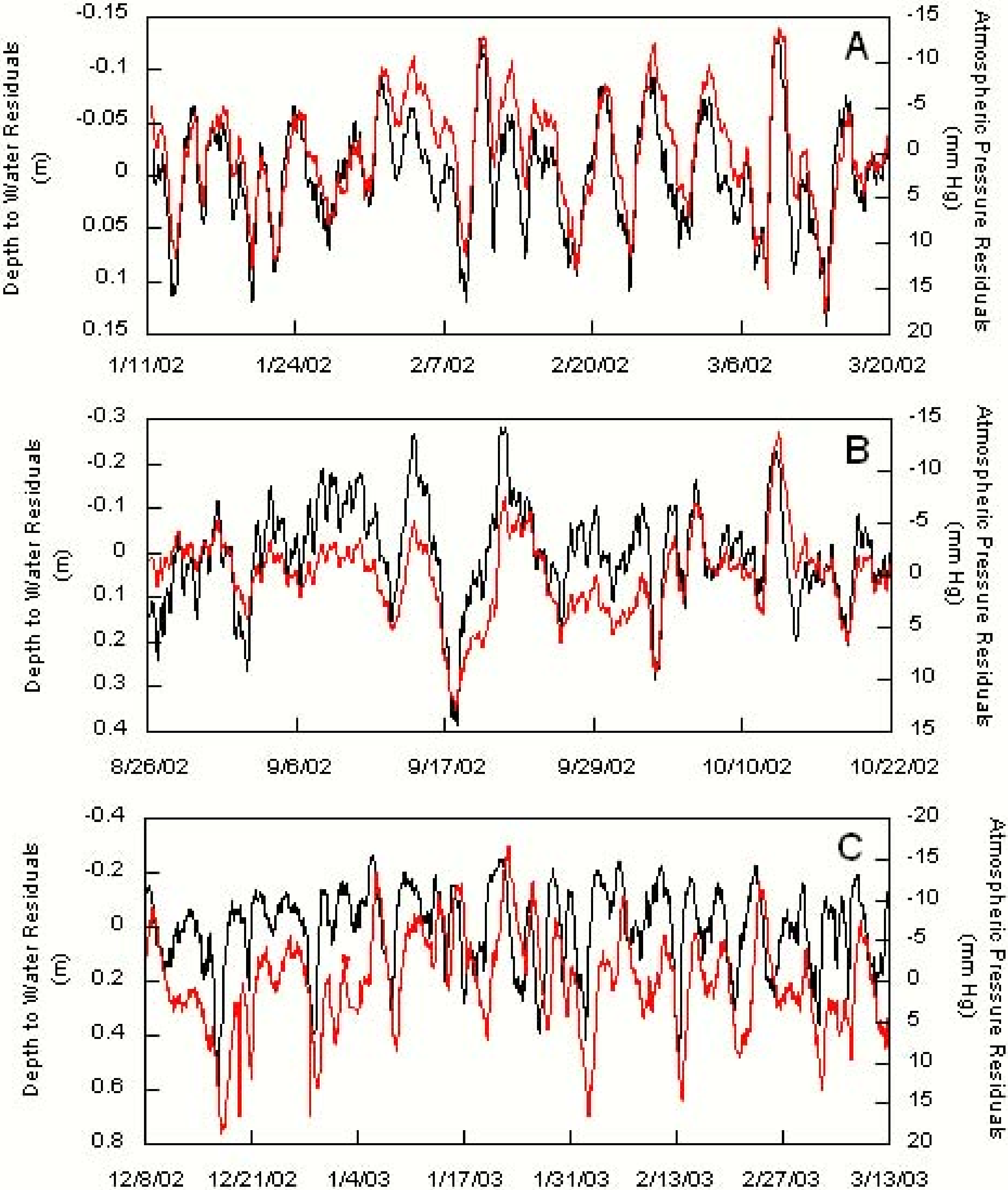


Figure 47. Time vs depth to water (black) and atmospheric pressure residuals (red) from linear regression for the periods 1/11 - 3/20/2002 (A), 8/26 - 10/22/2002 (B), and 12/8/2002 - 3/13/2003 (C).

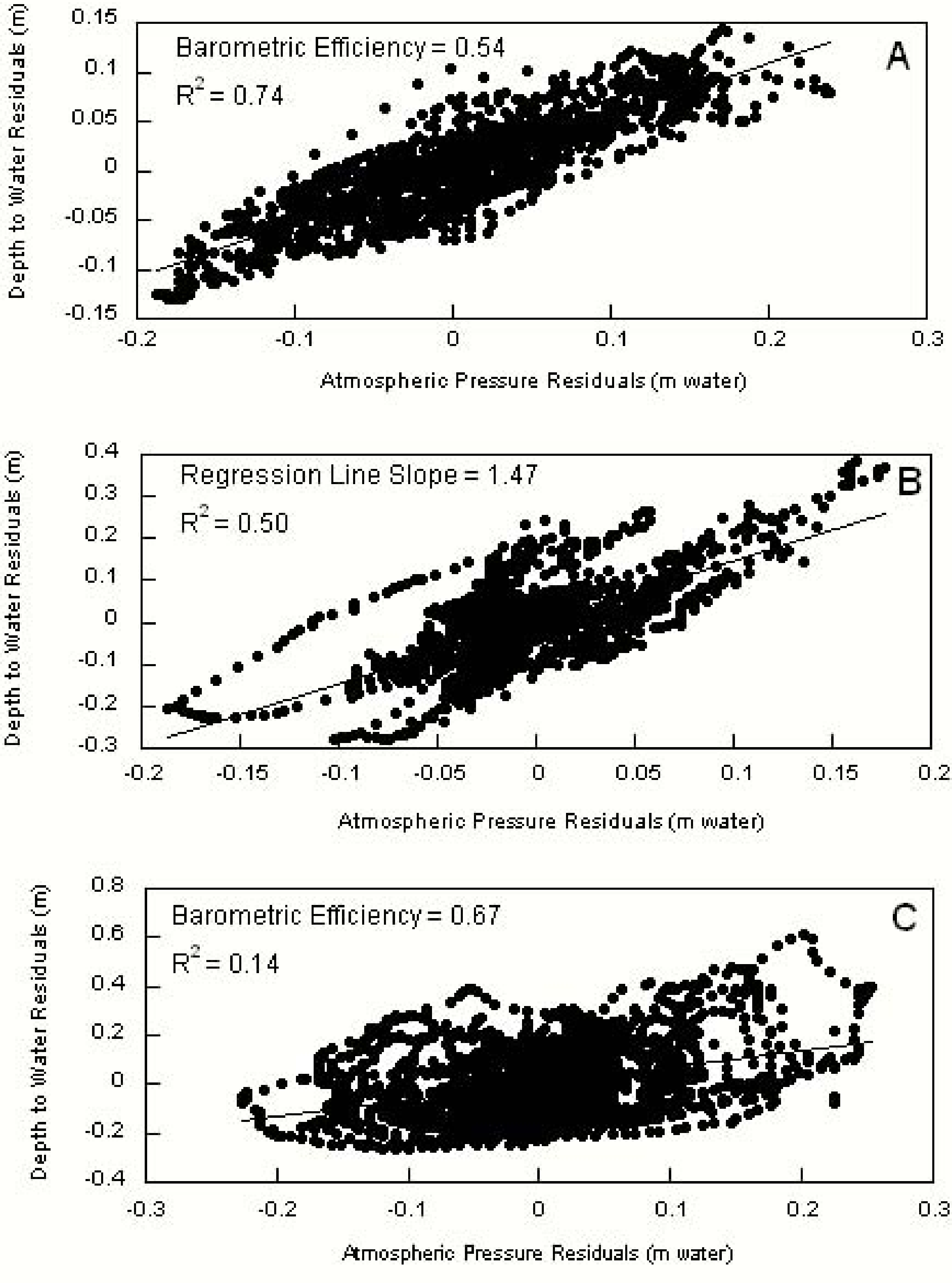


Figure 48. Scatterplots of the depth to water residuals vs atmospheric pressure residuals in m of water with a best-fit line from linear regression for the periods 1/11 - 3/20/2002 (A), 8/26 - 10/22/2002 (B), and 12/8/2002 - 3/13/2003 (C).

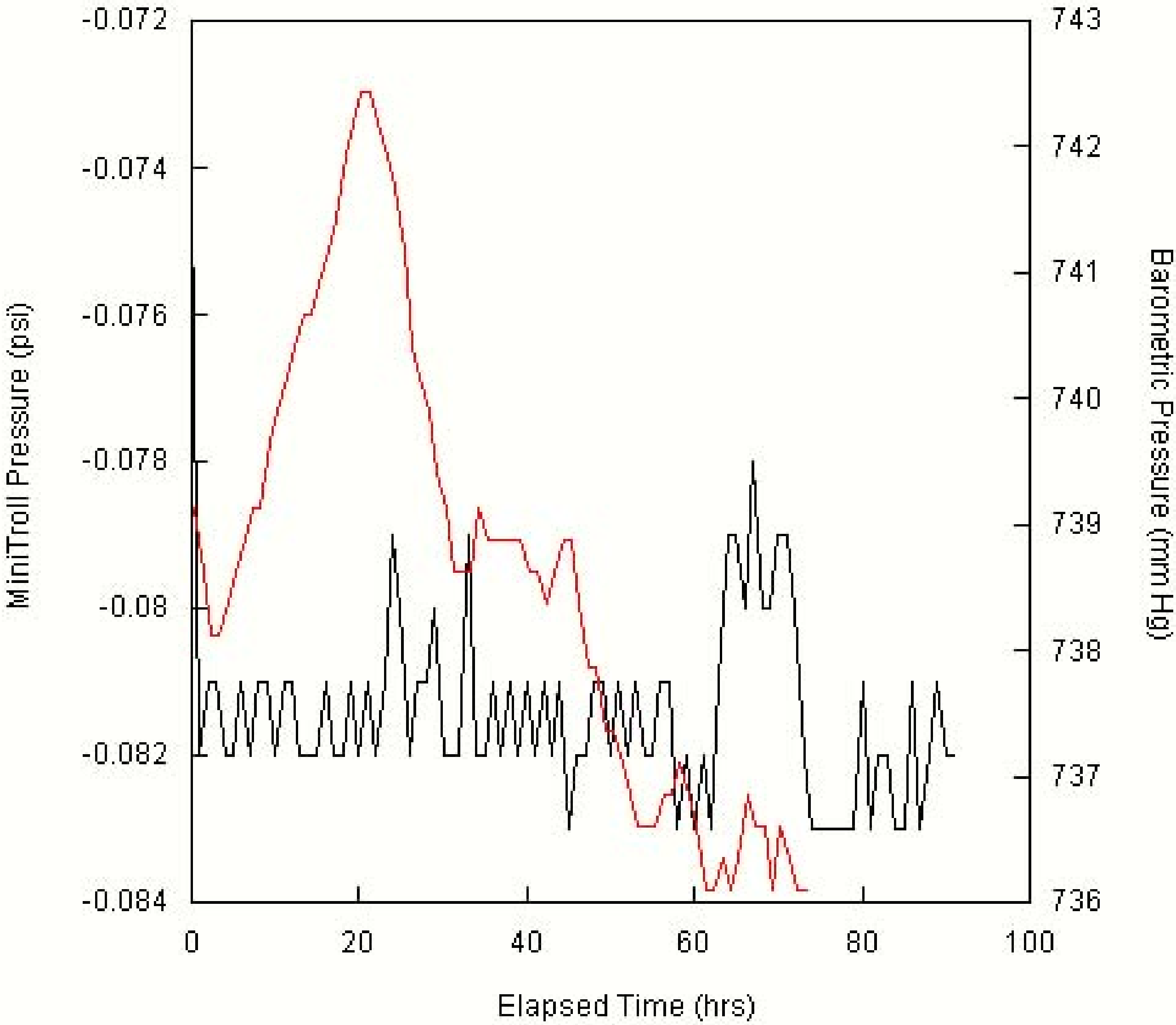


Figure 49. Elapsed time vs MiniTroll pressure (black) and atmospheric pressure (red) measured at the Municipal Airport, Lawrence, Kansas.

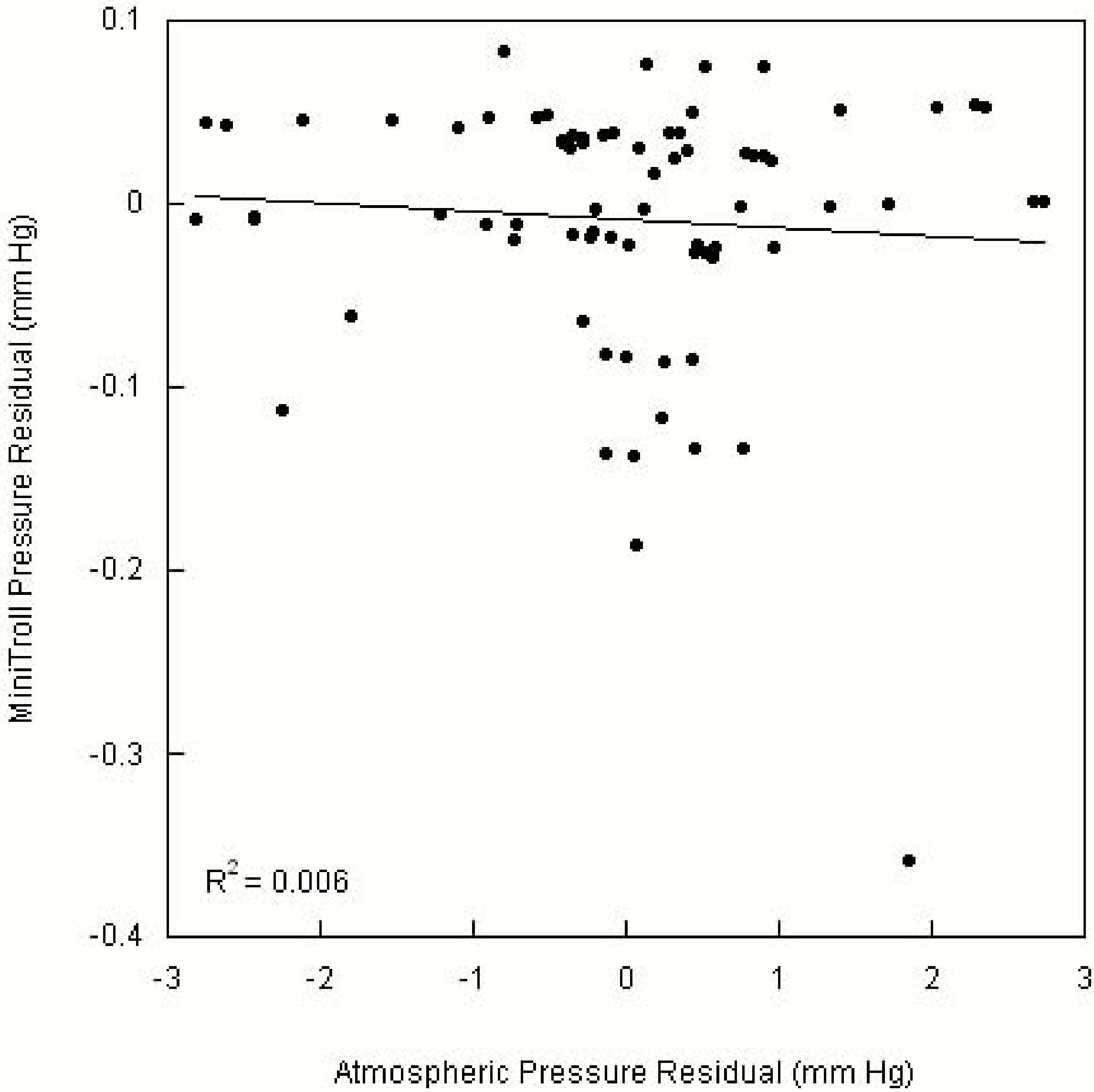


Figure 50. Atmospheric pressure residuals vs MiniTroll pressure residuals in mm Hg for the "out-of-water" test with a best-fit line from linear regression.

and the beginning of the June 2002-June 2003 monitoring periods. The range of the atmospheric pressure residuals expressed in equivalent hydraulic head change was 0.46 m, 0.36 m, and 0.48 m (1.40 ft, 1.19 ft, and 1.58 ft) for the January – March 2002, the August 26 – October 22, 2002, and the December 8, 2002 – March 13, 2003 periods, respectively. The mean of the residuals for all data sets was 0 and the standard deviation of the equivalent atmospheric pressure residuals was 7.92 cm, 5.18 cm, and 8.23 cm (0.26 ft, 0.17 ft, and 0.27 ft) for the January – March 2002, the August 26 – October 22, 2002, and the December 8, 2002 – March 13, 2003 periods, respectively. A t-test of the residuals reveals no statistical difference between the three populations of residuals, which suggests that the distribution of atmospheric pressure fluctuations did not change statistically between the two monitoring periods.

11.0 Barneston Aquifer Water Quality

11.1 Application to This Investigation

Hydrogeochemical data can provide information on sources of recharge or the ground-water flow paths taken through the karst aquifer system to springs. Several examples follow. In a classic study, Shuster and White (1971) classified 14 springs in the central Appalachians into diffuse-flow feeder-system types (centimeter-scale openings) or conduit-flow feeder systems types (centimeter to meter or larger-scale). They found higher variability in hardness of the discharge from conduit-flow feeder system types over time than in the discharge from diffuse-flow feeder system type springs. More recently, the data analysis has focused on development of geochemical mixing models to quantify sources of recharge to spring discharge. Katz et al. (1998) used the chemical data from well samples and a binary mixing model to quantify recharge to the Floridan aquifer from high flow events in Little River, an ephemeral stream. Martin and Dean (2001) and Swenton et al. (2002) used selected solutes and physical parameters to investigate the timing and relative magnitude of the exchange of water between the matrix and conduits as a result of high flow events in the Santa Fe Sink/Rise system of the Floridan aquifer. Lee and Krothe (2001) developed a geochemical mixing model to partition spring discharge from precipitation, soil water, epikarstic water, and phreatic diffuse flow sources in a catchment in south central Indiana.

The analysis of turbidity data from springs has followed a similar evolutionary pathway. Traditionally, turbidity data are correlated with precipitation data to show the rapid response of the karst system to rainfall events in the catchment. Very recently, turbidity data were used to evaluate the transport and remobilization of trapped fine sediments to a spring in northwestern France using turbidigraph separation (Massei et al., 2003).

In this investigation, hydrogeochemical information was used to provide information on the sources of water contributing to the discharge in Crystal spring. Early in the investigation, attention was drawn to the sinkholes as a primary source of water to the spring. The dye traces that were conducted later in the study confirmed the rapid response of the spring to the addition of water into the karst system through these entry points. Martin Creek is an intermittent stream for most of the year and thus the contributions to spring flow from this source were expected to be substantial, but episodic. It was also expected that other sources could contribute water to the spring. These include water that infiltrated through the soil zone and through the exposed jointed limestones in the Barneston Limestone outcrop belt and ground water from flow systems larger than the system that encompasses the spring catchment (intermediate-scale flow systems). The hydrogeochemical and spring discharge data from previous surveys of the spring were too few to make an initial assessment of these potential contributors to spring flow.

Hydrogeochemical data were also developed in this study to provide a better-defined water quality baseline than had been attempted previously. In the KGS study of the hydrogeology of Marion County (O'Connor, 1982, 1984; O'Connor and Chaffee, 1983, 1984; Chaffee, 1984, 1988; and Chaffee and O'Connor, 1986) and in the later Flint Hills spring survey (Sawin et al., 1999), Crystal spring was sampled only once and an estimate of the discharge was made. In both studies, water samples were collected when the discharge exceeded 189 L/s. Based on the discharge measurements taken during this study, this rate of discharge is only exceeded about 20% of the time (Figure 30). The high variability in spring discharge in response to rainfall and periods of flow in Martin Creek suggests that Crystal spring water quality could be different when these sources are not contributing to discharge. This study helps to define a baseline envelope in recognition that spring water quality depends on discharge.

11.2 Sample Sites

A total of 55 water samples were collected for chemical analysis from surface water in Martin Creek and in an unnamed drainage adjacent to Crystal spring, wells and springs in the study area that were believed to tap supplies in the Barneston aquifer (Figure 51). Wells were selected for sampling based on interpretation of the driller's log included with the WWC-5 record of the well on file at the KGS. Only those wells that are screened across the Barneston Limestone were selected for sampling.

Initially, water samples were collected during August 2000 as part of the study area reconnaissance. Sampling locations included Crystal spring, two sites on Martin Creek, one site on the unnamed drainage adjacent to the spring, and the upper Hett spring that discharges from the Florence Limestone into Bruno Creek in SE, NE, NE, SW Sec. 10. T. 20 S., R. 5 E. Later in 2000, samples were occasionally collected from Crystal spring as the opportunity arose. Winter and late spring sampling of surface water in Martin Creek and wells and springs was undertaken in 2002. Ground-water samples from 8 wells were collected in the winter survey and from 9 wells in the spring survey. Seven wells that were sampled in the winter survey were revisited in the spring. Samples of the discharge from the upper and lower Hett springs were also collected both times. These surveys were taken to determine areal and seasonal differences (wet vs. dry) in water chemistry. Monthly sampling and discharge measurements at Crystal spring were undertaken beginning in 2002 and continued to the end of the study in June 2003 to assess the variability in spring water quality and discharge over the 18-month period of sampling.

11.3 Sampling Procedures

At each sampling site care was taken to insure that the samples were collected as close to the exit point from the aquifer as possible. In the case of domestic wells, samples were taken downstream of the pressure tank. The well was allowed to pump while the temperature and specific conductance of the produced water were being monitored downstream of the pressure tank. When it appeared that recently discharged water from the well had replaced all of the water in the tank and the lines to the sampling point, a water sample was collected. If there was a water softener in use, the well was not sampled. In the case of springs, samples were collected where the main flow of water discharged to the surface. Stream samples were collected near the middle of the stream or where the stream was flowing most vigorously.

Prior to sample collection both the water temperature and the specific conductance (in units of μS/cm) were measured. Unfiltered, unacidified samples were collected in 500-ml polyethylene

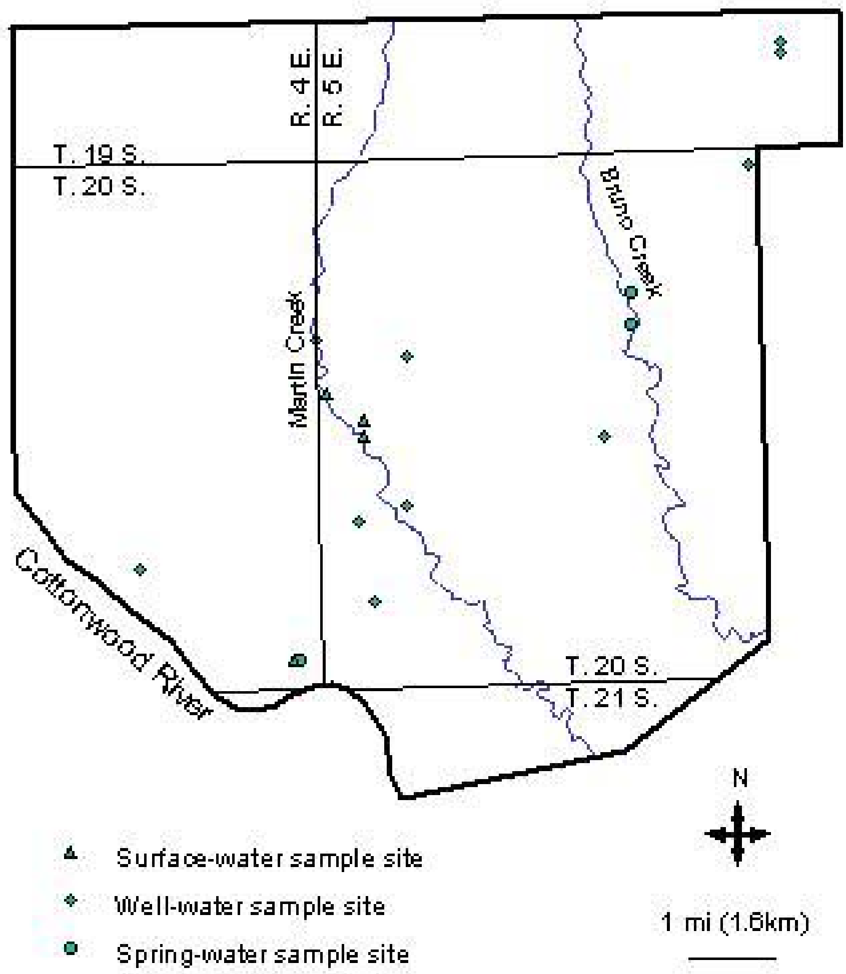


Figure 51. Distribution of spring, surface-water and well sample sites in the project area.

bottles for inorganic analyses and if appropriate, acidified samples were collected in 250-ml polyethylene bottles for other inorganic analyses. One-liter glass bottles were used to collect samples for the DOC analyses. Care was taken to eliminate the possibility of contamination from residuals left in the sample bottle by rinsing the bottle and bottle cap in the stream of water flowing from the well or spring, where possible, prior to collecting the sample. All bottles were filled to the top to minimize head space. Sample bottles were labeled with the date and an assigned sample number. In a field book, the sample numbers were keyed to the particular well or spring being sampled along with the date and time of sample collection, the location, and the temperature and specific conductance of the sample. The collected samples were placed in an ice chest to keep them cool. Upon arrival at the KGS the field data were entered into the Analytical Services sample log and the samples were stored in a refrigerator in the Analytical Services Section until they were analyzed.

11.4 Chemical Parameters and Analysis

Water samples from the 2000 reconnaissance were analyzed only for specific conductance, and chloride (Cl), sulfate (SO_4) and nitrate (NO_3) concentrations. Most of the monthly samples were also analyzed for these dissolved constituents, but some of the samples collected early in 2002 were analyzed only for specific conductance, silica (SiO_2), calcium (Ca), magnesium (Mg), sodium (Na), potassium (K), strontium (Sr), bicarbonate (HCO_3), fluoride (F), and boron (B) concentrations. Water samples collected during the winter and spring 2002 surveys were analyzed for specific conductance, silica, Ca, Mg, Na, K, HCO_3, Cl, SO_4, NO_3, F, B, and dissolved organic carbon (DOC).

Specific conductance, and Cl and SO_4 were selected for analysis because they are indicators of local and larger-scale ground-water flow (O'Connor, 1984). Twiss (1991) reported finding gypsum and halite in core samples of the Barneston Limestone collected downdip of the outcrop belt. Specific conductance is an indicator of the total dissolved solute load in water (Hem, 1985). It was expected that specific conductance and the Cl and SO_4 would be lower in ground water coming from surface water input through the sinkholes in Martin Creek because of its lower dissolved solids concentration. Nitrate was selected as an indicator of contamination from human activities. Nitrate concentrations in ground water from the larger-scale flow system may be lower than in ground water from shallow sources. DOC was believed to be a good indicator of surface-water inflow through the sinkholes in Martin Creek to the conduit system and eventually to Crystal spring. Miller et al. (1990) reported concentrations of 0.5 and 0.8 mg/L DOC in well samples from the Nolans and the Wellington formations in Marion County.

The Analytical Services Section, KGS, performed inorganic analyses of the water samples and Dr. Steve Randtke, Department of Civil and Environmental Engineering, University of Kansas, performed the DOC analyses. Dr. Stephen A. Macko Professor in the Department of Environmental Sciences at the University of Virginia, Charlottesville, VA, performed nitrogen-15 isotope analyses on water samples.

11.5 Analytical Methods, Quality Control, and Quality Assurance

The Analytical Services laboratory at KGS maintains a system of quality control and quality assurance that includes periodic analysis of US Geological Survey reference waters. Results from this program indicate low error in KGS analyses.

Prior to the chemical analysis, the water samples were filtered through 0.45-µm membrane filters to insure removal of most of the suspended solids. The analyses therefore represent dissolved constituent concentrations. Measurements of sample specific conductance were taken in the laboratory using a conductivity meter as a check on the specific conductance measured in the field at the time of sample collection. The samples were acidified prior to NO_3 analysis.

An automated, colorimetric, ferricyanide method on an Alpkem Flow Solution IV System was used for the chloride determination. The method is given in Standard Methods and U.S. Geological Survey and U.S. Environmental Protection Agency publications. The method range used will be 0-100 mg/L. Samples with constituent concentrations greater than the analytical method range were diluted to give concentrations within the range. The detection limit for the method as operated for this range at the Kansas Geological Survey is 0.3 mg/L. The actual detection limit for each diluted sample as analyzed is less than 1% of the concentration. Standards were analyzed before, during, and after the water samples. Internal blanks were distilled, deionized water inserted as samples at intervals of ten samples and before and after standards during the determination and used to verify or adjust the baseline. The estimated maximum error in the determinations is 3%.

An automated, colorimetric, complexometric methylthymol blue method on a Technicon AutoAnalyzer was used for SO_4 determinations. The methods are given in Standard Methods and U.S. Geological Survey and U.S. Environmental Protection Agency publications. The method range used was 0-100 mg/L. The detection limit is 2 mg/L for the methylthymol blue method as run for the ground-water samples. Standards were analyzed before, during, and after the water samples in the methylthymol blue method and before the samples in the turbidimetric method. Internal blanks were distilled, deionized water inserted as samples at intervals of ten samples and before and after standards and used to verify or correct the baseline in the methylthymol blue method, and inserted before the samples in the turbidimetric method. The estimated maximum error in the determinations is 4%.

Nitrate was determined using an ultraviolet-spectrophotometric screening method adapted from Standard Methods using the Technicon AutoAnalyzer II System. The method range used was 0-88.55 mg/L as NO_3. The detection limit for the ground-water samples is on the order of 0.1 mg/L. Standards were analyzed before and after the water samples have been analyzed. Internal blanks were distilled, deionized water inserted as samples and used to verify or correct the baseline in the method. The estimated maximum error in the determinations is less than 5%.

Water samples for nitrogen-15 isotope analyses were collected in unacidified, 100-ml, polyethylene bottles. The bottles were kept chilled at 4°C on ice until return to the laboratory, where the samples were then frozen until they were sent to the University of Virginia for analysis. If a sample was turbid, it was kept chilled until it could be filtered in the KGS analytical laboratory. A 100-ml aliquot was obtained from the filtered sample and kept frozen until it was sent for analysis.

For total nitrogen, the samples were acidified to remove carbonate and dried. The carbonate free residues were weighed into tin capsules and converted to N_2 for isotope analysis using a Carlo Erba elemental analyzer (EA), which is coupled to an OPTIMA stable isotope ratio mass spectrometer (GVI, Manchester, UK). Nitrogen isotopes were determined with a single combustion using a dual furnace system composed of an oxidation furnace at 1020ºC and a reduction furnace at 650ºC. The resulting gases were chemically dried and directly injected into the source of the mass spectrometer. The stable isotopic ratio is reported as follows:

$$\delta^{N}E = [R_{sample}/R_{standard} - 1]\ 10^{3}\ (‰) \qquad \text{Eqn. 5}$$

where N is the heavy isotope of the element E and R is the abundance ratio of the heavy to light isotopes ($^{15}N/^{14}N$) of the element. The standard for nitrogen is atmospheric N_2 (air), which is assigned and $\delta^{N}E$ value of 0.0‰. The reproducibility of the measurement is typically better than ±0.2‰ using the continuous flow interface on the OPTIMA. In the laboratory, the samples are commonly measured against tanks of nitrogen gas, which have been calibrated against atmospheric N_2.

Dissolved organic carbon concentrations were determined using a Dohrmann DC-80 analyzer according to the persulfate-ultraviolet oxidation method described in Standard Methods for the Examination of Water and Wastewater (1985). Samples were acidified to pH<2, purged with nitrogen to remove carbon dioxide, and shaken prior to injection into the analyzer. The system blank was subtracted from of the measured values of DOC. Analytical precision was ± 2% for DOC concentrations >1.0 mg/L.

Turbidity was measured in by city of Florence water treatment personnel using a portable turbidimeter from the Hach Chemical company.

11.6 Data Reporting and Analysis

Both the data water chemistry data collected in the field and the laboratory determinations are included in this report in Appendices 1 and 2. Specific conductance and temperature were reported as µS/cm to the nearest whole integer value and °C to the nearest 0.1°C, respectively. Constituent DOC concentrations were reported as mg/L for each sample. Nitrogen-15 values were reported relative to the standard as per mil (‰). Turbidity data are reported in NTU units with an accuracy of ±2% and a repeatability of ±1%.

Chemical types of water are classed according to the three major cations and the three major anions that usually comprise most of the naturally occurring concentrations of ions in solution. The three major cations are Ca, Mg, and Na and the three major anions are HCO_3, SO_4, and Cl. In this project, the concentration of K was grouped with the Na and the concentration of Sr was grouped with the Ca for classification purposes. The concentrations of the major and selected minor constituents were converted to milliequivalents to determine the dominant and subdominant cations and anions in solution. Calcium and Mg water types are included together in one class rather than forming separate classes because calcite and dolomite equilibria are the dominant and similar controls on the concentrations of Ca and Mg and are related in that they are the principle control on the HCO_3 concentration. If additional cations and anions have appreciable concentrations in addition to those with the highest contents, then secondary types within a main type can be defined.

To classify each water sample according to its type, the concentrations of Ca, Mg, Sr, Na, K, HCO_3, SO_4, and Cl were first converted to milliequivalents/L. The milliequivalent/L values were summed for the anions and cations and the relative proportion of each contributing constituent was calculated. The first cation and anion in a water-type designation is that which contributes the most to the total milliequivalents/L of the positive or negative charges in solution, respectively. Designation of secondary types uses the following rules. If the major cation or anion comprises greater than 50% of the total concentration of cations or anions, respectively, the secondary ion must constitute more than 33% of the total millequivalents/L. If the major ion was less than 50% of the total, the secondary ion must constitute more than 30%.

The total dissolved solids concentration for each complete analysis was determined by summing the concentrations of the dissolved constituents, except in the case of the HCO_3. The HCO_3 contribution to the dissolved solids was taken into account by converting the measured concentration to an equivalent carbonate concentration before adding it to the sum (Hem, 1985). This was done by multiplying the bicarbonate concentration in mg/L by 0.4917. Figure 52 shows the relationship between the measured specific conductance in the laboratory and the calculated total dissolved solids concentration. The extremely high coefficient of variation between the variables indicates that the specific conductance is a good estimator of the total dissolved solids in the water samples.

11.7 Crystal Spring Water Quality

Winter and Spring 2002 Crystal spring samples are both Ca-HCO_3 type waters. TDS concentrations decreased from 410 mg/L in the winter sample to 311 mg/L in the spring sample and hardness from 345 mg/L to 270 mg/L. DOC increased from 0.46 mg/L in the winter sample to 2.49 mg/L in the spring sample. It is worth noting that the Winter 2000 sample was collected when there was little, if any flow in Martin Creek, but the Spring 2002 sample was collected when the stream was nearly bank full. The DOC concentration of the Spring 2002 Martin Creek water sample was 11.8 mg/L.

During the course of the study, the specific conductance of water samples from Crystal spring ranged from 450 mS up to 689 mS. Decreases in specific conductance coincided wetter periods and increases with the drier periods (Figure 53). Fluctuations in specific conductance are caused by changes in constituent concentrations, which are directly related to the transitions between

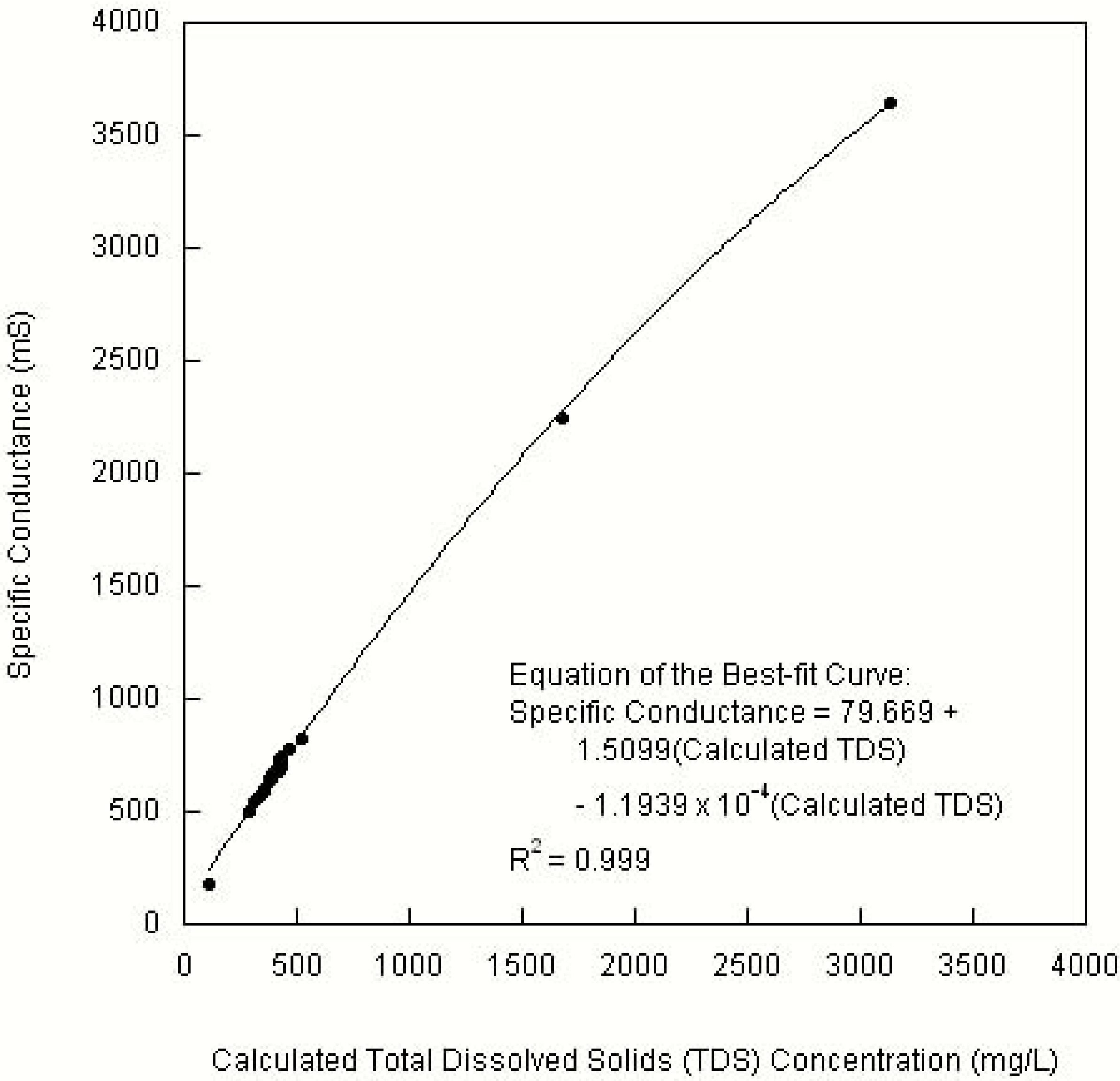


Figure 52. Calculated total dissolved solids concentration vs laboratory measured specific conductance with a best-fit curve from polynomial regression.

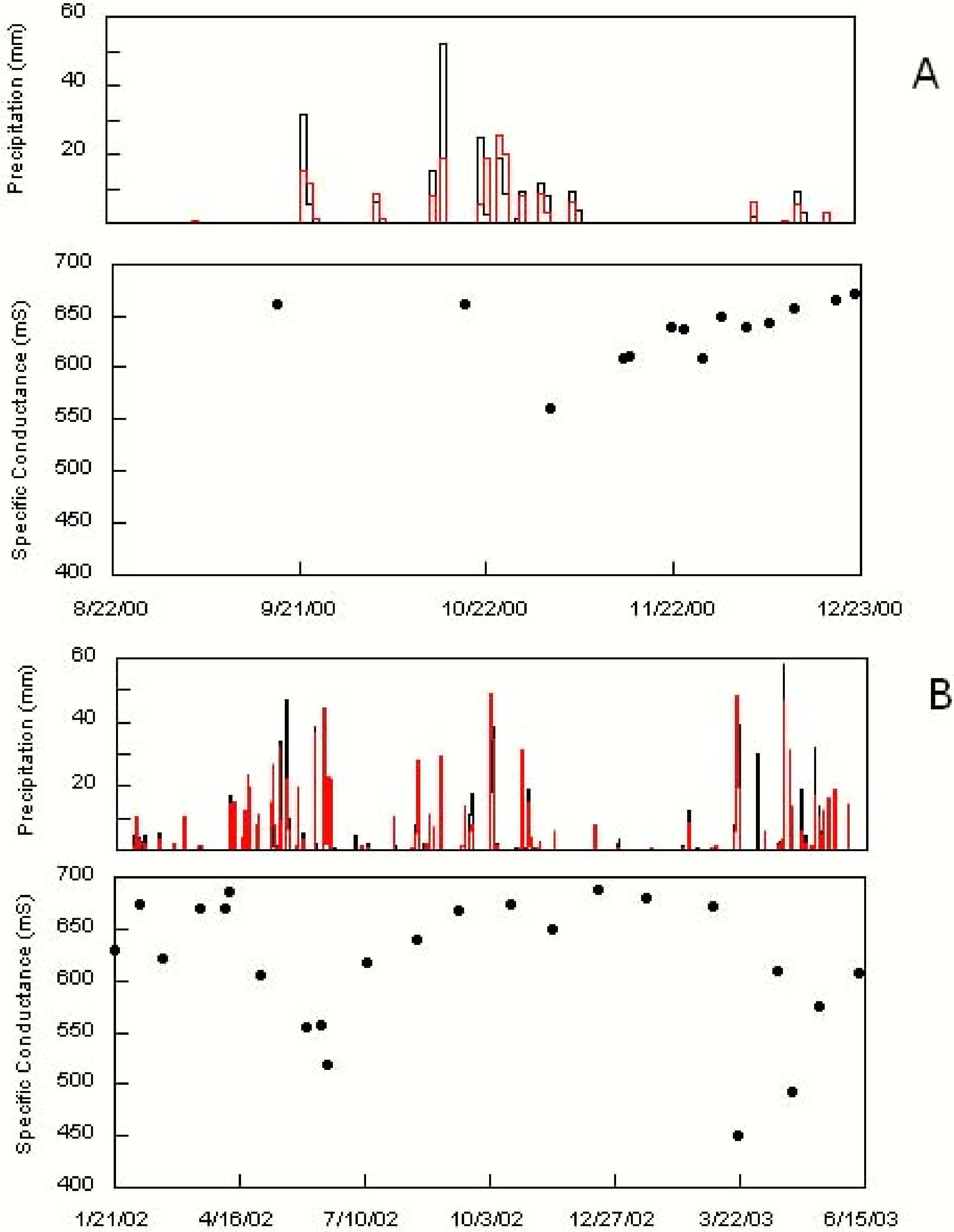


Figure 53. Time vs specific conductance in Crystal spring discharge and precipitation in 2000 (A) and in 2002-3 (B).

wetter and drier periods in the catchment. Concentrations of the major constituents, Ca, Mg, Na, Cl, and SO_4, and the minor constituent, Sr, decreased during wet periods from their higher, dry season levels, whereas the concentrations of the minor constituents, K, B, and NO_3, seemed unaffected by the transition between wet and dry periods (Figures 54-56). Table 2 shows the concentration ranges of Ca, Mg, Na, Cl, SO_4, K, Sr, B, and NO_3 in the samples collected during the study.

Table 2. Range of concentrations of selected chemical constituents in water samples from Crystal spring.

Constituent	Concentration Range	Median Concentration	Number of Samples	Sampling Period
Ca	77.9-90.4 mg/L	85.3 mg/L	7	1/4/2002-6/10/2002
Mg	29.2-18.3 mg/L	28.8 mg/L	7	1/4/2002-6/10/2002
Na	7.9-15.8 mg/L	14.3 mg/L	7	1/4/2002-6/10/2002
Cl	3.7-8.6 mg/L	6.9 mg/L	16	8/14/2000-12/12/2000; 1/4/2002-6/11/2003
SO_4	5.1-42.4 mg/L	31.55 mg/L	16	8/14/2000-12/12/2000; 1/4/2002-6/11/2003
K	0.8-1.9 mg/L	1.1 mg/L	7	1/4/2002-6/10/2002
Sr	1.26-5.9 mg/L	5.58 mg/L	7	1/4/2002-6/10/2002
B	<26-71 ppb	48 ppb	7	1/4/2002-6/10/2002
NO_3	2-9 mg/L	3.6 mg/L	16	8/14/2000-12/12/2000; 1/4/2002-6/11/2003

11.7.1 In Relation to Spring Discharge

Plots of Cl, SO_4, Sr, hardness, and the SO_4/Cl ratio vs. spring discharge show that these constituents and paramaters are all inversely related to spring discharge (Figures 57-59). Log-linear regression analysis of the data shows that the coefficient of variation associated with the best-fit line through the data ranges from 0.72 to 0.92. This indicates that majority of the variation in these chemical parameter values is explained by variation in spring discharge. In contrast, the fluctuations in the concentrations of NO_3, B, and K do not appear to be related to variations in spring discharge (Figures 60-62).

11.7.2 Turbidity

The turbidity of the water is a measure of its suspended solids content. A sample of turbid water was collected from the spring in April 2000 and analyzed by the KGS Analytical Services Section to determine the mineralogic composition of its suspended solids using X-ray diffraction. The results showed that the inorganic portion of the suspended solids in the sample consisted mostly of quartz with minor amounts of feldspar (anorthite and microcline), illite/mica, and dolomite.

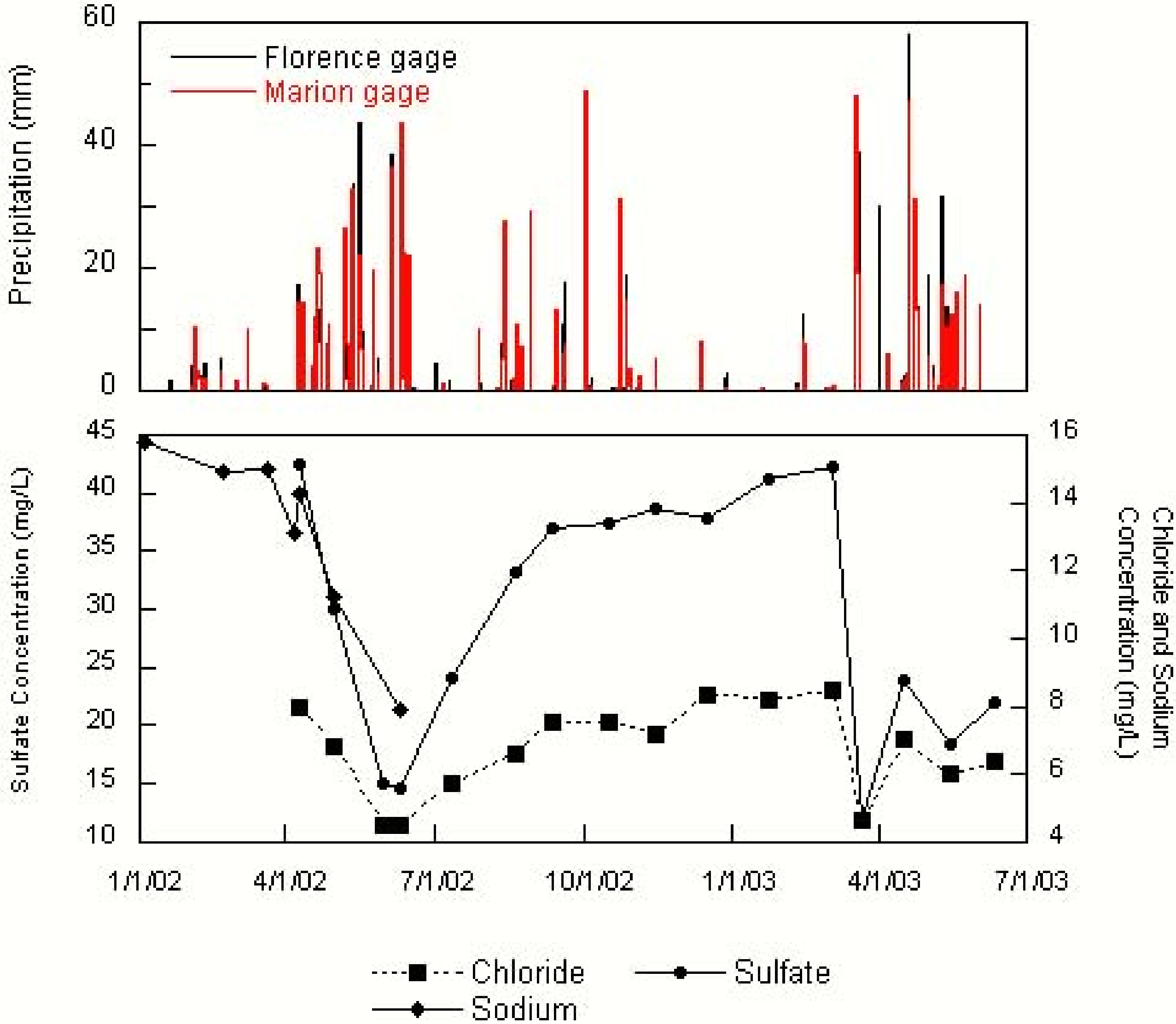


Figure 54. Time vs sodium, sulfate, and chloride concentration also showing time distribution of precipitation at Marion (red) and Florence stream gages.

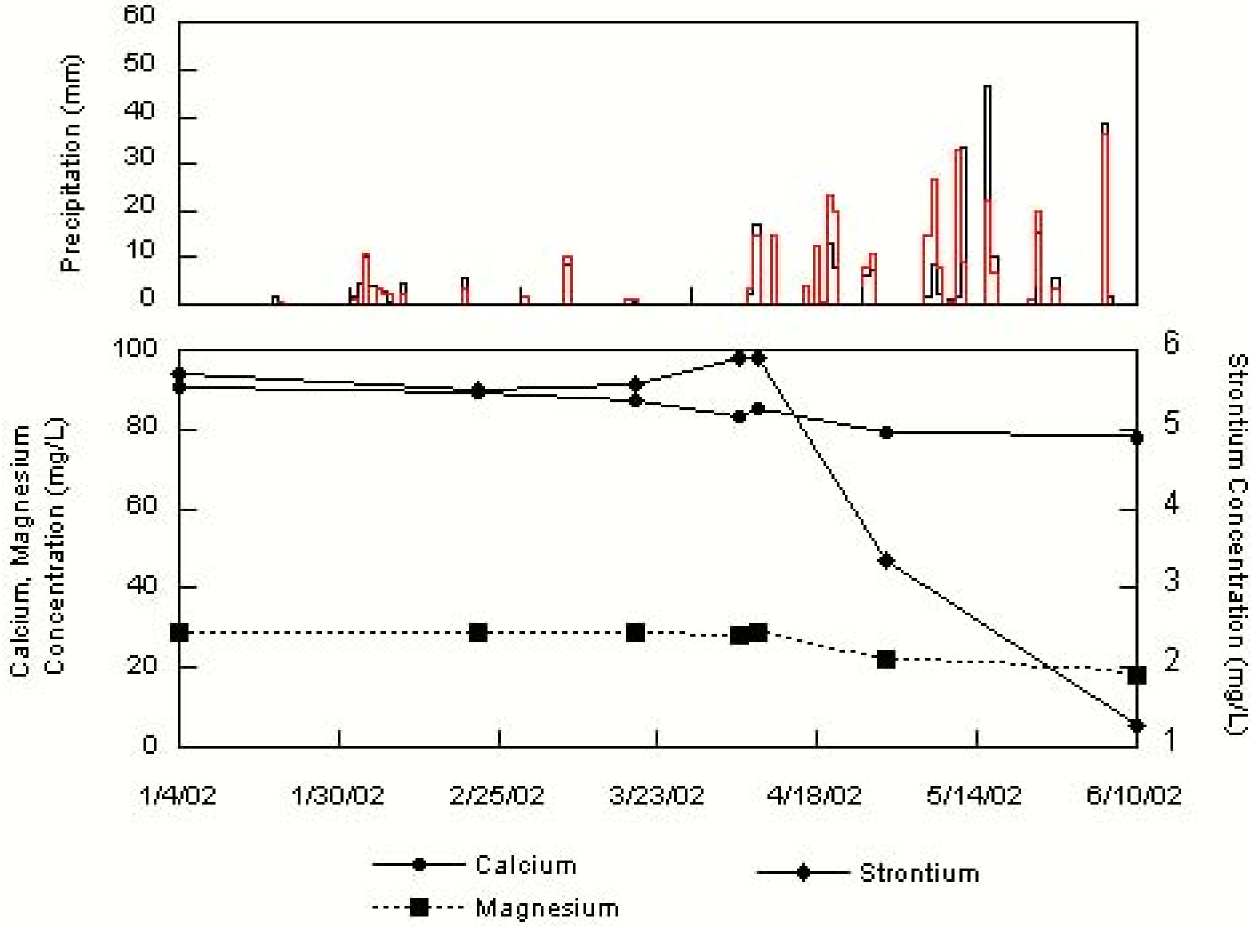


Figure 55. Time vs calcium, magensium and strontium concentration also showing time distribution of precipitation at Marion (red) and Florence stream gages.

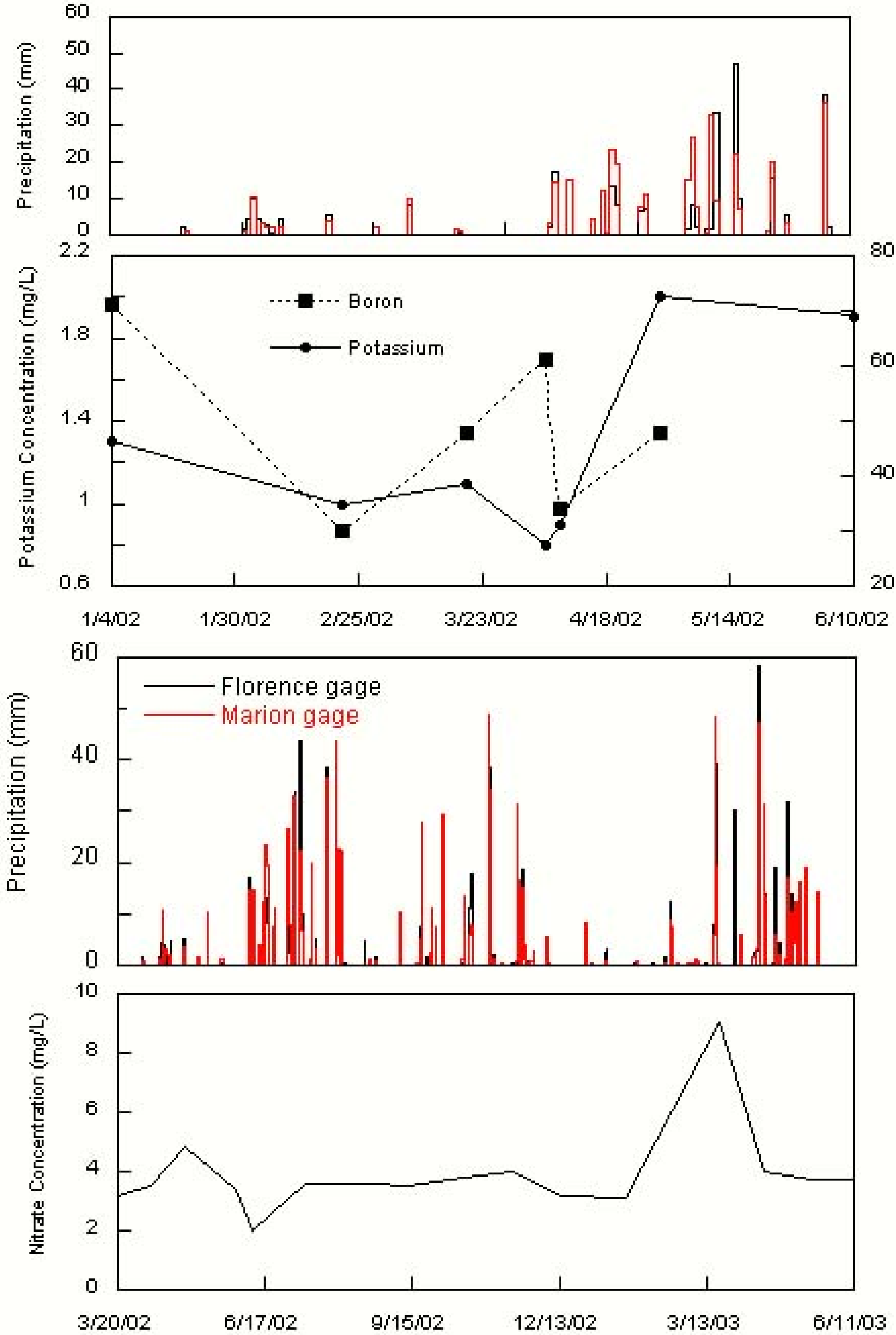


Figure 56. Time vs potassium, boron, and nitrate concentration also showing time distribution of precipitation at Marion (red) and Florence stream gages.

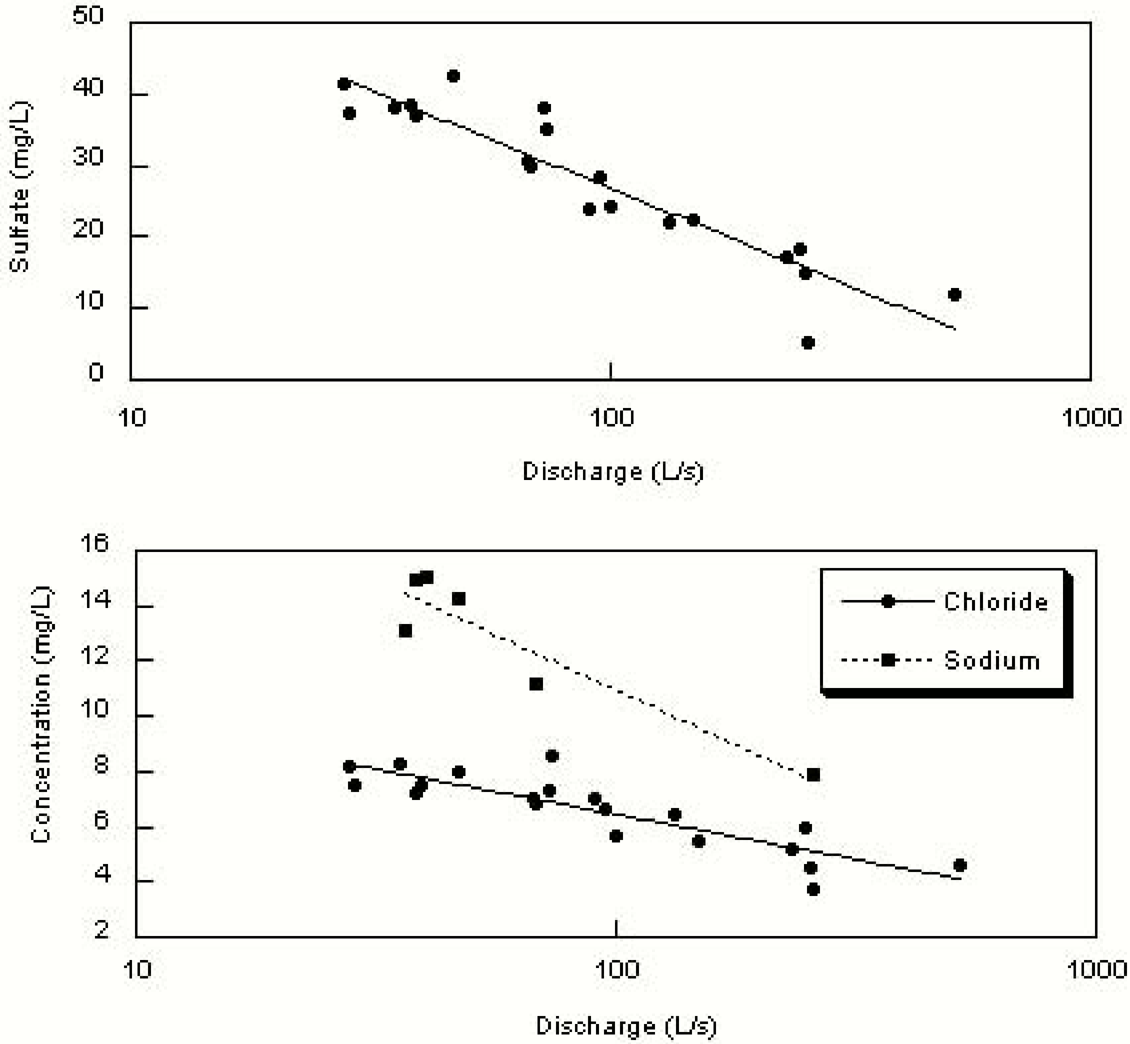


Figure 57. Discharge vs sulfate, chloride, and sodium concentration for Crystal spring.

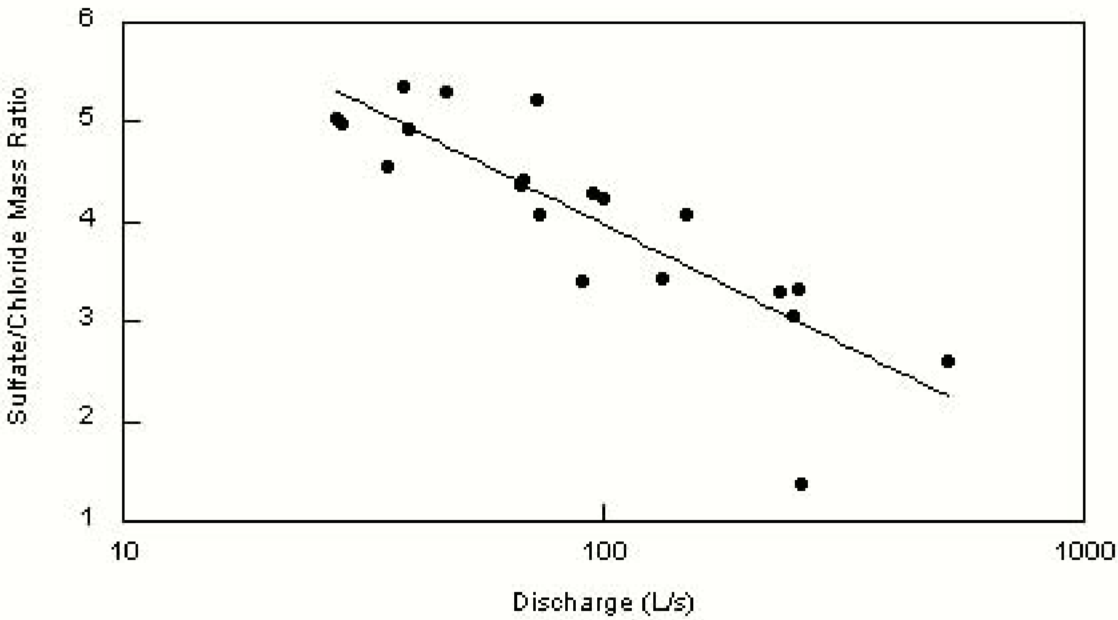


Figure 58. Discharge vs the sulfate/chloride mass ratio for Crystal spring.

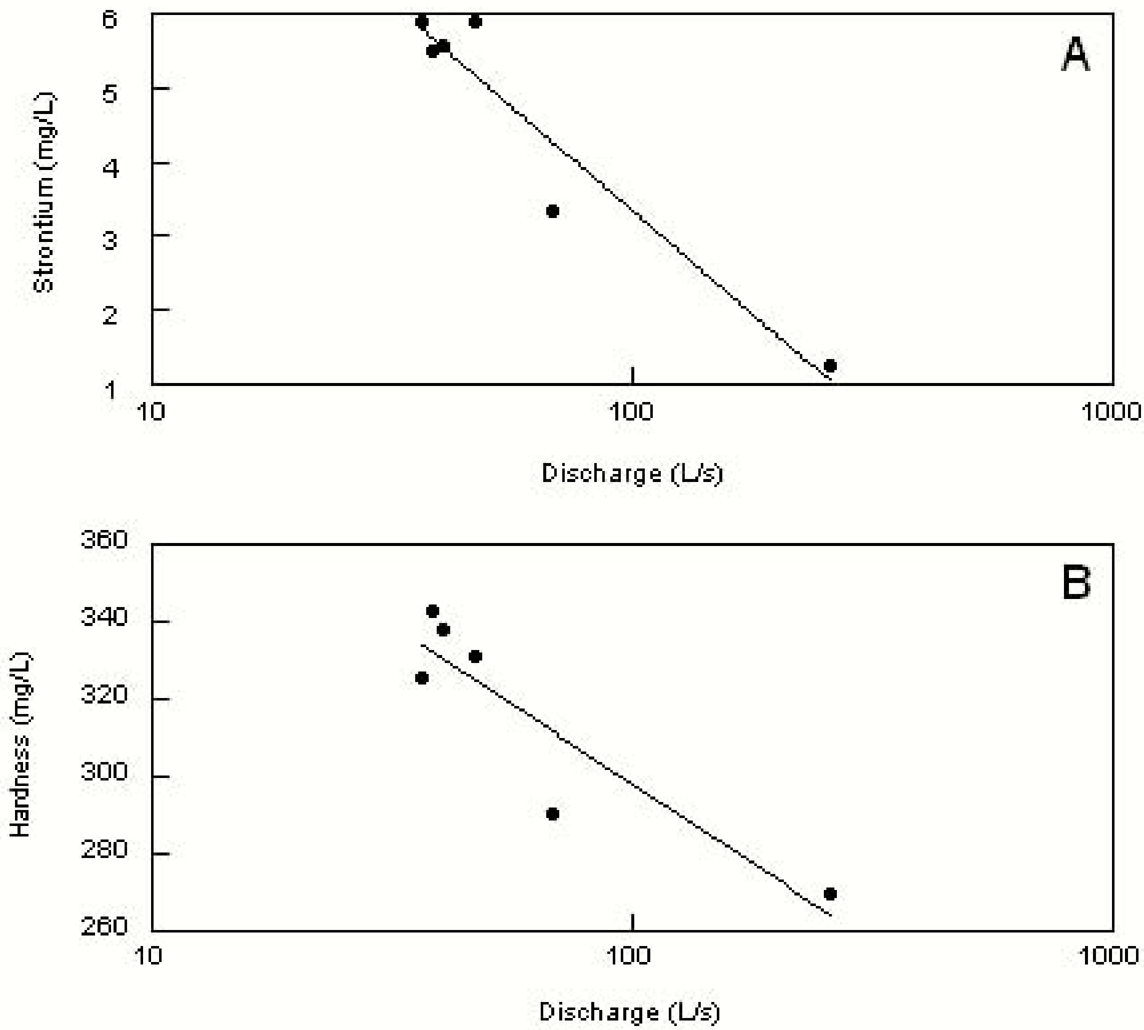


Figure 59. Discharge vs strontium concentration (A) and hardness (B) for Crystal spring.

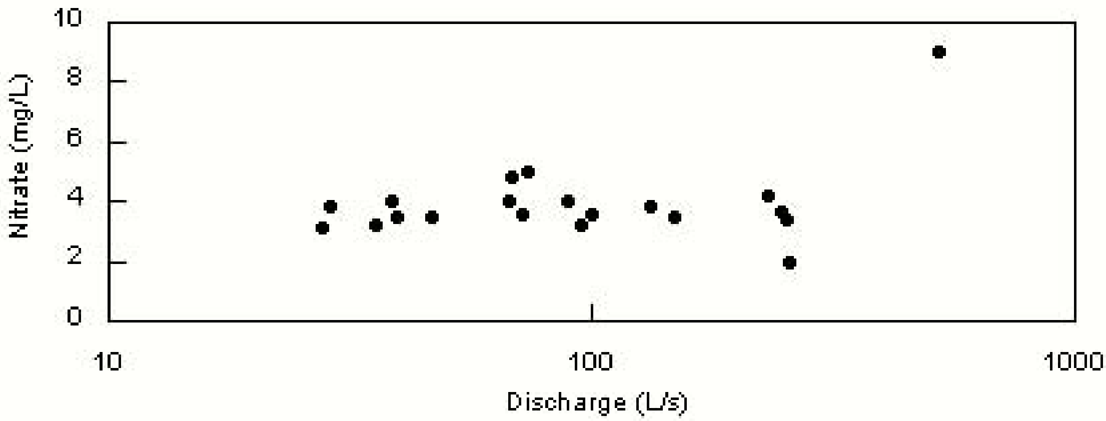


Figure 60. Discharge vs nitrate concentration for Crystal spring.

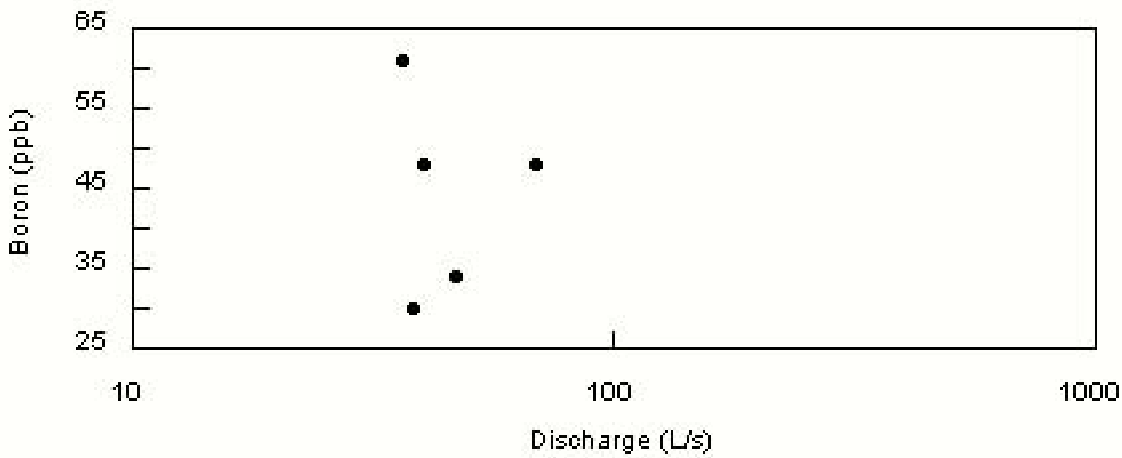


Figure 61. Discharge vs boron concentration for Crystal spring.

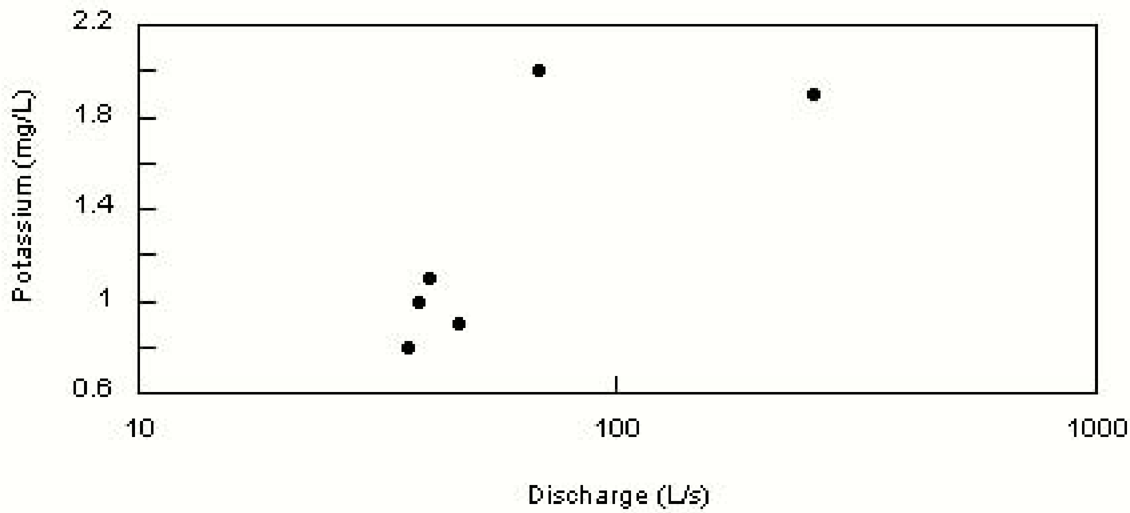


Figure 62. Discharge vs potassium concentration for Crystal spring.

Spring discharge turbidity is frequently related to precipitation events (See for example Massei et al., 2003). Daily measurements of spring discharge turbidity are taken by city water plant personnel to assist them in assessing the quality of the finished water from their treatment process. Time series plots of spring discharge turbidity were generated for the 2000 and the 2002-2003 time periods of the study (Figures 63, 64). Time series plots of precipitation at the Marion and Florence stream gages were generated for the same time periods to compare with the turbidity time series. (Figures 63, 64). The plots show that discharge turbidity is low with small fluctuations during extended dry periods. During the later summer and fall months rainfall events tend to produce only small amounts of precipitation insufficient to mobilize and flush some of the accumulated sediment from the karst conduit system. However, during high intensity precipitation events and extended wet periods there is sufficient water to input new sediment from the surface into and entrain accumulated sediment within the karst system. As a result the discharge turbidity abruptly increases by an order of magnitude or more over pre-storm event levels and is followed by a recession that parallels a recession in spring discharge (Figures 63, 64). As expected, discharge and turbidity are highly correlated (Figure 65).

A comparison of the 2002-2003 precipitation, spring discharge, and turbidity data with the water-level data from the monitoring well shows that most of the order of magnitude or greater fluctuations in turbidity are correlated with large-scale fluctuations in water levels in the aquifer adjacent to Martin Creek (Figure 66). The high water levels in June 2002 and in March through early June 2003 occurred when there was flow in Martin Creek following periods of precipitation. The June through September 2002 recession in turbidity follows a wet spring period in March through mid-June of that year. A series of storms from October 21 through November 3, 2002 caused an abrupt rise in turbidity up to 100 TU at Crystal spring and resulted in a 3.5-ft (1.07-m) rise in water level in the observation well. Field observation on November 15 indicated that there was no flow into Sinkhole 3, but Sinkhole 1 was totally submerged from this previous series of events and receiving flow from Martin Creek.

11.8 Hett's Springs Water Quality

An upper spring is located in SW, NW, NW, SE Sec. 10, T. 20 S., R. 5 E. and discharges from the middle of the Florence Limestone directly into Bruno Creek. The upper spring was sampled during the August 2000 reconnaissance and the Winter and late Spring 2002 sampling events and the lower spring during the Winter and Spring 2002 sampling events. The lower spring, located in SW, SW, SW, SE Sec. 10, T. 20 S., R. 5 E., is used for domestic and stock-watering purposes and also discharges to the creek from the middle of the Florence Limestone. Both springs are less than less than 1,500 m apart and appear to discharge from approximately the same horizon within the Florence.

The data from the Winter and Spring 2002 sampling events show that the quality of water discharging from the upper spring differs slightly from that of the lower spring (Appendices 1 and 2). Both springs discharge a low dissolved solids, Ca-HCO_3 type water. Water samples from the upper spring have a higher specific conductance, Ca, Na, HCO_3, Cl, SO_4, and NO_3 than the samples from the lower spring.

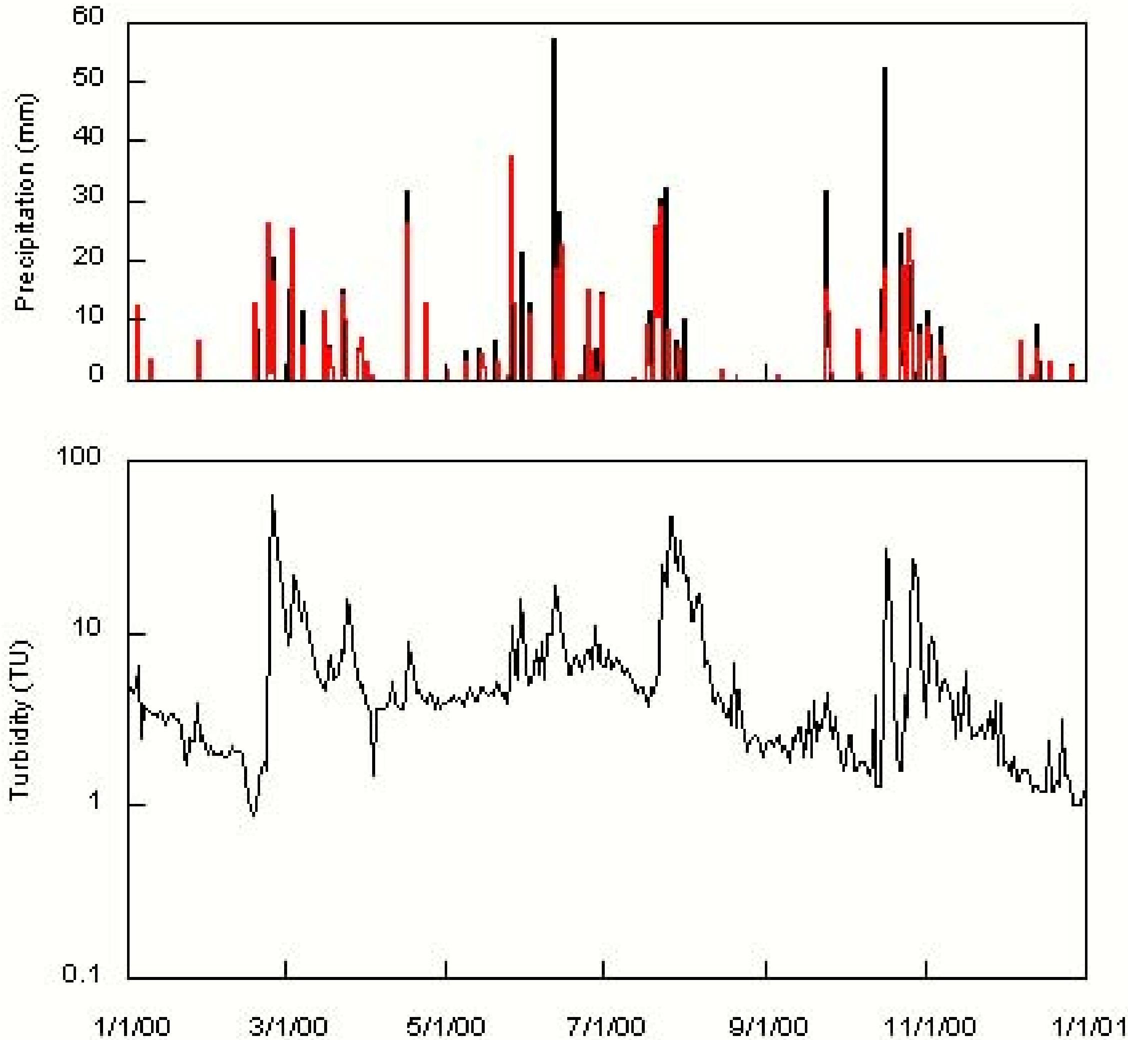


Figure 63. The 2000 turbidograph for Crystal spring. Daily precipitation data are from the Marion (red) and Florence (black) weather stations.

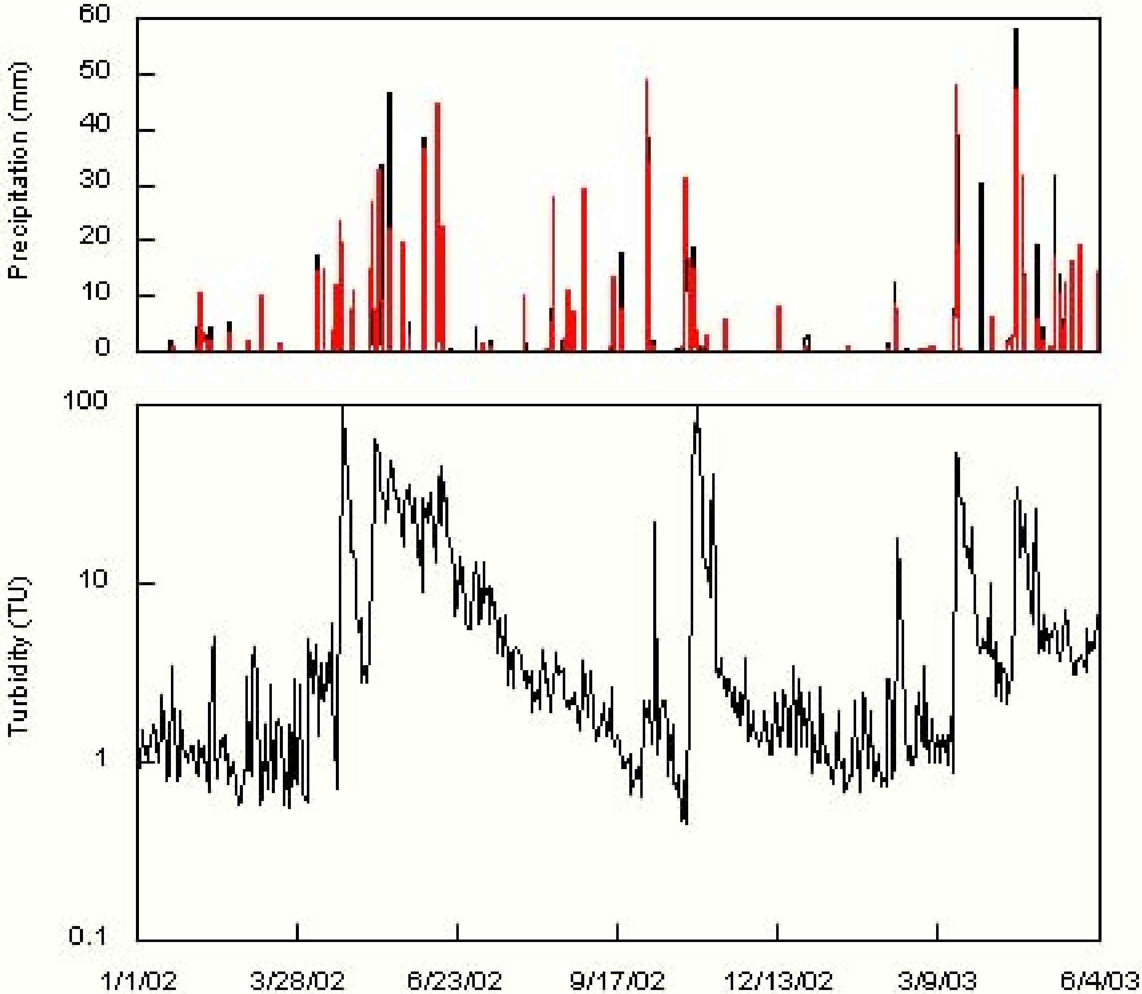


Figure 64. The 2002-3 turbidograph for Crystal spring. Daily precipitation data are from the Marion (red) and Florence (black) stream gages on the Cottonwood River.

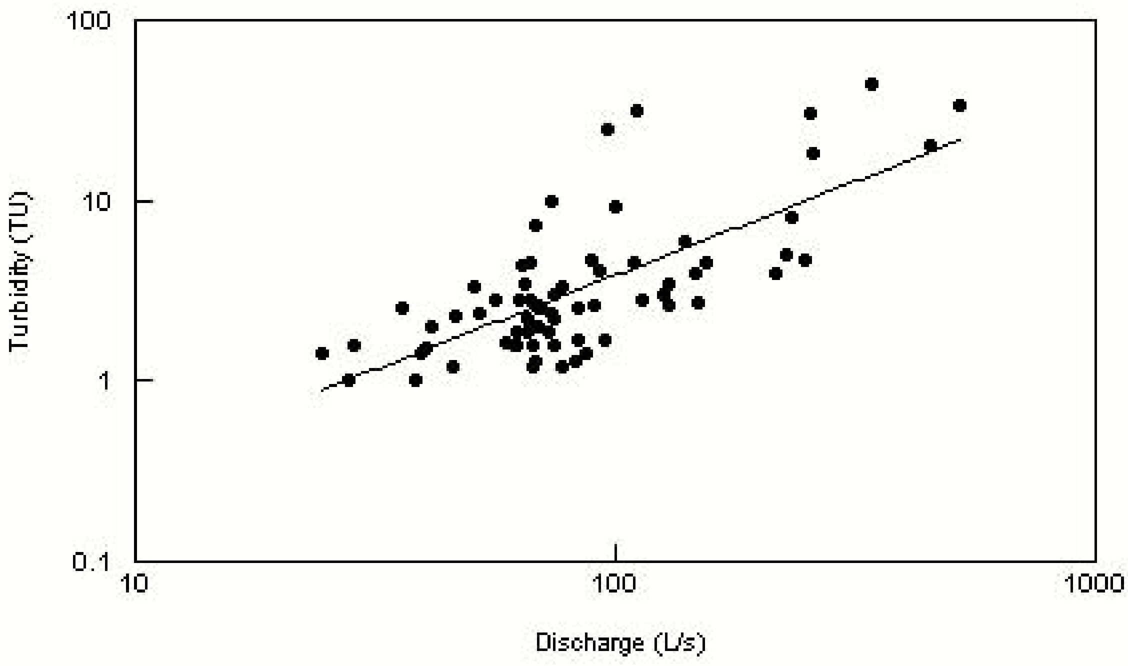


Figure 65. Crystal spring discharge vs turbidity for the 2000 and 2002-3 project years.

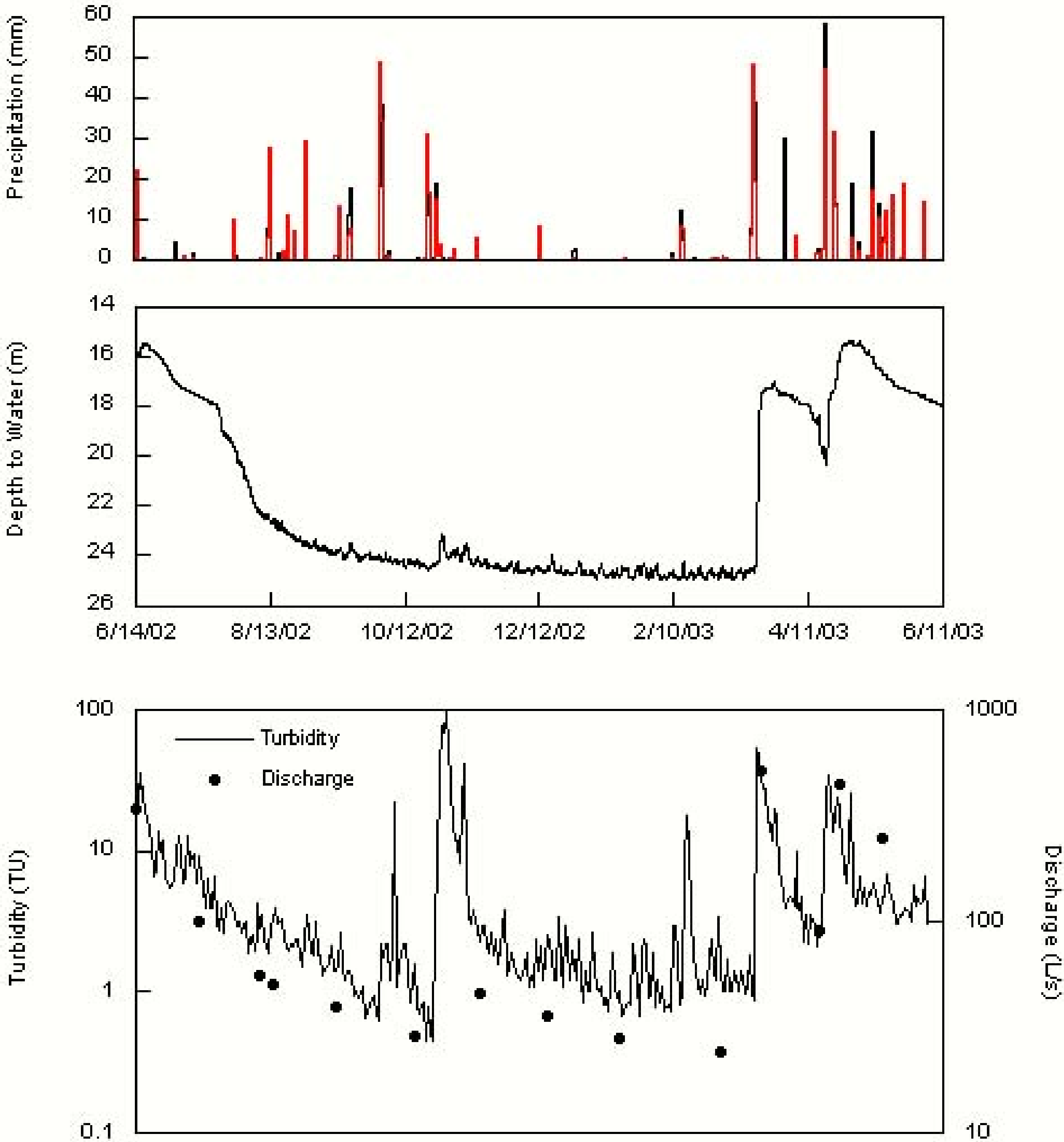


Figure 66. Time vs depth to water in the monitoring well, Crystal spring discharge and turbidity, and precipitation at the Marion (red) and Florence (black) stream gages on the Cottonwood River.

11.8.1 Wet/Dry Season Fluctuations

Comparison of the analytical results from the Winter and Spring 2002 sampling events shows the impact of the spring rainfall events on water quality. Specific conductance values for the samples collected from both springs in the spring are much lower than for the samples that were collected during the dry winter period. Chloride and the SO_4/Cl ratio changed very little between the winter and spring sampling events in water samples from the lower spring (Appendices 1 an 2). However, the Cl concentration was higher and the SO_4/Cl ratio slightly lower in the winter sample from the upper spring than the concentrations of these constituents in the spring sample. The lower spring experienced a slight increase in NO_3 between the winter and spring sampling events (2.4 to 3.1 mg/L), while the upper spring experienced a larger increase (7.6 to 10.2 mg/L). The increase in nitrate concentrations in the discharge from both springs is attributable to an influx of nitrate from infiltrated water that has moved through the soil zone and into the ground-water system.

11.9 Ground-water Quality

Most of the well water samples are either a Ca, Mg - HCO_3 or Ca - HCO_3 water with generally low total dissolved solids concentrations (Appendices 1 and 2). These water types and the low dissolved solids concentrations are fairly typical of ground waters from shallow aquifers in limestones and dolomites. Total dissolved solids range from 353 mg/L to 3,138 mg/L in the Winter 2002 samples and 311 mg/L to 1,679 mg/L in the Spring 2002 samples. Strontium concentrations ranged from 0.57 mg/L to 21.9 mg/L in the Winter 2002 samples and 0.56 mg/L to 19.3 mg/L in the Spring 2002 samples. Fluoride concentrations ranged from 0.31 mg/L to 1.05 mg/L in the Winter 2002 samples and 0.30 to 0.96 mg/L in the Spring 2002 samples. Boron concentrations ranged from 33 ppb (parts per billion) to 4,441 ppb in the Winter 2002 samples and <15 ppb up to 2,068 ppb in the Spring 2002 samples.

Nitrate concentrations in the Winter 2002 ground-water samples range from 1.0 mg/L to 25.4 mg/L and from 1.7 mg/L to 85.1 mg/L in the Spring 2002 samples. From their national assessment of water-quality, Mueller and Helsel (1996) used a background NO_3 concentration of 2 mg/L as the upper limit for ground water that has not been impacted by human activities. The wide range of nitrate concentrations in the samples is indicative of varying degrees of contamination from near surface sources of nitrate. It is likely that most of the contamination has resulted from poor well construction (O'Connor, 1984, Chaffee, 1988). In this study, nearly all of the samples show an increase in NO_3 concentration between the Winter and Spring 2002 water sampling events (Appendices 1 and 2).

To minimize the possibility of drawing erroneous conclusions from the limited chemical data, wells were eliminated from further analysis if the NO_3 concentrations in the Winter 2002, the Spring 2002 or both samples exceeded 9 mg/L. This limit was chosen because the NO_3 concentration range in Crystal spring discharge during the study was 2 mg/L to 9 mg/L with a median concentration of 3.6 mg/L. Using this criteria, the sites eliminated included well 2 in the SW, SE, SW Sec. 25, T. 19 S., R. 5 E., the windmill well in NW, NW, NE Sec. 31, T. 20 S., R. 5 E., and the stock well in SE, SE, SE, Sec. 12, T. 20 S., R. 4 E. Well 1 in the SW, SE, SW Sec.

25, T. 19 S., R. 5 E. was also eliminated because it is located approximately 100 m away from well 2.

The two wells located in SE, NE, NE Sec 2, T. 20 S., R 5 E. were also eliminated from further analysis. Winter and Spring 2002 samples are a blend of waters from a battery of two wells completed in the Barneston aquifer (Appendices 1 and 2). The dissolved solids concentrations of 3,138 mg/L and 1,679 mg/L for the winter and spring samples, respectively, and the Ca, Mg, Na – HCO_3 water type indicate that these are not typical ground water samples from the Barneston aquifer. The site is located on the Towanda Limestone Member outcrop. The high NO_3 values for both the winter and spring samples and the large decrease in total dissolved solids concentration from winter to spring gives a strong indication of contamination by leakage of water into one or both wells from the near surface. The older of the two wells is a windmill well of uncertain age that is located near a livestock pen and a plowed field. It is likely that the high NO_3 in the ground-water samples is coming from this source.

Plots of the SO_4/Cl mass ratio vs. Cl for the dry season (combined August 2000 and Winter 2002) and wet season (Spring 2002) data reveal distinct chemical differences between Barneston ground water and the spring and surface water in the study area (Figures 67, 68). The August 2000 and the Winter 2002 samples were grouped together because both sets were collected during an extended dry period.

The Cl concentration range in the dry season samples appears to be less for the ground water than for the spring discharge and surface water in the project area. Chloride concentrations in wet and dry season well samples generally fall in the range of 4 mg/L to 8 mg/L. The dry season SO_4/Cl ratios tend to be higher and more variable in the ground water than in the spring discharge from the Barneston or surface water with the values ranging from 6 up to about 15. The SO_4/Cl ratios in spring discharge are similar to those in the surface water. The SO_4/Cl ratios in the Spring 2002 (wet season) in some of the well samples generally remain higher than the ratios in the spring and surface water samples. However in the sample from a well in SE, NE, NE Sec. 18, T. 20 S., R. 5 E., the ratio value is comparable. The SO_4/Cl ratios in the spring and surface waters, including Crystal spring, are in the range of 3 or less.

Field observations in the project area indicate that baseflow in the upper reaches of the streams is sustained by spring discharge from the Towanda Limestone Member (Doyle Shale) during the drier parts of the year. Perennial spring discharge from the Stovall Limestone Member (Gage Shale), Cresswell Limestone Member (Winfield Limestone) and the Herrington Limestone (Nolans Limestone) was not observed within the project area. A number of perennial springs discharge from the Towanda to the unnamed drainage in Secs. 25 and 36, T. 20 S., R. 4 E and to the upper reaches of Martin Creek in Sec. 13, T. 20 S., R. 4 E. and in Sec. 18, T. 20 S., R. 5 E. Other sources, including the soil zone may also contribute to the observed baseflow. Streamflow enters the Barneston aquifer where the streams cross its outcrop through Sinkhole 1 and elsewhere through fractures in the Crystal spring catchment. The flow into Sinkhole 1, the uppermost in the Martin Creek drainage, was visually estimated to be on the order of 0.17 L/s and 0.06 L/s, respectively, during the August 2000 and Winter 2002 sampling events. Note that the Cl and SO_4/Cl values are slightly lower in the August 2000 sample than for the Winter 2002

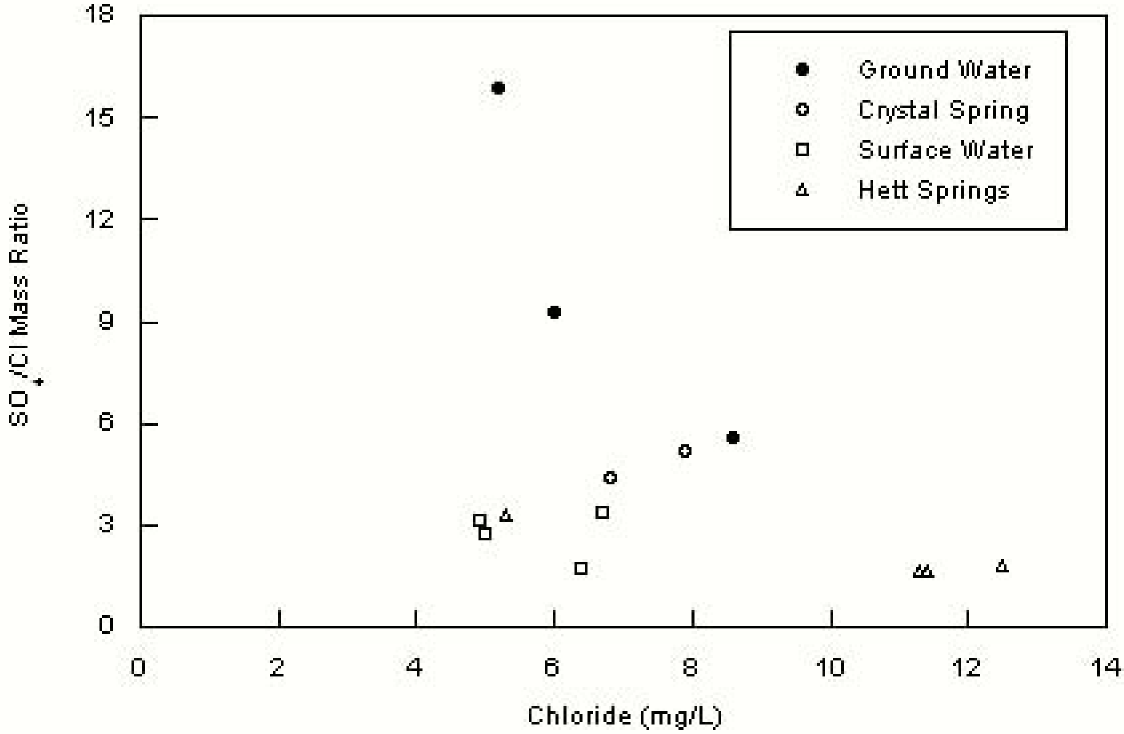


Figure 67. Dry season (August 2000 and Winter 2002) chloride and sulfate to chloride mass ratios for ground water, surface water, and spring samples.

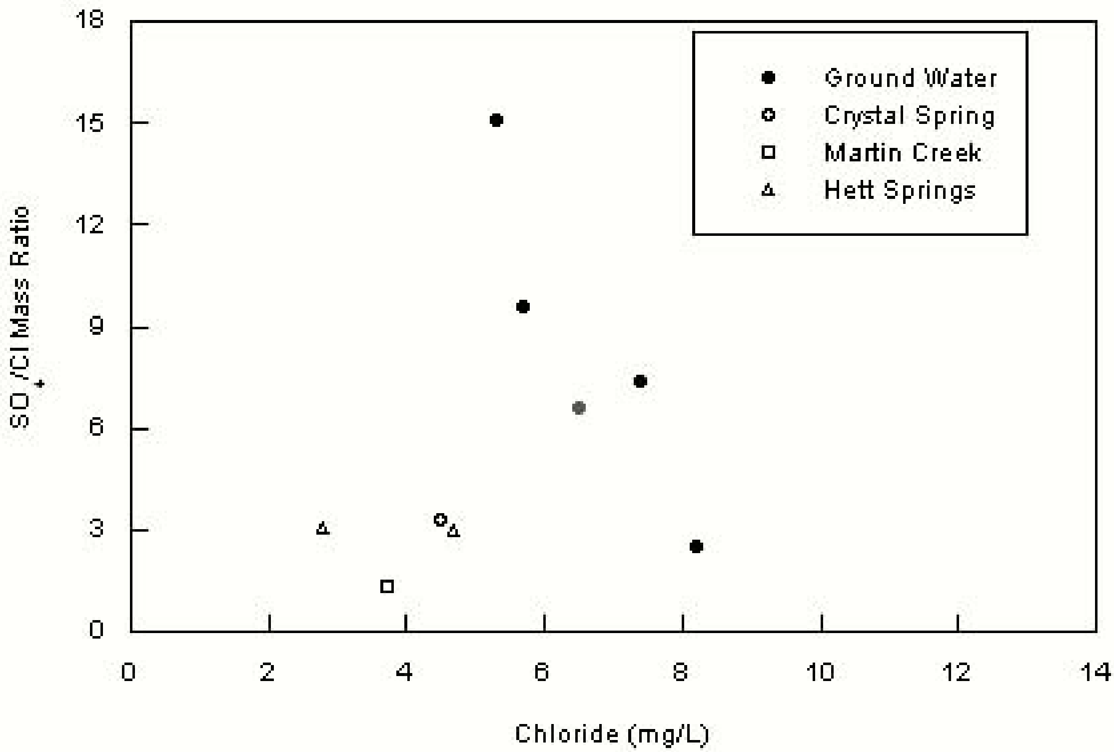


Figure 68. Wet season (Spring 2002) chloride and sulfate to chloride mass ratios for ground water, surface water, and spring samples

sample. Thus, the dry season discharge from the springs is a mixture of recharge that originated as baseflow to the stream and regional ground-water flow from the Barneston. Similar sources of recharge may supply water to the Hett springs during dry periods, but this was not investigated in any detail during the project.

11.9.1 Wet/Dry Season Fluctuations in Ground-water Quality

With the small number of wells that were sampled it is not possible to fully evaluate seasonal effects on ground-water quality in the project area. However, two wells located near Martin Creek and one well located on the north valley wall of the Cottonwood River (NE, SW, NE, SE Sec. 27, T. 20 S., R. 4 E.) near Crystal spring are suitable for this purpose. Winter and Spring 2002 Cl and SO_4/Cl ratios illustrate the changes in these parameters between the dry and wet season sampling events (Figure 69). Also plotted on the figure are the changes in these parameters in the surface water in Martin Creek and the discharge from Crystal spring. The well in NW, NE, NW Sec. 30, T. 20 S., R. 5 E. is situated south of Martin Creek between the sinkholes and Crystal spring. The water samples from this well and the well near the Cottonwood River show very little change in Cl and the SO_4/Cl ratio over time. In contrast, the Spring 2002 water sample from an unused domestic well located east of Martin Creek in SE, NE, NE Sec. 18, T. 20 S., R. 5 E. shows a significant reduction in the SO_4/Cl ratio between the sampling events with negligible change in the Cl concentration. This decrease mirrors a similar decrease between the winter and spring samples from Crystal spring and from Martin Creek.

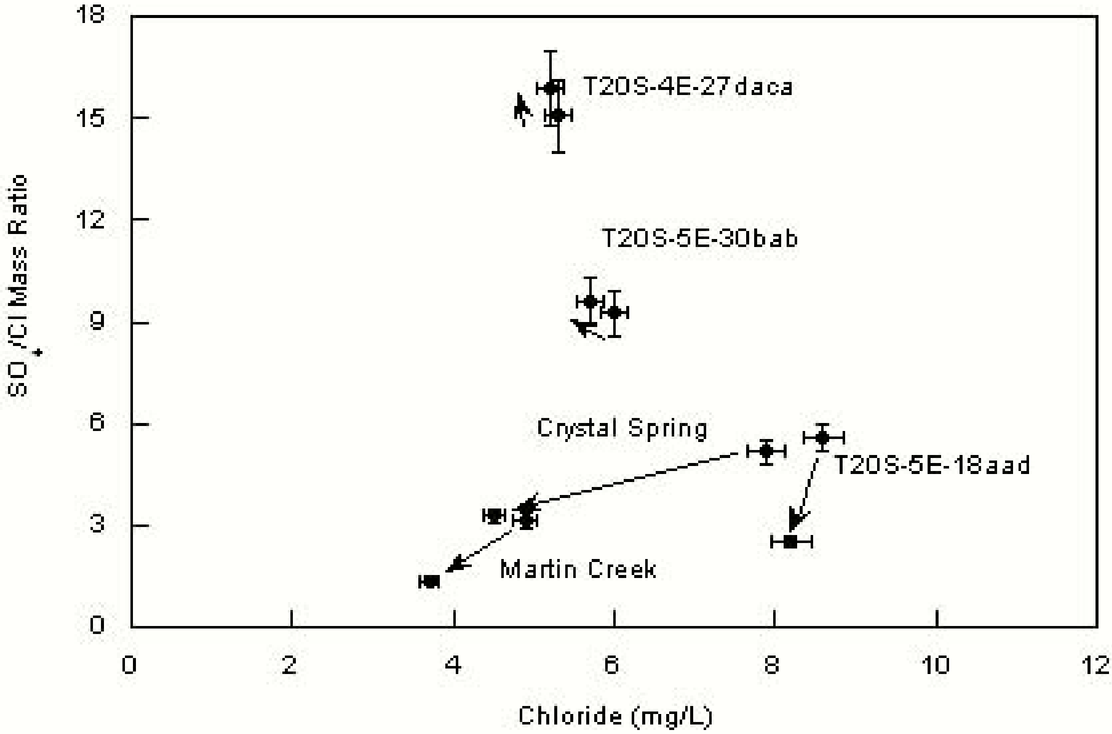


Figure 69. Changes in chloride and the sulfate to chloride mass ratio in ground water, Martin Creek surface water, and Crystal spring discharge between the Winter 2002 and Spring 2002 water sampling events. Also shown are the error bars associated with each parameter.

12.0 A Preliminary Assessment of the Flow of Water Through the Crystal Spring Catchment

12.1 Overview of hydrogeologic classifications and models of carbonate aquifers

Karst aquifer systems consist of two coupled dynamic systems (White, 1971). In the conduit part of the flow system the residence time of water is on the order of hours to days in duration, but in the diffuse part the residence time is much longer and can be on the order of millennia. White (1969) defined the characteristics of diffuse- and free-flow (conduit flow) karst aquifers in low-relief, flat-lying carbonate sequences. These are presented in Table 3.

Table 3. White's (1969) classification of carbonate aquifers (from White, 1971).

Flow Type	Hydrological Control	Associated Cave Type
I. DIFFUSE FLOW	GROSS LITHOLOGY: Shaly Limestones; crystalline dolomites; high primary porosity	Caves rare, small, and have irregular patterns
II. FREE FLOW	THICK, MASSIVE, SOLUBLE ROCKS	Integrated conduit cave systems
A. PERCHED	Karst system underlain by imperviuos rocks near or above base level.	Cave streams, perched; often have free air surface
1. Open	Soluble rocks extend upward to level surface	Sinkhole inputs: heavy sediment load; short channel morphology caves
2. Capped	Aquifer overlain by impervious rock	Vertical shaft inputs: lateral flow under capping beds; long integrated caves
B. DEEP	Karst system extends considerable depth below base level	Flow is through submerged conduits
1. Open	Soluble rocks extend upward to level surface	Short, tubular abandoned caves, likely to be sediment-choked
2. Capped	Aquifer overlain by impervious rock	Long. Integrated conduits under caprock. Active level of the system inundated
III. CONFINED FLOW	STRUCTURAL AND STRATIGRAPHIC CONTROLS	
A. ARTESIAN	Impervious beds which force flow below regional base level	Inclined 3-D network caves
Sandwich	Thin beds of soluble rock between impervious beds	Horizontal 2-D network caves

In “diffuse flow” aquifers, the dominant lithology is shaly limestone or dolomite and the solutional cavities are limited in size and number to a network of solution widened joints and bedding planes that exhibit a high degree of connectivity. In outcrop karst landforms are subdued. Diffuse-flow dominated aquifer systems are characterized by a well-defined water table and discharge is through a small number of seeps and springs. In contrast, the defining characteristic of “free-flow” or “conduit-flow” aquifers is ground-water flow that is localized in a well-integrated system of conduits with flow velocities on the order of 0.03 m/s (0.1 ft/s) and frequently in the turbulent range (White, 1971). Hydrostatic head, conduit hydraulic characteristics, and the volume of recharge govern conduit flow. In these systems the main conduits are often underground extensions of the surface drainage with the flow carrying a significant suspended load and passageways that act as temporary storage of the suspended sediment following a flood crest. The hydraulic gradients within the conduits are typically very low. Discharge from these systems is through a big spring with a catchment area of 25.9 to 259 km^2 (10s to 100s of mi^2) (White, 1971).

Of special note here is the sandwich aquifer, shown at the bottom of Table 3. The sandwich aquifer is capped and perched and thin in comparison to the total bed thickness above regional base level and typically less than 12.2 m (40 ft) thick. Ground-water flow is along solution-widened joints and is retarded by a lack of concentrated recharge from overlying beds (White, 1969). In his 1971 revision of the classification, White implies that sandwich aquifers are local in extent but can grade into regional aquifers governed by diffuse flow if they are laterally extensive. Most of the karst aquifers in the Flint Hills region are of the sandwich type (Macpherson, 1996).

The 1971 revision of the free-flow systems section is a two-way classification based on whether the catchment relief, aquifer thickness, and relative position of the base of the carbonate aquifer with respect to regional base level (Table 4).

Table 4. Classification of free-flow, carbonate aquifers based on the thickness with respect to regional base level (After White, 1971).

	Regional Base Level Elevation > Carbonate Aquifer Base Elevation	Regional Base Level Elevation < Carbonate Aquifer Base Elevation
Vertical Relief Above Regional Base Level > Carbonate Aquifer Thickness	Capped	Perched/Capped
Vertical Relief Above Regional Base Level < Carbonate Aquifer Thickness	Open	Perched/Open

White (1971) also recognized that the transient response of springs to recharge events is also a characteristic that distinguishes diffuse from conduit-flow. The response of the conduit to recharge can be flashy with a steep rise in spring discharge immediately following recharge up to a peak discharge value, followed by a rapid recession to pre-recharge flows. In contrast, the transient response of springs fed by diffuse-flows is muted and delayed in time with a very long recession following peak discharge.

Smith et al. (1976) defined a mixed or hybrid aquifer system in which the relative importance of conduit and diffuse flows changes in response to transient events. They envisaged two main compartments for the temporary storage of water entering a mixed aquifer, the main conduits and the surrounding interconnected secondary porosity in the carbonate rock. During and immediately following recharge, the hydraulic head in the conduits is higher than in the surrounding aquifer and conduit water moves into the secondary porosity of the surrounding carbonate rock to create a form of bank storage. Once the floodwave has passed and the hydraulic gradient within the conduit begins to return to its pre-recharge event state, the flow direction between the conduit and the surrounding secondary porosity in the carbonate rock reverses, and the bank storage and the infiltration from the recharge event flows into the conduit. The resulting spring discharge hydrograph from this system displays the steep rise of the conduit-fed spring but also has the extended recession of the diffuse-flow spring. Diffuse- and conduit-flow systems are end members along a continuum of flow system types. Individual spring and catchment systems posess the characteristics of both end members to varying degrees (Quinlan, 1989). Examples of mixed systems include the Floridan aquifer in the Santa Fe River (Martin and Dean, 2001) and in the Little River drainage (Katz et al., 1998) of northern Florida and the Devils Ice Box in the Missouri Ozarks (Hallihan, 1996, 1998).

12.2 Conceptual model of the study area hydrologic system

It is obvious from the data collected during this study that the characteristics of both conduit- and diffuse-flow systems are present in the Barneston aquifer and the dominance of one flow system over the other is determined by the intensity and duration of recharge events within the Crystal spring catchment. Thus, the Barneston aquifer seems to act as a mixed aquifer system within the catchment. In terms of its extent most of the ground-water system is encompassed by diffuse flow and it is flow from this part of the aquifer that constitutes most of the discharge from Crystal spring. The conduit-flow part of the system takes up a small fraction of the aquifer but can dominate the flux as a result of recharge from storm events or wet periods. The following is a summary of the evidence in support of this conclusion.

12.2.1 Hydrostratigraphic evidence

The Barneston Limestone is underlain by the Matfield Shale and in parts of the catchment by the Holmesville Shale Member of the Doyle Shale. These low permeability units act as confining units. According to the White (1969) classification, the Barneston is not a sandwich aquifer because its thickness is greater than 12.2 m (40 ft). Of the bedrock units in the Council Grove and Chase Groups that underlie the Flint Hills, the Barneston has the greatest average thickness of 24.4 m (80 ft; Figure 70). The Cottonwood River acts as the local base level for the karst system. With the exception of where the Barneston underlies or is adjacent to the Cottonwood River valley westward from the vicinity of Crystal spring, the aquifer base is above the base level. Thus, the Barneston is a perched and locally capped aquifer (Tables 3 and 4).

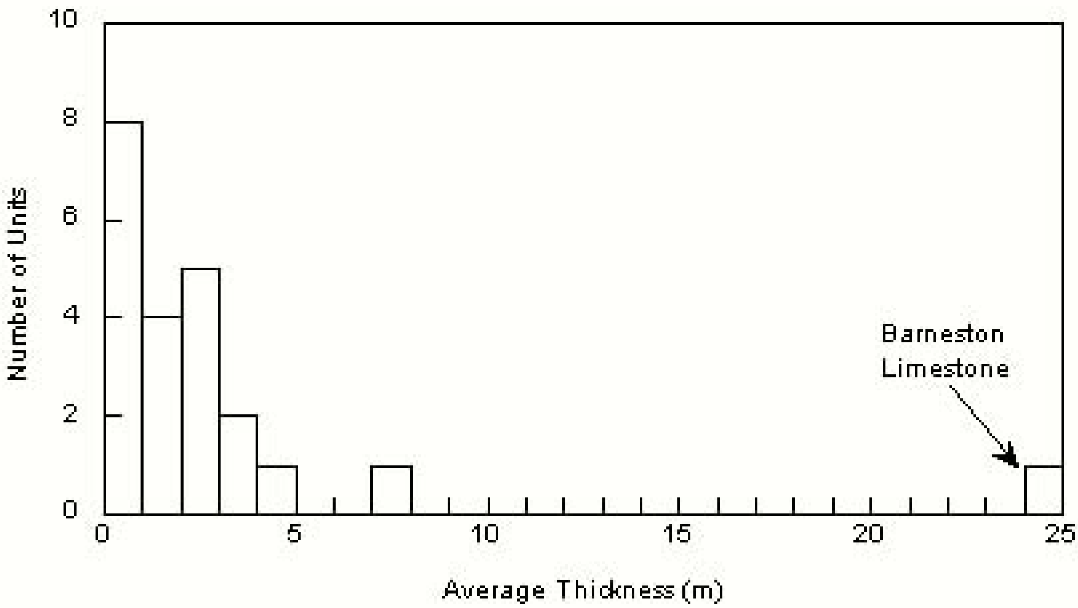


Figure 70. Histogram showing the average thickness of limestone units in the outcrop belt of the Council Grove and Chase Groups in Kansas based on average unit thicknesses in Zeller (1968).

The dominant lithology of the Barneston is limestone. However, the lithology of the upper Ft. Riley is shaly limestone interbedded with thin shales. Thin shales and shaly units are also sporadically distributed through the Florence. These units are not conducive to the development of conduits in limestone units (White, 1969; Hess et al., 1989). However, gypsum and halite have been reported in this part of the Barneston and in the Florence Limestone downdip of the outcrop belt. Early dissolution of these evaporites by infiltrating fresh water most likely initiated the process of secondary porosity development, but later limestone dissolution by fresh ground water continued and enhanced the process primarily along bedding planes.

Observations of the Barneston Limestone in outcrop, in the quarry, and from the drilling of the monitoring well near Martin Creek indicate a diversity of secondary porosity types from joints to solution conduits up to more than 1 m in diameter. In the Ft. Riley Limestone these features include solution-widened joints and bedding planes and short conduit segments 15 to 20 cm (6 in to 8 in) in diameter and approximately 2 m (6.5 ft) in length. Secondary porosity in the Florence is characterized primarily by solution-widened joints, thin zones consisting of small to large vugs (cm to several decimeters in size) near the bottom of the unit, and a small cave half or more filled with sediment near the top of the unit in non-cherty limestone. Hett spring issues from a small cave opening near the middle of the Florence in a section of cherty limestone which apparently maintains its size for a short distance behind the rock face. Both cave features seem to be associated with a solution-widened joint. No cavities were detected in the Barneston during the cable-tool drilling of the borehole for the monitoring well.

True sinkhole features normally associated with karst terrains do not appear to be present within the catchment. True sinkhole formation has been documented near Potwin in Butler County (Gordon, 1938). Sinkholes typically form as a result of limestone dissolution and collapse of overlying rock to form a depression. The features referred to as sinkholes in this project are areas where a bowl-shaped depression has been formed by fluvial erosion of highly weathered and fractured bedrock. Entry into the bedrock from the depression is through solution-enlarged joints. It is interesting to note that these sinkhole-like features have been identified only where there are bends in Martin Creek and not where the channel is straight. The general alignment of these sinkhole-like features also suggests control by fractures in the bedrock.

12.2.2 Hydrologic evidence

The hydrologic evidence suggests that the Barneston possesses many of the characteristics of the free-flow (conduit-flow) aquifers in White's (1969) classification (Table 3). The Barneston aquifer has only a single natural discharge point in a catchment area on the order of 25.9 km^2 to 51.8 km^2 (10 mi^2 to 20 mi^2). Measured flow rates ranged from approximately 28.3 L/s to 509 L/s (1 ft^3/s to 18 ft^3/s) measured during this study. Dye trace experiments indicate that Martin Creek stream bottom is in direct hydraulic connection with Crystal spring. Flow velocities from the October 2002 dye trace were calculated to be 0.018 m/s (0.06 ft/s) with an average residence time of approximately 2.5 days when spring discharge was approximately 28.3 L/s (1 ft^3/s). This suggests the existence of a well-developed conduit system between the sinkholes in Martin Creek and the spring.

The response of the Barneston aquifer and Crystal spring to recharge events provides further insight into the cycling of water through the karst system. The turbidograph and hydrograph of spring discharge show a rapid increase in turbidity and spring discharge from runoff-producing storm events and extended wet periods. In tandem with increases in these parameters at the spring, streamflow was generated in Martin Creek and there is an abrupt rise in water level in the monitoring well, located in the diffuse-flow part of the aquifer near Martin Creek. Two examples follow that describe what happens during stormy periods under different antecedent moisture conditions.

A series of storms passed over the study area over a 17-day period in late October - early November 2002 with most of the rainfall occurring on October 23, 24, and 27 (Figure 71). Prior to the storms, rainfall events had taken place sporadically during the hot summer months of the growing season. These storms had not produced sufficient precipitation to alter the trend of declining discharge at Crystal spring or the decline in the monitoring well water level over time. The monitoring well water level was approximately 4.6 m (15 ft) below the top of the Florence Limestone. Dry conditions had curtailed flow into Sinkhole 1 and the streambed was also dry. On November 15, the stream was flowing into a pool that covered Sinkhole 1, but there was no evidence that a pool had formed above Sinkholes 3 or 4 near the monitoring well or that streamflow had moved past Sinkhole 1. Spring discharge turbidity and monitoring-well water level did not begin to rise until October 26 and October 27, respectively (Figure 71). No water could be heard cascading into the well from above. Later in early November, a small increase in turbidity and a rise of water level accompanied some small rainfall events with a similar time lag.

In the second example, a series of higher intensity storms passed through the study area March 13-15, 2003 (Figure 72). Prior to these storm events, on March 3, the snow cover was melting and the soil was saturated with snowmelt. Martin Creek was dry downstream of Sinkhole 1 and the flow into Sinkhole 1 was estimated to be less than 0.06 L/s (0.002 ft^3/s). Crystal spring discharge turbidity and the monitoring well water level began to rise on the same day on March 14, when most of the rainfall occurred. Precipitation was reported from the Cottonwood River stream gage near Florence on April 1, but not at the Marion stream gage and apparently did not have any impact on water levels or turbidity.

The spring discharge and turbidity and monitoring-well water-level data from stormy periods suggest that the diffuse-flow part of the aquifer system acts is the overflow for conduit water depending on the intensity and duration of input from precipitation and streamflow generation. Smaller rainfall events may not generate sufficient runoff for streamflow in the reach of Martin Creek to cross the width of the Barneston Limestone outcrop belt if there is little antecedent moisture in the basin or during the growing season.

In the first example, the time interval between the start of the rainy period and the increase in spring turbidity is close to the 60-hr residence time for water in the conduit system calculated from the dye trace when discharge was on the order of 28.3 L/s (1 ft^3/s). The increase in water level in the monitoring well could have resulted from downward infiltration of water through the soil zone and into the solution-enlarged fractures of the diffuse-flow part of the system.

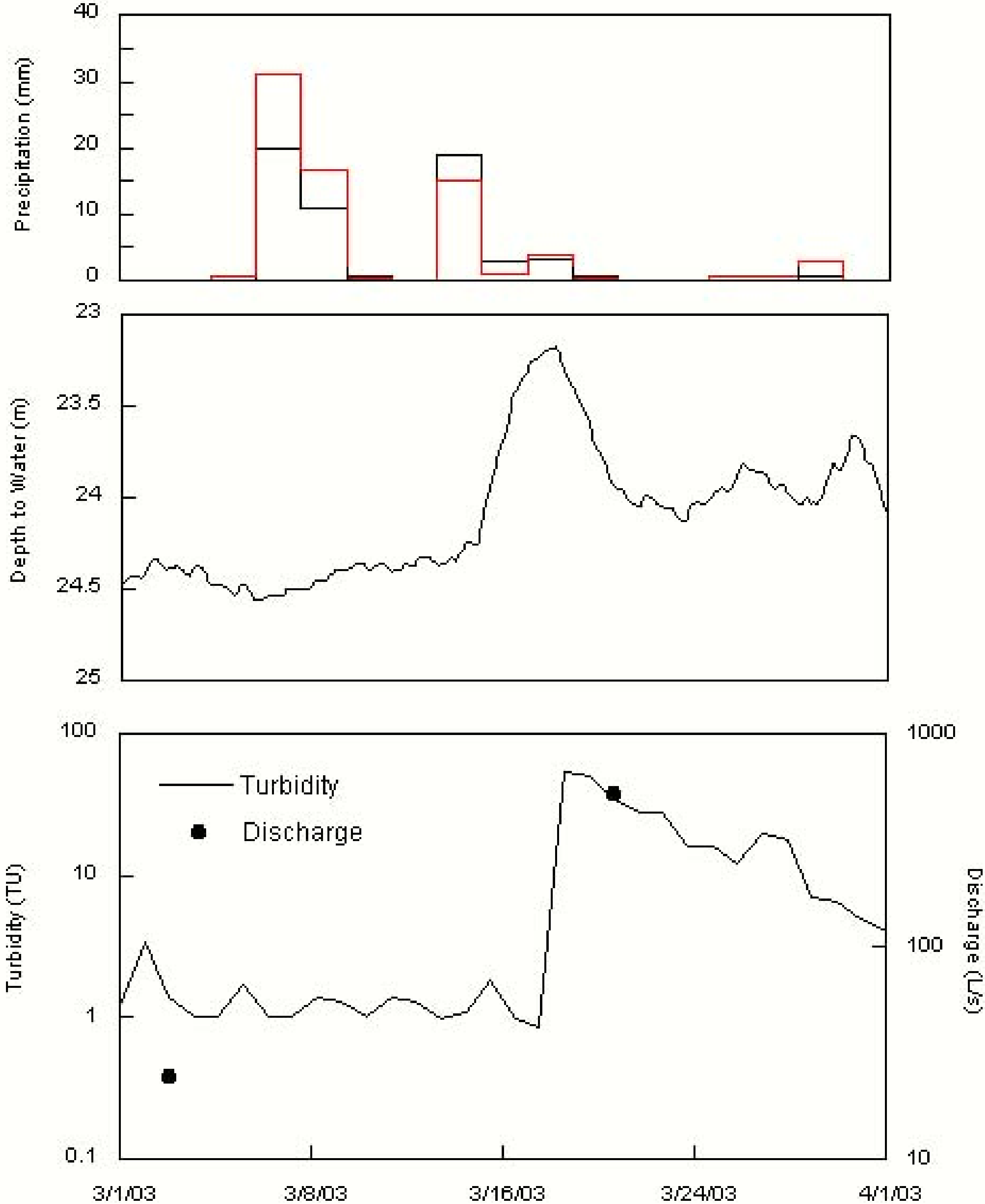


Figure 71. Precipitation, depth to water in the monitoring well, and turbidity in the discharge from Crystal spring for the period October 20 - November 6, 2002. Precipitation data are from the Marion (red) and Florence (black) stream gages on the Cottonwood River.

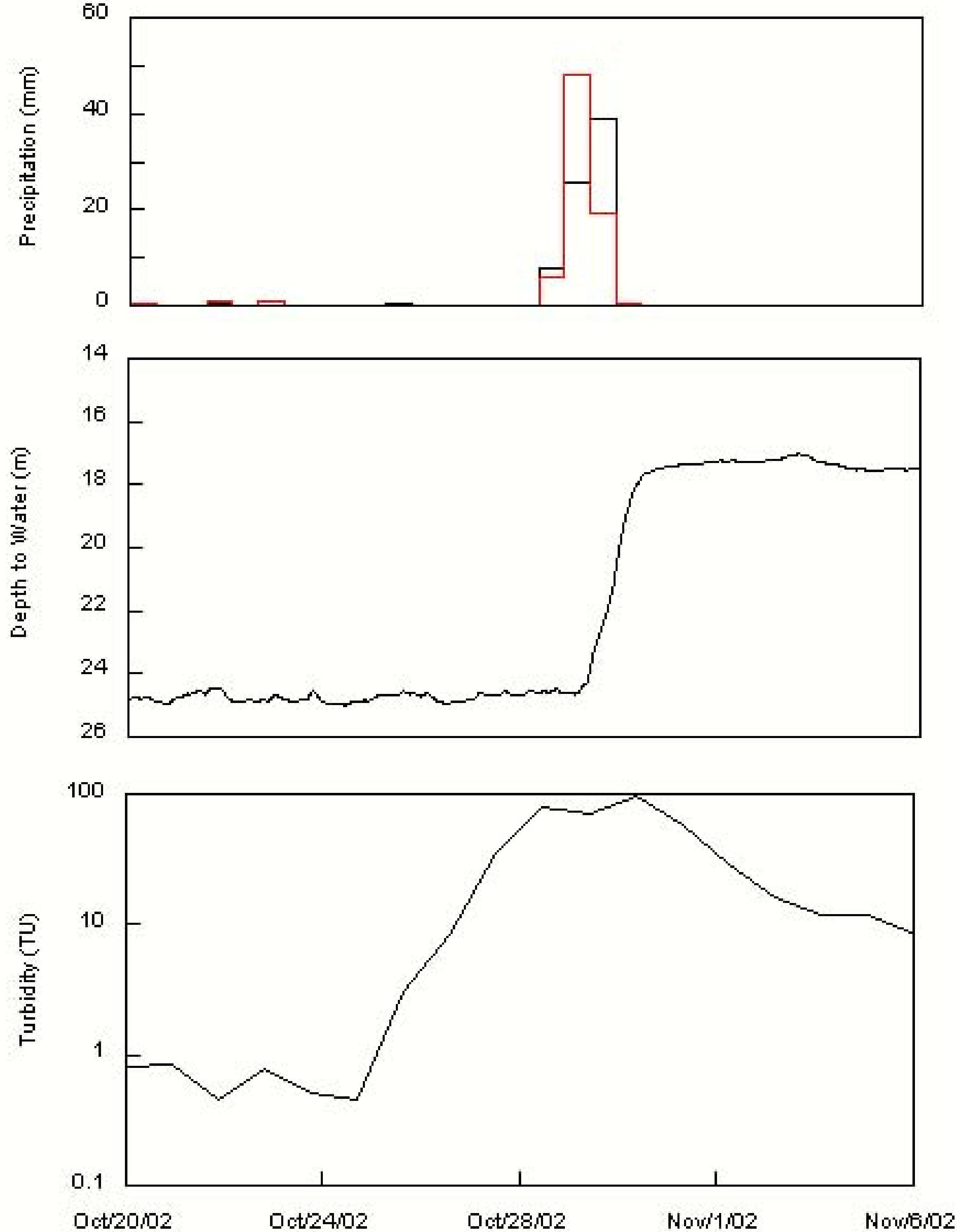


Figure 72. Precipitation, depth to water in the monitoring well, and turbidity and discharge at Crystal spring for the month of March 2003. Precipitation data are from the Marion (red) and Florence (black) stream gages on the Cottonwood River.

However, it is more likely that small amounts of recharge moved from the filled conduits in the Florence Limestone flowed into the diffuse-flow part of the aquifer toward the monitoring well.

The second example contrasts with the first in that there was significant antecedent moisture in the soil and the storms were of higher intensity. These conditions produced runoff and flow in Martin Creek. When the stream began to flow throughout its reach across the Barneston outcrop is unknown and measurements of turbidity were taken only twice each day, once in the morning and once in the late afternoon (Funk, S., personal communication, 2003). However, it appears that the monitoring-well water level began to rise at least a few hours before the increase in turbidity at the spring. This suggests that recharge from Martin Creek quickly filled the conduits in the Florence and the overflow moved into the adjacent diffuse-flow part of the aquifer before the pulse of recharge arrived at the spring. The movement of water away from the conduits and into the diffuse-flow part of the aquifer would have caused a water-level rise in the monitoring well. Figure 72 shows that the monitoring well water level began to level off early on March 21, some time after the turbidity had peaked in the spring discharge and following the end of the period of precipitation.

Spring discharge measurements and turbidograph and monitoring-well hydrograph recessions appear to be related to the intensity and duration of precipitation events. Short recessions follow short periods of precipitation and the more intense runoff-producing rainfall events because a smaller volume of overflow has moved into the diffuse-flow part of the aquifer. Much longer recession periods accompany extended wet periods. The long period of decline in the monitoring-well water level from the wet Spring 2002 period matches the decline in spring discharge and turbidity over the same time period (mid June to late October 2002). By the end of July 2002 flow into Sinkhole 1 from Martin Creek had declined to a trickle of less than 0.06 L/s (0.002 ft^3/s). Thus input to the conduit system from Martin Creek effectively ended during July. However, the water-level recession continued through until the next rainfall event in late October as the secondary porosity in the diffuse-flow part of the aquifer drained back into the conduits.

12.2.3 Hydrochemical evidence

Cl, SO_4, hardness, Sr, and Na concentrations are negatively correlated with spring discharge and are depressed during extended wet periods (Figures 57, 59). Hardness is a function of the Ca and Mg concentrations (Freeze and Cherry, 1979). NO_3, K, and B concentrations fluctuated during the study but do not appear to be negatively correlated with spring discharge and their concentrations did not appear to be affected by wet or dry periods (Figures 60-62). The monthly data collected during this study are insufficient to determine if shorter periods of precipitation and individual storms have the same effect on constituent concentrations as the extended wet periods.

The apparent depression of constituent concentrations and changes in mass ratio during wet periods suggests mixing of the ground water in the limestone aquifer with fresher water, such as Martin Creek surface water or precipitation. Table 5 is a comparison of the Winter and Spring 2002 samples from Crystal Spring, Martin Creek, a stock well in NW, NE, NW Sec. 30, T. 20 S.,

R. 05 E. and a domestic well in NE, NE, SE Sec. 18, T. 20 S., R. 5 E. The two wells are located near Martin Creek in the diffuse-flow part of the aquifer. The Ca/Mg and SO_4/Cl ratios are lower and the concentrations of Na, Sr, and Cl are higher in the Winter and Spring 2002 samples from these wells than in the respective samples from Martin Creek. In most instances, the values of these chemical parameters are intermediate between the two end members.

The chemical composition of the water from the stock well located between Martin Creek and Crystal spring is essentially unchanged between Winter and Spring 2002 sampling events. The potentiometric surface map suggests that the water produced by this well is not from Martin Creek (Figure 35) but rather from a different part of the ground-water flow system. In the domestic well, located northeast and upgradient of Martin Creek, Sr and SO_4 decrease from 21.9 to 11.8 mg/L and from 47.9 to 21.8 mg/L, respectively, between the Winter and Spring 2002 sampling events. Mg and Na also decrease slightly in concentration from 32.6 to 26.8 mg/L and from 17.5 to 15.2 mg/L, respectively. The decreases in constituent concentration between sampling events could have resulted from mixing of ground water with Martin Creek surface water in the diffuse

Table 5. Changes in constituent concentrations and ratios in Winter and Spring 2002 water samples from Crystal spring, Martin Creek, and two wells located in the diffuse-flow part of the Barneston aquifer near adjacent to Martin Creek. All concentrations are in mg/L and constituent concentration ratios are based on mass.

Sample Location/Time of Year Sampled	Ca	Mg	Ca/Mg	HCO_3	Na	Sr	SO_4	Cl	SO_4/Cl
Winter 2002 Crystal spring	90.4	29.1	3.1	404	15.8	5.71	40.9	7.9	5.18
Martin Creek	94.0	24.2	3.9	423	12.1	0.63	15.4	4.9	3.14
Stock well, NW, NE, NW Sec. 30, T. 20 S., R. 05 E.	75.2	32.2	2.3	387	16.2	18.6	55.5	6.0	9.25
Domestic well, NE, NE, SE Sec. 18, T. 20 S., R. 5 E.	81.3	32.6	2.5	406	17.5	21.9	47.9	8.6	47.9
Spring 2002 Crystal spring	77.9	18.3	4.2	337	7.9	1.26	14.7	4.5	3.27
Martin Creek	21.7	4.7	4.6	98.3	2.8	0.09	5.1	3.7	1.38
Stock well, NW, NE, NW Sec. 30, T. 20 S., R. 05 E.	77.5	33.2	2.3	388	16.2	19.3	54.8	5.7	9.61
Domestic well, NE, NE, SE Sec. 18, T. 20 S., R. 5 E.	83.9	26.8	3.1	412	15.2	11.8	20.8	8.2	2.54

flow part of the aquifer. During wet periods when there is flow in Martin Creek, a recharge mound is most likely created near the stream and the hydraulic gradient between the stream and the surrounding aquifer reverses. The extent of the area in the aquifer affected by mounding depends on the amount of recharge entering the aquifer from the streambed (Moench and Kisiel, 1970; Walton, 1985) and the slope of aquifer potentiometric surface and the nature of the hydraulic conductivity anisotropy. The anisotropy will depend on the joint pattern and the degree to which individual joints have been solution enlarged (Domenico and Schwartz, 1992).

12.2.4 Preliminary Descriptive Catchment Conceptual Model

Figure 73 is the schematic for a preliminary descriptive conceptual model of the flow of water through the Crystal spring catchment. No attempt has been made to quantify any of the flows between the boxes because there is insufficient data to do so at this time. This would be inappropriate because the relative importance of the pathways for water moving through the catchment most likely changes depending on precipitation intensity and duration (Smith et al., 1976). The primary features illustrated show the infiltration of water into solution-enlarged fractures and small diameter solution channels from Martin Creek and eventually into the conduits or into the adjacent fissures and solution channels in the rock surrounding the conduits. The arrows between the Large-scale Conduits box and the Joints, Solution-enlarged Joints and Small-scale Solution Channels box indicate that water is exchanged between these two storage compartments. Discharge to Crystal spring is most likely through a larger diameter or master conduit.

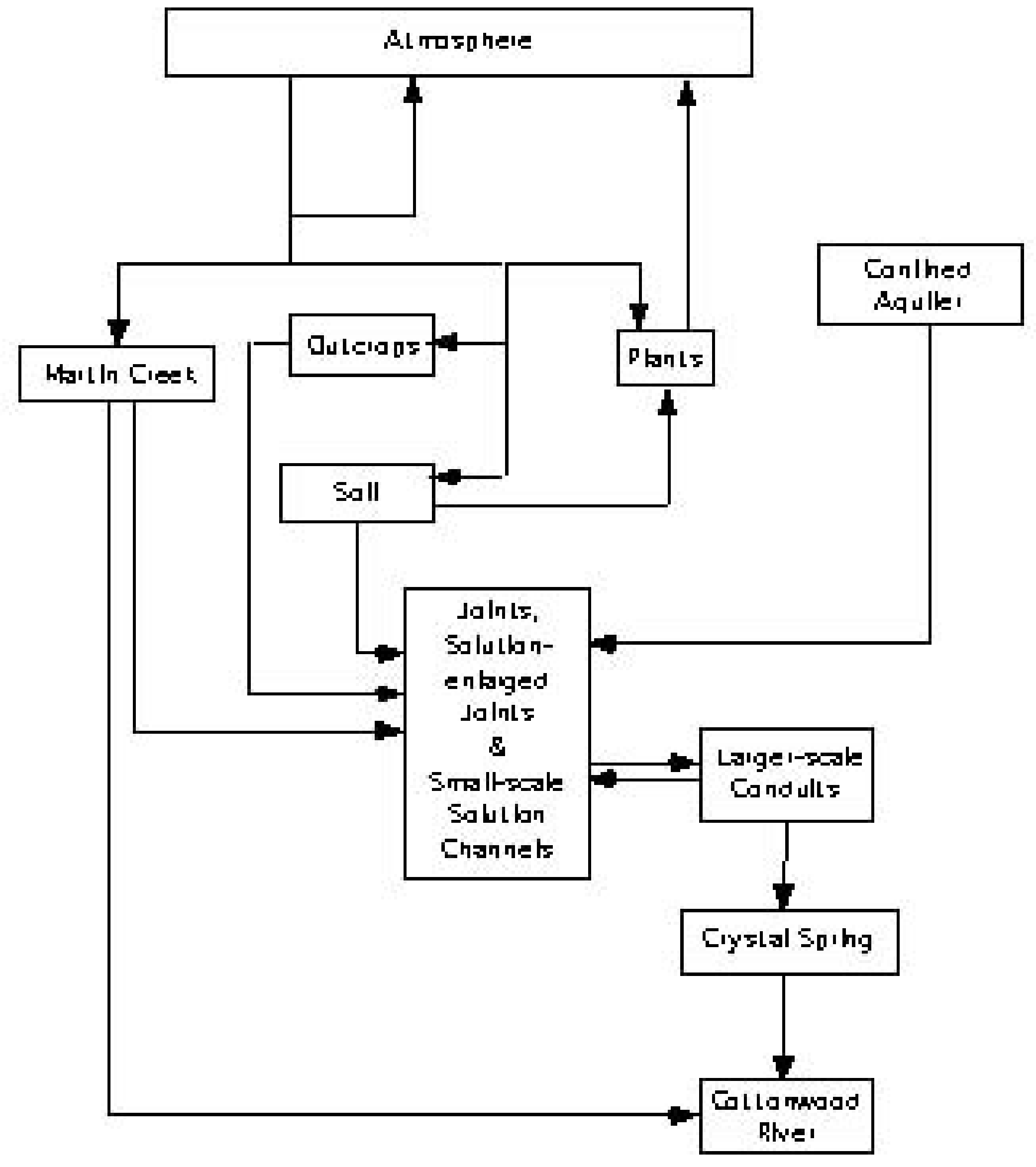


Figure 73. Schematic of a preliminary conceptual model showing the movement of water through the Crystal spring catchment to the Cottonwood River.

13.0 Sources of Recharge to Crystal Spring

13.1 Source contribution fluctuations

Fluctuations in water chemistry, turbidity, and spring discharge in response to wet-dry period transitions indicate that both diffuse and conduit flow sources contribute to the discharge from Crystal spring. Spring discharge during and shortly after wet periods is dominated by conduit flow, but diffuse flow dominates during dry periods following recession from flood events. The diffuse source most likely originates from regional and local ground-water flow through small-scale interconnected secondary porosity in the Barneston aquifer. Diffuse flow seeps into the conduit system and is eventually discharged at the spring. The Barneston aquifer potentiometric surface indicates that this flow originates primarily from areas where the Doyle Shale and other overlying Permian units confine the aquifer. Water/rock chemical interactions are more likely in this part of the aquifer because of the longer time and greater surface area contact of the ground water with soluble carbonate and evaporite minerals in rock framework. Spring discharge is likely to contain a greater loading of dissolved constituents than ground water from conduit sources. The conduit source originates as lower dissolved solids streamflow in Martin Creek that enters the Barneston aquifer through the wider aperture, solution-enlarged joints and bedding planes and the larger diameter conduits in the sinkhole-like features at the top of the exposed limestone streambed. The dye trace experiments demonstrate short travel times between the sinkholes and the spring and thus reveal a direct hydraulic connection through a conduit system. Significant changes in water chemistry between the Winter and Spring 2002 sampling and the behavior of water levels in the monitoring well suggests that dilute ground water in the conduits could be infiltrating into the fine scale secondary porosity features in the aquifer (Martin and Dean, 2001).

The Winter and Spring 2002 Martin Creek samples represent the contribution to the spring from conduit flow sources and the samples from the domestic well in SE, NE, NE Sec. 18, T. 20 S., R. 5 E. are representative of the chemistry of the diffuse flow in Barneston aquifer. Plots of SO_4/Cl ratio vs Cl in the monthly samples from Crystal spring demonstrate the transient impacts of these flow types on spring discharge (Figure 74). Nearly all the scatter in the SO_4/Cl vs. Cl point data from Crystal spring is bracketed by the region on the plot bounded by the Winter and Spring 2002 samples from Martin Creek and the domestic well. This suggests that the Winter and Spring samples are end members of a geochemical mixing process. The Crystal spring SO_4/Cl vs. Cl data points are the fluctuating resultant mixtures of these end members that depend on the relative contribution of water from diffused and focused sources.

13.2 Determination of source contribution using geochemical mixing curves

The observed SO_4 and Cl concentrations in spring discharge depend on the flow rates and concentrations of these constituents in the sources to the spring and the rate of spring discharge. The relationship between these factors and observed SO_4 and Cl concentrations in spring discharge can be expressed as a mass balance equation (Hem, 1985):

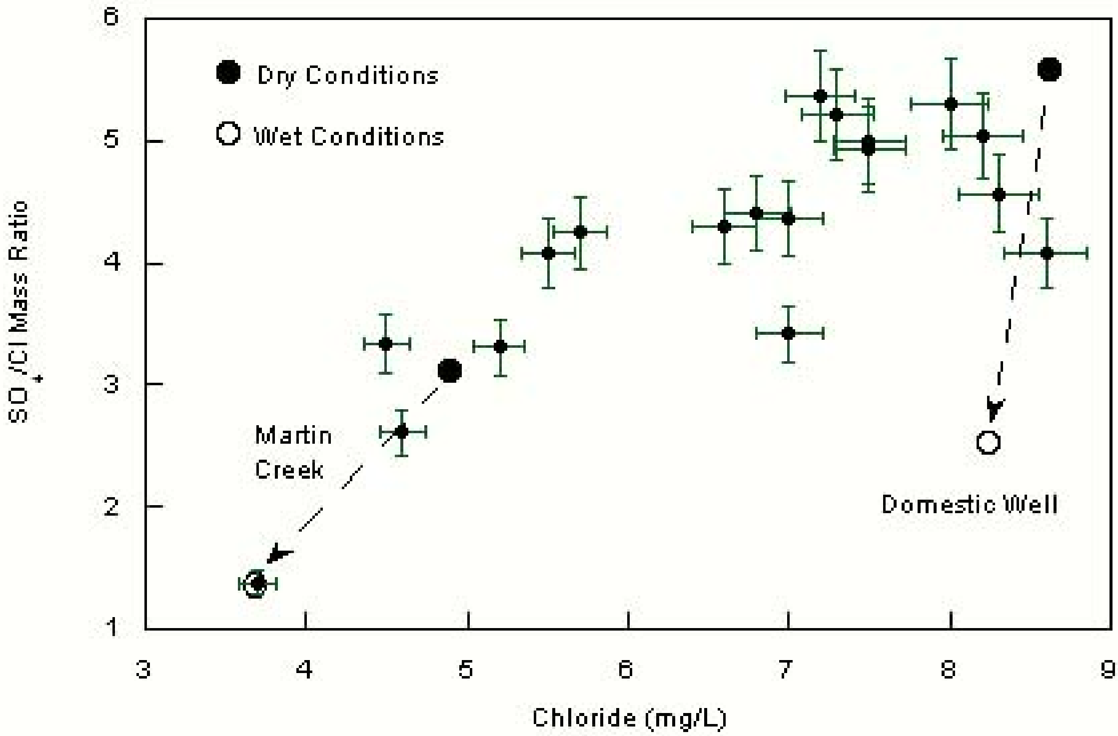


Figure 74. Changes in chloride and the sulfate to chloride mass ratio in the monthly samples of the discharge from Crystal spring with respect to the changes in these parameters in Martin Creek surface water and ground water between the Winter 2002 and Spring 2002 water sampling events.

$$C_1Q_1 + C_2Q_2 = C_3(Q_1 + Q_2) \qquad \text{(Eqn. 6)}$$

where C and Q are the constituent concentrations and the flow rate from sources 1 and 2 and C_3 and $(Q_1 + Q_2)$ are the constituent concentrations and the flow rate from Crystal spring. Application of this simple relationship to derive the wet and dry season mixing curves using the SO_4/Cl ratios and Cl concentrations of the end member waters yields the following relation (D. O. Whittemore personal communication, 2003):

$$SO_4/Cl=[SO_4^1+\{([(SO_4^2-SO_4^1)Cl]-[(SO_4^2-SO_4^1)Cl^1])/(Cl^2-Cl^1)\}]/Cl \quad \text{(Eqn. 7)}$$

Superscripts 1 and 2 refer to the Martin Creek and the domestic well samples, respectively, and the SO_4 and Cl in Eqn. 6 without superscripts refer to SO_4/Cl ratios and Cl concentrations in Crystal spring discharge along the continuum between the Martin Creek and domestic well sample end members.

Application of Eqn. 7 using the Winter and Spring 2002 samples from Martin Creek and the domestic well as end members yielded two mixing curves between the end member wet season samples and the end member dry season samples from both sources (Figure 75). The extended Winter season curve beyond the Martin Creek Winter sample to lower SO_4/Cl ratio and Cl values comes very close to the Spring Martin Creek sample which suggests that this sample is an end member for both mixing curves. Also plotted on the graph are the SO_4/Cl ratios and Cl concentrations for the samples collected from Crystal spring during this project under varying climatic conditions. Most of the data points plot between the wet and dry season mixing curves. This indicates that most of the flow from Crystal spring is derived from the mixing of surface water from Martin Creek and ground water from sources upgradient of Martin Creek in the Barneston aquifer in the direction of the domestic well end member. Those samples that plot above the dry season mixing curve are most likely mixtures that contain small amounts of water from other areas of the aquifer.

13.3 Relative source contribution to Crystal spring discharge

If the flow rates are redefined relative to the discharge from Crystal spring, the relative discharge from each contributing source can be computed using the relation:

$$Q_1 + Q_2 = 1. \qquad \text{(Eqn. 8)}$$

Rearranging the terms to solve for the relative contribution from source 2 and substituting into Eqn. 5 yields:

$$C_1Q_1 + C_2(1 - Q_1) = C_3 \qquad \text{(Eqn. 9)}$$

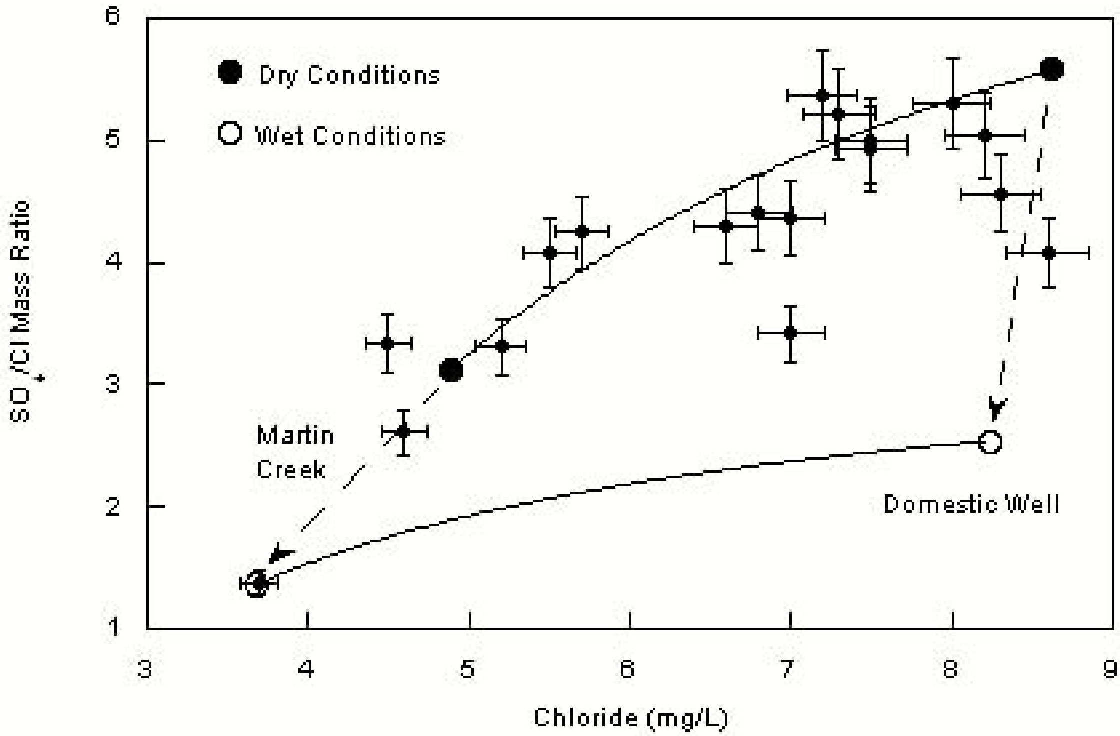


Figure 75. Geochemical mixing curves showing the mixing of surface water from Martin Creek with ground water from the Barneston aquifer in the discharge from Crystal spring using the chloride and sulfate to chloride mass ratios in the monthly samples with error bars.

Using this relation, the relative contribution, Q_1, from Martin Creek was calculated using the Cl concentrations and plotted as a function of spring discharge (Figure 76). The relative contribution of Martin Creek inflow to spring discharge ranges from less than 1% at the lowest flow rates up to 100% at the highest flow rates. Similar results have been reported for a sink-rise system in the Santa Fe River in northern Florida (Martin and Dean, 2001). The plot shows that input to the Barneston aquifer from Martin Creek dominates when spring discharge (>50%) exceeds 124.6 L/s (4.4 ft^3/s). Spring discharge turbidity spikes generally occur at discharge values greater than approximately 85 L/s (3 ft^3/s) (Figures 63, 64) and are usually associated with Martin Creek streamflow generation and aquifer recharge. The correlation of turbidity in Crystal spring discharge with these streamflow events is to be expected given the turbid streamflow in Martin Creek observed frequently during these recharge events and the lack of turbidity in the streamflow input to Sinkhole 1 in between these events.

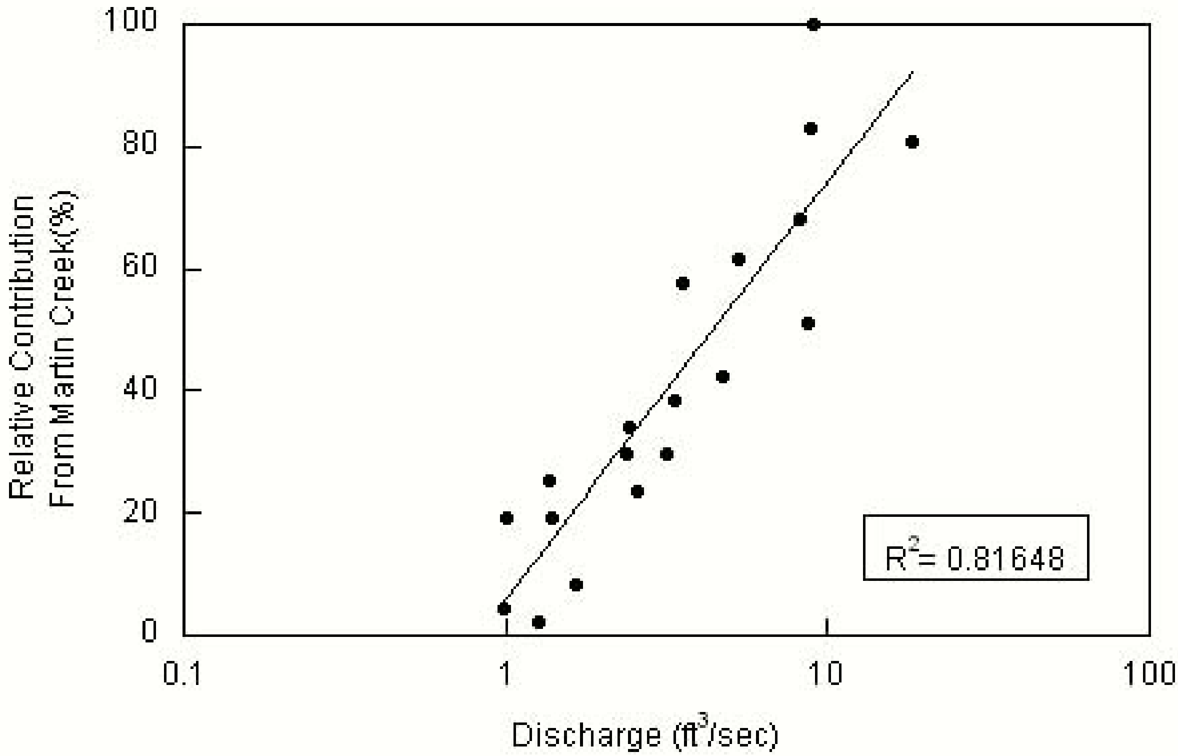


Figure 76. Scatterplot showing the percent of water from Martin Creek in the discharge from Crystal spring estimated from the chloride concentrations in spring discharge.

14.0 Additional Research Needs

With the completion of this study of Crystal spring and its catchment, several research problems remain for exploration by others in order to develop the details of the karst aquifer system and its relationship to the spring.

Carbonates that have been affected by dissolution are highly heterogeneous with respect to their permeability and storage properties. The observations made in this study at the Sunflower rock quarry and elsewhere indicate that the effects of dissolution are nonuniform vertically within the Barneston Limestone. These effects range from solution-widened joints and bedding planes to conduits up to more than 1 m (3 ft) in diameter. The diffuse part of the flow system develops in carbonates dominated by smaller-scale dissolution features, whereas the conduit-flow system develops where the larger-scale features dominate. An improved characterization of these properties should focus on *in situ* packer testing of the diffuse flow part of the aquifer and possibly dye tracing experiments in the sinkholes and using wells to determine differences in travel times between the conduits and the spring. Packer and other well testing methodologies could provide information on the variation in hydraulic conductivity laterally (dip and strike directions) and vertically. Further dye testing under a variety of flow conditions using continuous sampling could provide more insight into the hydraulics of the conduit system between the sinkholes and the spring.

Further research needs to focus on the quantification of flows through the conduits and diffuse-flow systems in the Barneston in response to storm events of varying magnitude. To achieve this end will require continuous monitoring of the stream to determine the occurrence and magnitude of streamflow events in Martin Creek coupled with continuous monitoring of (1) water levels in observation wells adjacent to the stream, (2) discharge and turbidity at Crystal spring, (3) weather conditions and (4) atmospheric pressure. The monitoring wells should be targeted to monitor discrete intervals of the Barneston and not completed as an open borehole. The data collected from this effort could then be used to develop analytical models relating storm events to response of the ground-water system monitored at the spring. These models could be used to predict spring discharge and turbidity depending on intensity and duration of precipitation events during wet periods.

Along this line, discharge water quality and turbidity are concerns because the spring is the sole source for the city's public water supply system. Thus there is a need to treat the water to remove suspended solids that may contain pathogens and to ensure that the treated water meets the water quality standards for drinking water. This study demonstrated a relationship between streamflow events in Martin Creek, abrupt increases in the discharge and turbidity and decreases in most dissolved constituents in spring discharge, and water-level changes in the aquifer at the monitoring well near Martin Creek. Further exploration of these relationships could result in an improved understanding of the spring and aquifer response to storm events. Only a preliminary descriptive model of the catchment could be generated from the data collected during this study.

Crystal spring is located within 100 m (325 ft) of a small drainage that has cut down through most of the Ft. Riley Limestone and has breached the north valley wall of the Cottonwood River (Figure 77). For most of the year the flow in the western branch of this drainage is maintained by spring discharge from the Towanda Limestone Member of the Doyle Shale. The other branch of the drainage trends in a northeast-southwest direction in the uplands north of the spring. This part of the drainage is dry year-round and appears to be up gradient of the spring. It seems likely that much of the lower part of the unnamed drainage could be contributing some recharge to the Barneston aquifer because the stream flows over the exposed Ft. Riley bedrock and the proximity of this area to the spring. Further research is needed to investigate this is as a possible recharge area for the Barneston aquifer and a source for the discharge from Crystal spring.

Sources of SO_4, Cl, NO_3 and other dissolved constituents in spring discharge were not determined from this study. Possible sources are ground-water flow from the diffuse flow system in the outcrop belt or from areas where the Barneston is covered by younger Permian bedrock units. Gypsum and halite have been identified in core samples of the Barneston and overlying units downdip of its outcrop belt (Twiss, 1991). Sulfate and Cl may also be contributed from the soil zone and the upper bedrock where they may have accumulated from evaporation of infiltrating water from precipitation (Macpherson, 1996). Further insight into the sources of these constituents in spring discharge could be gained using the nitrogen-15 and sulfur isotopes in water samples from the spring and the wells in the study area.

Currently, the demand for water from Crystal spring and its catchment is low relative to availability. During this study spring discharge ranged from less than 28.3 L/s (1 ft^3/s) up to more than 510 L/s (18 ft^3/s) or less than 450 gallons per minute up to more than 8,078 gallons per minute). Pumpage into the public water supply system is intermittent at less than 12.6 L/s (200 gallons per minute). This may change with the possibility that a bottled-water plant may locate in Florence and use Crystal spring as its source of supply. The rates of supply needed by this new industrial development are unknown at present. With the potential of adding another high-volume user of discharge from the spring, long term monitoring of spring discharge should be contemplated in order to conduct a rigorous flow-duration analysis. The results would be beneficial to the planning process for the facility and city over the long term.

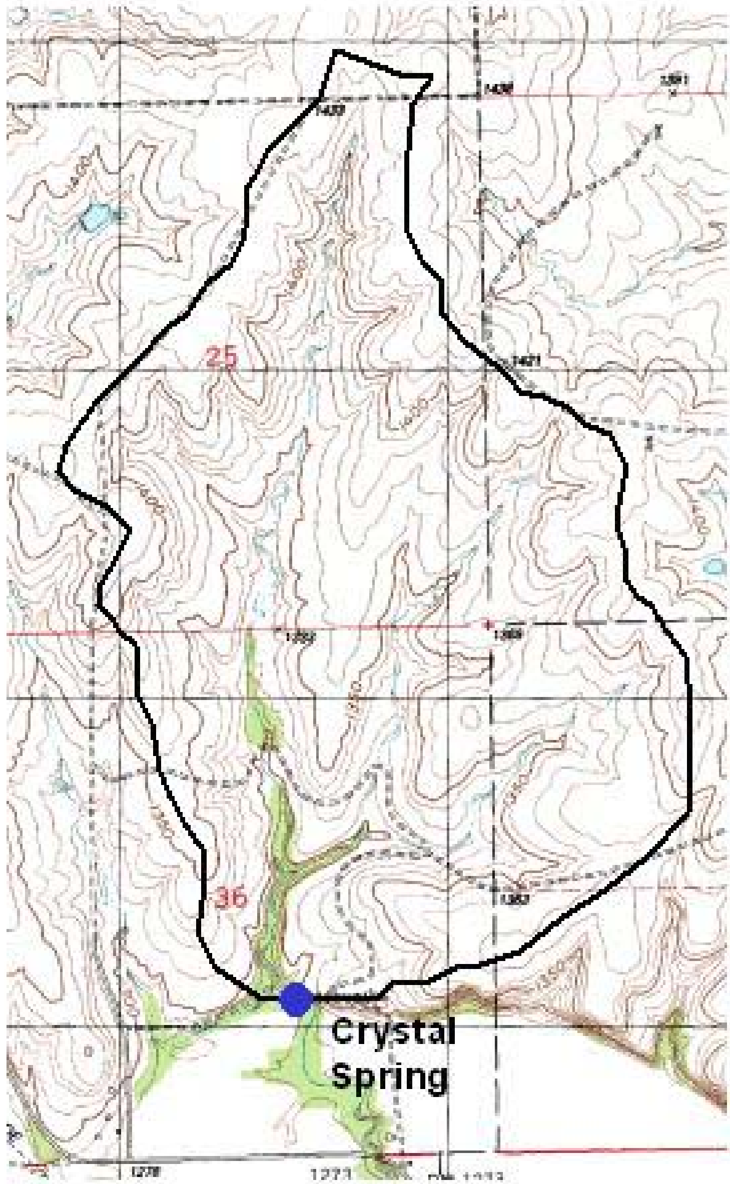


Figure 77. Extent of the unnamed drainage west of Crystal spring. The lower part of the drainage crosses the outcrop of the lower Ft. Riley Limestone and may contribute to the discharge from the spring.

Part III: Delineation of the Source Water Assessment Area

15.0 The Crystal Spring Source Water Assessment Area

15.1 The Source Water Assessment Area for Crystal spring

This study has identified Martin Creek as a major contributor of recharge to Crystal spring where it crosses the Barneston aquifer outcrop. Recharge enters through solution-enlarged joints when there is sufficient run-off to produce stream flow and increases conduit-flow. Contiguous zones on either side of this reach of the stream also provide recharge to the spring from two sources. Recharge is provided by water infiltrating through the soil and outcrops to the water table and this part of the aquifer functions as a temporary storage in the diffuse-flow part of the ground-water system following the input from wet periods or run-off producing storm events. The areal extent of these contiguous zones and the stream reach is defined by where the Barneston Limestone is at the surface in the Martin Creek drainage and is believed to contain a saturated thickness (SWAA 1, Figure 78). The reach of Martin Creek upstream of the Barneston Limestone outcrop belt is also included as a source of water to the spring because the sinkhole-like features in the outcrop belt intercept flow from upstream areas.

A second source area includes the Barneston Limestone outcrop in the unnamed drainage north of Crystal spring (Figure 77). The water infiltrating through the thin, rocky soils and outcrops and into solution-enlarged joints provides recharge. Because of the proximity of this area to the spring, it is possible that some of this recharge is contributed to Crystal spring discharge. This area is identified as SWAA 2 (Figure 78).

15.2 Factors influencing the delineation

The factors considered in defining the Source Water Assessment Area (SWAA) for Crystal spring include the following: soils, adsorption potential of the aquifer materials, direct vertical pathways to the ground-water system, overlying confining layers, and time-of-travel from sources of recharge to the spring. These factors have been extensively discussed in Part 2 of this report and only a summary of the discussion will be presented here as it relates to defining the SWAA.

Soils: Soils within the SWAAs belong to the Labette-Tully-Sogn association, which consists of well drained to somewhat excessively drained, moderately thick to thin (<1 m) soils with clayey or silty subsoil. Chert and limestone gravel lenses are common near the bedrock surface. In many areas of both SWAAs a soil cover has been removed by erosion and the jointed bedrock is at the surface. In the reach of Martin Creek that spans the Barneston Limestone outcrop belt, the stream flows over a bedrock surface or over a thin poorly sorted layer of silt and limestone rubble overlying the bedrock surface. The relatively high soil permeability and thin to absent soils suggest that contaminants could potentially move quickly into solution-enlarged joints in the Barneston Limestone outcrop belt, especially in the stream bottom of Martin Creek.

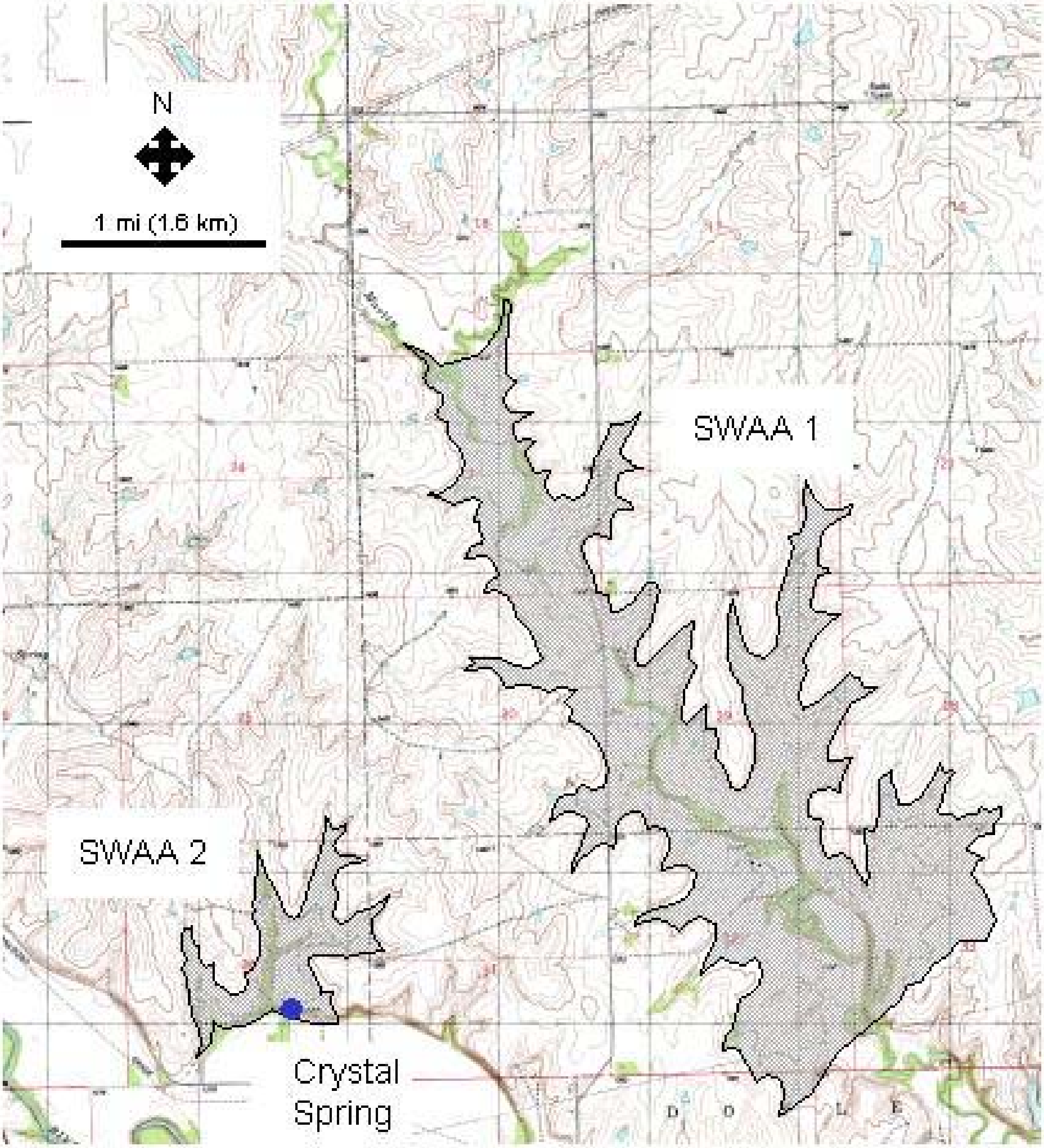


Figure 78. The Source Water Assessment Area (SWAA) for Crystal spring shown in a gray screen. The extent of the two SWAAs is defined by the outcropping Doyle Shale-Barneston Limestone contact and the approximate extent of the significant saturated thickness in the aquifer zones of the Barneston Limestone beneath Martin Creek.

Adsorption potential of the aquifer materials: The primary contaminants of concern in the Crystal spring catchment that are most likely to be of concern are bacteria, viruses, nitrate, agrichemicals. The Barneston Limestone consists primarily of limestone and chert, which have a low potential for the adsorption of these contaminants. The silt fraction stored in the conduits consists primarily of quartz, mica, and feldspar. These constituents have a low adsorption potential.

The dye trace experiments confirm that the residence time of ground water in the conduits is less than 3 days under low spring discharge conditions. Ground-water residence time in the diffuse-flow part of the outcrop belt of the Barneston aquifer is unknown but is expected to increase with distance from the conduits. This indicates that the short residence time and lower surface area available for rock/water chemical interactions is short within and near the conduits. The longer residence time of ground water in the diffuse-flow part of the aquifer may increase the opportunities for rock/water chemical interactions. Thus, the adsorption potential of the aquifer near the conduits is very low.

Direct vertical pathways: Direct vertical pathways to the ground-water system in the Barneston outcrop belt include the solution-widened joints in stream bottom of Martin Creek and in the upland areas where erosion has removed the soil cover on the bedrock surface. Much of the stream bottom and the upland outcropping bedrock surfaces are a pavement of weathered limestone blocks separated by joints. Joint aperture widths range up to several centimeters. In the uplands, it is apparent that the joints are eroded-soil or limestone-residuum filled. In the stream bottom, the solution-widened joints at the bottom of the sinkhole-like features are open. Elsewhere, the joints are filled with fine sediment or appear to be open for some distance below the stream bottom.

Overlying confining layers: Confining layers overlying the Barneston aquifer consist of interbedded shale and limestone. The shale and limestone units form aquitard and aquifer units, respectively because of the permeability contrast between the two lithologies. Where these overlying units are present, it is likely that recharge to the Barneston aquifer from infiltrating precipitation is low.

Time-of-travel from sources of recharge to the spring: Time of travel through the conduit-flow part of the ground-water system between the sinkhole-like features in Martin Creek and Crystal spring is 2-3 days at low spring discharge. The time of travel should be much lower following wet periods and runoff-producing storm events because of the increase in hydraulic gradient in the conduits between the stream and the spring. Time of travel through the diffuse-flow part of the ground-water system to Crystal spring from the aquifer near Martin Creek was not estimated, but it is expected to be longer because of the lower ground-water velocities in this part of the flow system.

In the area surrounding Crystal spring and north of the Cottonwood River valley wall, the Ft. Riley Limestone is at the surface and there is a small, unnamed southward drainage approximately 100 m west of the spring (Figure 75). Where it has breached valley wall, the stream has cut down through most of the Ft. Riley. The soil cover in most of the lower part of the drainage is very thin to absent. The base flow of the main branch of the stream is fed by spring discharge from the Towanda Limestone Member of the Doyle Shale. The other main branch appears to be dry most of the time. A reconnaissance of the drainage suggests that the lower part of this drainage could possibly be a recharge area for the Barneston and provide a source of discharge to Crystal spring. Because of its close proximity to the spring, it is likely that recharge from this source area could arrive at the spring in a relatively short period of time.

15.3 Delineation based on apparent sensitivity of the aquifer to contamination

Two subareas can be defined with the Crystal spring SWAAs based on sensitivity of the aquifer to contamination. Delineation is based on time-of-travel of contaminants from a source at the surface to Crystal spring. On this basis the most highly sensitive area is the stream bottom of Martin Creek. SWAA 2 could also be considered a highly sensitive area because of its proximity to Crystal spring. Assuming that contaminants would take the shortest route to the spring, ground-water velocities were determined to be approximately 1,500 m/day (0.93 mi/day) when spring discharge was approximately 56.6 L/s (2 ft^3/s). No dye-trace tests were conducted in the diffuse-flow part of the ground-water system to verify flow paths and estimate travel times between the monitoring well and the spring. However, it is believed that the travel times would be much longer because of the slower ground-water velocities in this part of the ground-water system.

16.0 Summary

The goals of this project were to conduct a hydrogeologic investigation Crystal spring and its catchment and to use the results of the investigation to delineate the spring's SWAA. For the most part these goals have been accomplished although more research is needed to understand the movement of water through the catchment and the sources of the dissolved constituents in the discharge from the spring. Very little is known of the Flint Hills regional hydrogeology. This study has provided more detailed information on a very small part of that regional system, but much more remains to be done. The following is a summary of the project results.

The shallow subsurface geology consists of a sequence of thick to thin alternating Permian limestone and shale units that belong to the Chase Group. Because of the permeability contrasts across lithologic boundaries, the limestones behave as aquifer units and the shales form the confining units. The thickest of these limestone aquifers is the Barneston. Stratigraphically, the Barneston Limestone consists of the Ft. Riley Limestone, the Oketo Shale, and the Florence Limestone members from youngest to oldest.

Field observation of the Barneston Limestone in outcrop and at the Sunflower rock quarry and examination of samples of the cuttings from the monitoring well drilling indicate development of secondary porosity, primarily from solution-enlargement of joints and fractures and secondarily from the dissolution of gypsum and halite. The secondary porosity developed in the Ft. Riley Limestone differs somewhat from that developed in the Florence Limestone. Vuggy zones near the top of the unit, solution-enlarged joints, and small-diameter conduits characterize Ft. Riley secondary porosity whereas vuggy zones from evaporite dissolution near the bottom and rare larger-diameter conduits near its top characterize most of the secondary porosity in the Florence. A 1.3-m (4.2-ft) diameter conduit in the upper Florence was found in the north wall of the Sunflower rock quarry. The conduit was half full of stratified silt that had washed into it from overlying sources over time. Solution-enlarged joints were noted in the Florence exposed in the quarry but they did not seem to be as common as those that were observed in the Ft. Riley. Where Martin Creek crosses the Ft. Riley outcrop belt solution-enlarged joints in the streambed with aperture widths up to 10 cm (4 in) provide entry points for flow into the Barneston aquifer at the bottom of sinkhole-like features. Discharge from Crystal spring comes from near the top of the Florence Limestone Member of the Barneston Limestone. The point of discharge from the bedrock could not be observed because of the springhouse location atop the spring.

According to the Meinzer (1923) classification of springs based on discharge, Crystal spring is a second magnitude spring during the wet spring season when discharge is at its highest and a third or fourth magnitude spring during extended dry periods. During this study, measured spring discharge ranged from slightly less than 28.3 L/s up to more than 510 L/s (1 ft^3/s up to more than 18 ft^3/s) and at least half of the time the discharge was 73.6 L/s (2.6 ft^3/s).

Water-level measurements in wells within the study area and the dye trace experiments confirm that the sinkhole-like features in the stream bottom of Martin Creek are directly connected to Crystal Spring, 4.8 km (3 mi) to the south. The Barneston aquifer potentiometric surface indicates generally southward ground-water flow from the upland area to the Cottonwood River valley. Two sets of dye trace experiments were conducted using the sinkholes in Martin Creek as injection points with monitoring at Crystal spring. Both sets of experiments were conducted during dry periods when spring discharge was less than 71 L/s (2.5 ft^3/s). The amount of time for the injected dyes to travel from the sinkhole-like features to the spring was at most 2.5 days. Plans were made to conduct a similar set of tests using the monitoring well near Martin Creek, but were never acted on due to time limitations.

Precipitation duration and intensity within the catchment directly influence Crystal spring discharge, turbidity, and water quality. Extended wet periods cause discharge and turbidity to rise abruptly and result in lower concentrations of dissolved constituents in the discharge. Depending on the duration and intensity, shorter wet periods and storm events may cause momentary fluctuations in discharge, turbidity, and water chemistry. During extended dry periods discharge and turbidity are low but the dissolved constituent concentrations rise up to levels close to those found in well-water samples. Nitrate, B, and K appear to be the exception to this overall pattern. Concentrations of these constituents do not seem to be related to discharge or to temporal variations in precipitation. Plots of the SO_4/Cl ratio vs. Cl show that conduit- and diffuse-flows in the Barneston Limestone contributes to spring discharge. Analysis of the hydrograph from the monitoring well near Martin Creek shows that fluctuations in spring discharge and turbidity are correlated with water-level fluctuations in the diffuse-flow part of the Barneston aquifer. The well hydrograph also reveals that the diffuse-flow part of the aquifer provides temporary storage when the volume of recharge from streamflow input exceeds the capacity of the adjacent conduits. Following wet periods, streamflow input wanes and is eventually shut off. In the conduits, the hydraulic gradient declines over time and the stored water moves from the diffuse-flow part of the aquifer to the conduits. This is similar to the creation and depletion of bank storage in stream-aquifer systems in response to flood-wave passage.

Two SWAAs were delineated based on the results of this study: one associated with the reach of Martin Creek that traverses the Barneston Limestone outcrop belt and the other associated with the Barneston aquifer outcrop where the spring is located. SWAA 1 includes the Barneston outcrop belt near Martin Creek bounded on the southeast by the estimated limit of saturation in the aquifer. Most of the attention of this study has focused on the Martin Creek SWAA (SWAA 1). The results of the field investigation indicate that this area is an important source of recharge to Crystal spring. The contribution of recharge to the Barneston aquifer outcrop area in the unnamed drainage and its consequent discharge at Crystal spring is unclear. Conservatively, it would seem reasonable to conclude that this part of the outcrop belt is a source area because of its proximity to the spring and the generally southward direction of ground-water flow in the Barneston. However, it is possible that the local ground-water flow system in this part of the drainage is entirely separate from the flow system that involves the Crystal spring and SWAA 1. Additional research needs to be carried out in this area to determine the level of significance of recharge from SWAA 2 to Crystal spring. The factors considered in defining the Source Water Assessment Area (SWAA) for Crystal spring include the following: soils, adsorption potential of the aquifer materials, direct vertical pathways to the ground-water system, overlying confining layers, and time-of-travel from sources of recharge to the spring.

ACKNOWLEDGMENTS

This project was funded by grants from Nonpoint Source Section, Bureau of Water, Kansas Department of Health and Environment (KDHE) as project NPS-2K-014. The author is grateful to Don Snethan and Rob Beilfuss of KDHE for their assistance and oversight in this project. Many other individuals contributed to the success of this project. I wish to acknowledge the contributions of my colleagues at the Kansas Geological Survey in this project. Margaret Townsend, KGS, provided her library and her experience on karst hydrogeology, assisted with water sample collection, and arranged for nitrogen isotopic analyses of the water samples. The Analytical Services Section of KGS conducted the chemical analyses of water samples and the x-ray diffraction of the suspended solids in a sample from Crystal spring. John Healey, KGS, provided assistance in the field during the dye-trace experiments. Former members of the Survey also contributed, including Sharon Vaughn and Marcia Schulmeister. Marios Sophocleous reviewed the draft report and provided comments that improved its overall content. I wish to thank YSI, Inc, for the loan of their rhodamine-WT sensor and monitoring device for the dye-trace experiments. Dr. Steve Macko, University of Virginia, performed the N-15 analyses on water samples collected for this project. Dr. Steve Randtke, University of Kansas, analyzed water samples for dissolved organic carbon. Lastly, the assistance of Steve Hett and his family, and Stewart Funk and other city of Florence personnel are gratefully acknowledged. The Hett family allowed me entrance onto their property to conduct much of this research and provided valuable insights from their experience. Stewart Funk took many of the spring discharge measurements during the early part of the study when I could not. He also assisted with the dye-trace experiments, and provided turbidity data that he had collected as part of his duties for the city during the study.

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Appendix 1. Results of the chemical analysis of the monthly samples collected from Crystal spring.

Sample Collection Date	Specific Cond (μS/cm)	SiO_2 (ppm)	Ca (ppm)	Mg (ppm)	Na (ppm)	K (ppm)	Sr (ppm)	HCO_3 (ppm)	SO_4 (ppm)	Cl (ppm)	F (ppm)	NO_3 (ppm)	B (ppb)
1/4/02	680	16.9	90.4	29.1	15.8	1.3	5.71						71
2/22/02	670	15.9	89.1	29.2	14.9	1.0	5.50						30
3/20/02	673	15.8	87.4	29.1	15.0	1.1	5.58					3.2	48
4/6/02	670	16.2	83.5	28.5	13.1	0.8	5.90						61
4/9/02	670	16.5	85.3	28.8	14.3	0.9	5.89	392	42.4	8.0		3.5	34
4/30/02	584	15.3	79.5	22.3	11.2	2.0	3.35	345	30.0	6.8		4.8	48
5/31/02	530								15	4.5		3.4	
6/10/02	542	14.6	77.9	18.3	7.9	1.9	1.26	337	14.7	4.5	0.19	2.0	<26
7/12/02	618								24.2	5.7		3.6	
8/21/02	640								33.1	6.6		3.6	
9/12/02	667								37.0	7.5		3.5	
10/17/02	673								37.4	7.5		3.8	
11/15/02	649								38.6	7.2		4.0	
12/16/02	670								37.9	8.3		3.2	
1/23/03	679								41.3	8.2		3.1	
3/3/03	671								42.2	8.5			
3/21/03	439								12	4.6		9	
4/17/03	610								23.9	7.0		4.0	
5/15/03	576								18.3	6.0		3.7	
6/11/03	607								21.9	6.4		3.8	

Appendix 2. Results of the chemical analysis of the surface and ground water samples collected from the project area as part of the Winter 2002 and Spring 2002 sampling events. Specific conductance units are μS/cm. pH is in pH units. Boron (B) concentrations are given in units of parts per billion (ppb). Dissolved organic carbon (DOC) concentrations are given in mg/L. All other concentrations are given as parts per million (ppm).

Sample Location	Date	Source	Spec. Cond.	pH	SiO_2	Ca	Mg	Na	K	Sr	HCO_3	SO_4	Cl	F	NO_3	B	DOC
Winter 2002 Sampling																	
NW, NE, NW Sec. 30, T. 20 S., R. 5 E.	01/02/2002	Windmill	680	7.35	16.2	75.2	32.2	16.2	1.1	18.6	387	55.5	6.0	1.05	1.7	92	0.38
SE, NE, NE Sec. 18, T. 20 S., R. 5 E.	01/02/2002	Domestic Well	702	7.45	19.7	81.3	32.6	17.5	1.0	21.9	406	47.9	8.6	0.75	3.6	66	0.38
01 SW, SE, SW Sec. 25, T. 19 S., R. 5, E.	01/02/2002	Domestic Well	715	7.45	19.2	82.1	38.1	17.5	0.9	5.98	393	56.8	12.3	0.80	6.2	52	0.57
02 SW, SE, SW Sec. 25, T. 19 S., R. 5, E.	01/02/2002	Windmill	738	7.40	14.7	104	26.5	17.6	0.6	0.57	393	50.2	18.4	0.38	13.2	45	
SE, NE, NE Sec. 2, T. 20 S., R. 5 E.	01/02/2002	Battery 2 wells	3650	7.40	13.4	350	174	366	3.1	10.8	218	2010	77.7	0.83	25.4	4441	1.22
NW, NW, NE Sec 31, T. 20 S., R. 5 E.	01/03/2002	Windmill	590	7.95	17.7	60.0	20.9	39.3	0.7	0.92	323	23.4	19.2	0.31	11.9	33	
SE, SE, SE Sec 12, T. 20 S., R. 4 E.	01/03/2002	Stock Well	828	7.55	19.1	94.0	39.2	28.4	1.2	10.4	404	119	13.3	0.67	1.0	114	
SW, NW, SW Sec 18, T. 20 S., R. 5 E.	01/03/2002	Martin Creek	638	8.15	13.4	94.0	24.2	12.1	0.9	0.63	423	15.4	4.9	0.37	1.9	51	1.38
NE, SW, NE, SE Sec. 27, T. 20 S., R. 4 E.	01/03/2002	Domestic Well	679	7.45	16.3	89.1	30.1	10.5	1.0	10.5	354	82.7	5.2	0.87	2.5	60	
SW, NW, NW, SE Sec. 10, T. 20 S., R. 5 E.	01/04/2002	Spring Florence LS	661	7.60	15.6	90.0	24.4	17.7	0.6	1.30	399	22.4	12.5	0.36	7.6	43	
SW, SW, SW, SE Sec. 10, T. 20 S., R. 5 E.	01/04/2002	Spring Florence LS	596	7.60	14.9	91.6	19.6	10.1	0.8	1.54	381	17.4	5.3	0.33	2.4	36	
SW, NW, SE Sec. 36, T. 20 S., R. 4 E.	01/04/2002	Crystal Spring	680	7.70	16.9	90.4	29.1	15.8	1.3	5.71	404	40.9	7.9	0.46	3.2	71	0.46

Sample Location	Date	Source	Spec. Cond.	pH	SiO_2	Ca	Mg	Na	K	Sr	HCO_3	SO_4	Cl	F	NO_3	B	DOC
Spring 2002 Sampling																	
NW, NE, NW Sec. 30, T. 20 S., R. 5 E.	06/10/2002	Windmill	690	7.40	16.4	77.5	33.2	16.2	0.9	19.3	388	54.8	5.7	0.96	1.7	64	0.51
SE, SE, SE Sec 12, T. 20 S., R. 4 E.	06/10/2002	Stock Well	778	7.40	18.6	92.8	32.2	20.2	1.2	7.15	397	58.1	11.2	0.50	32.2	68	0.92
NE, NE, NE Sec. 30, T. 20 S., R. 5 E.	06/10/2002	Stock Well	648	7.45	17.7	87.1	26.1	17.6	1.3	2.54	347	54.7	7.4	0.38	3.5	<15	2.6
SE, NE, NE Sec. 18, T. 20 S., R. 5 E.	06/10/2002	Domestic Well	671	7.60	18.4	83.9	26.8	15.2	0.6	11.8	412	20.8	8.2	0.42	4.8	36	0.45
SE, NW, NW Sec 22, T. 20 S., R. 5 E.	06/10/2002	Windmill	650	7.60	13.3	88.7	22.0	17.3	0.7	2.76	373	43.0	6.5	0.40	3.0	<40	
SW, NW, NW, SE Sec. 10, T. 20 S., R. 5 E.	06/10/2002	Spring, Florence Ls.	570	7.50	12.7	94.8	14.0	9.5	0.8	0.37	353	14.0	4.7	0.29	10.2	<15	
SW, SW, SW, SE Sec. 10, T. 20 S., R. 5 E.	06/10/2002	Spring, Florence Ls.	505	7.60	11.4	80.8	13.8	4.3	1.0	0.28	324	8.5	2.8	0.25	3.1	29	
01 SW, SE, SW Sec. 25, T. 19 S., R. 5, E.	06/10/2002	Domestic Well	740	7.60	18.4	80.8	37.1	17.4	0.9	5.12	392	61.4	14.6	0.64	7.1	59	
02 SW, SE, SW Sec. 25, T. 19 S., R. 5, E.	06/10/2002	Windmill	730	7.60	14.1	103	23.3	15.4	0.7	0.56	389	39.5	19.6	0.30	15.9	27	
SE, NE, NE Sec. 2, T. 20 S., R. 5 E.	06/10/2002	Battery 2 Wells	2245	7.45	14.0	223	91.2	177	3.2	5.93	345	846	63.4	0.54	85.1	2068	
NE, SW, NE, SE Sec. 27, T. 20 S., R. 4 E.	06/10/2002	Domestic Well	690	7.60	16.9	88.3	30.7	10.1	1.0	11.1	355	79.9	5.3	0.81	2.8	<40	
SW, NW, SE Sec. 36, T. 20 S., R. 4 E.	06/10/2002	Crystal Spring	542	7.80	14.6	77.9	18.3	7.9	1.9	1.26	337	14.7	4.5	0.31	3.8	<26	2.49
SW, NW, SW Sec 18, T. 20 S., R. 5 E.	06/10/2002	Martin Creek	175	7.80	14.8	21.7	4.7	2.8	6.1	0.09	98.3	5.1	3.7	0.19	2.0	30	11.81