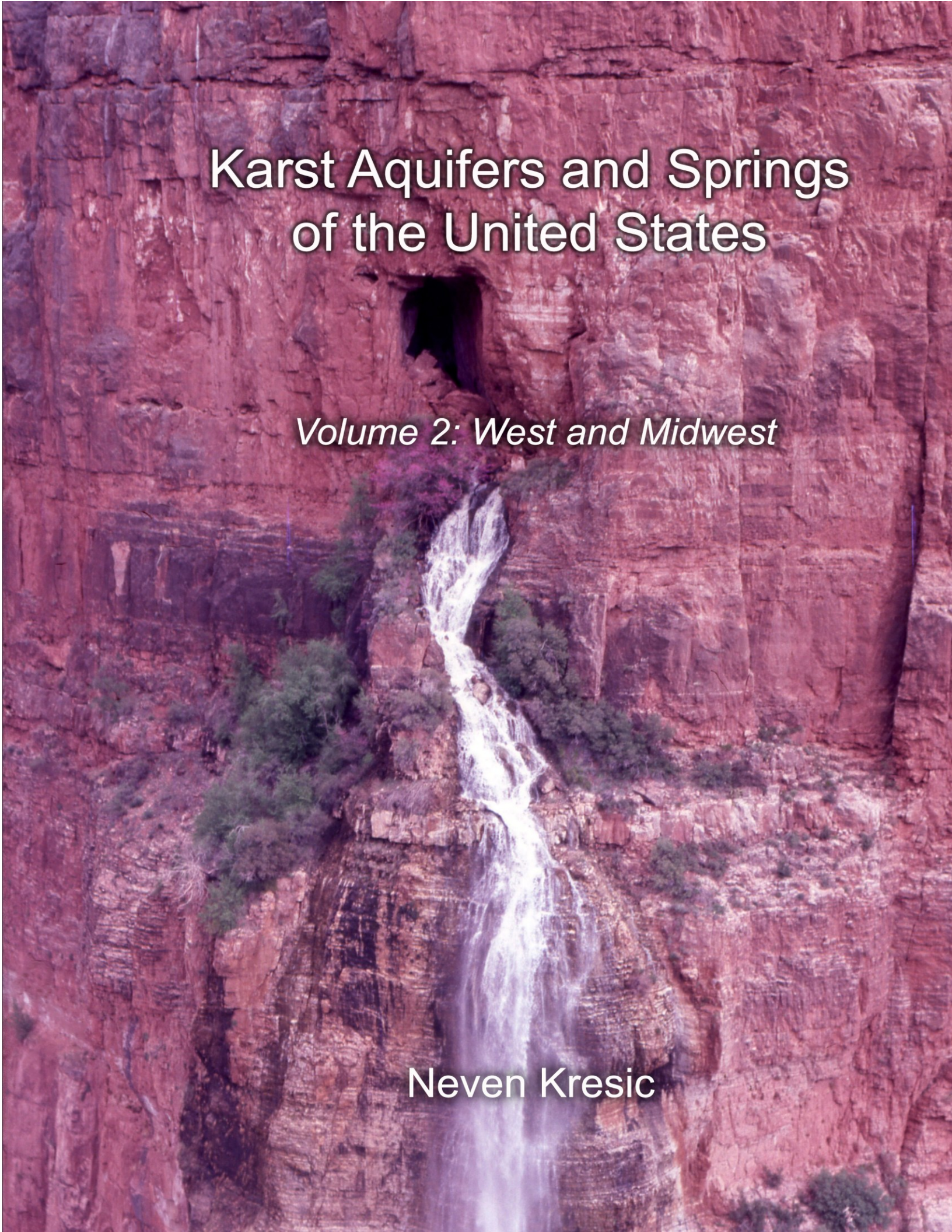
Karst Aquifers and Springs of the United States

Volume 2: West and Midwest



Neven Kresic

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GROUNDWATER ACADEMY

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Front cover: Cheyava Falls Spring in Grand Canyon, Arizona—photo from the USGS Grand Canyon collection of Billingsley et al., 2019.
Back cover: Alley Spring in Missouri (top); and Moth Spring in Minnesota (bottom)—photo courtesy of Martin Larsen.

The author is grateful to the following for hosting the electronic (pdf) version of *Karst Aquifers and Springs of The United States* on their web sites for free download.

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MIKAS
MOST IMPORTANT KARST AQUIFER SPRINGS

MIKAS (Most Important Karst Aquifer Springs) is a project by the Karst Commission (KC) of the International Association of Hydrogeologists (IAH) aimed at creating the first list of the most important karst springs at both the global and national levels, representing the natural heritage of countries with karst.Visit us at https://mikasproject.org/

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About The Author

Neven Kresic is an independent groundwater consultant with more than 40 years of experience focused on groundwater and surface water consulting, research, and teaching. He specializes in characterizing and analyzing complex groundwater systems for water supply and groundwater remediation projects and is an expert in the application of numerical models for designing remedial alternatives including simulation of contaminant fate and transport in groundwater. Over the course of his career, Dr. Kresic served in lead technical roles at major national and international engineering and environmental consulting companies based in the United States.

Dr. Kresic has worked on numerous projects for a variety of U.S. and international clients including federal, state, and local agencies; industries such as water, transportation, and power utilities; and oil, petrochemical, chemical, mining, and construction companies. He was Senior Fulbright Scholar at the United States Geological Survey in Reston, Virginia, and the George Washington University, Washington, D.C., where he conducted novel research on characterizing and modeling groundwater flow and contaminant fate and transport in fractured rock and karst aquifers. Prior to coming to the United States in 1991, Dr. Kresic was professor of groundwater dynamics at the University of Belgrade, former Yugoslavia.

Dr. Kresic taught academic courses and professional workshops in hydrogeology, groundwater resources, groundwater modeling, and groundwater remediation at universities, government agencies, and conferences around the World. He authored and co-authored numerous papers and eight books on the topic of groundwater. Dr. Kresic is a founding member of the Ground Water Modeling Interest Group sponsored by the National Ground Water Association, and former Co-chair of the Karst Commission of the International Association of Hydrogeologists. He is past Vice President for International Affairs of the American Institute of Hydrology and was committee member of the Groundwater Management and Remediation Specialty Group of the International Water Association, and Chair of the Board of Directors, Karst Waters Institute (KWI).

Through Groundwater Academy, Dr. Kresic and his collaborators support the protection and sustainable management of groundwater resources and springs around the world.

TABLE OF CONTENTS

Introduction 1

Chapter 1 Water in Karst 10

1.1 Karst and Karstification 10

1.1.1 Sinkholes 30

1.2 Types and Characteristics of Karst Springs 34

1.2.1 Spring vs. River Rise 48

1.3 Example of Public Health and Environmental Catastrophe 57

1.4 Web Resources 66

Chapter 2 Karst Aquifers of the United States: West and Midwest 67

2.1 The Ozarks Plateaus Aquifers 70

2.2 Edwards-Trinity and Edwards Carbonate Aquifers 74

2.3 Madison Limestone Aquifer 85

2.4 Basin and Range, and Bear River Range Carbonate Aquifers 88

2.5 Colorado Plateaus Karst Aquifers 91

2.6 Roswell Aquifer 95

2.7 Arbuckle-Simpson Aquifer 96

2.8 Northern Midwest Paleozoic Carbonate Aquifers 100

2.8.1 Upper Carbonate Aquifer (Paleozoic) 107

2.9 Marble Karst in California 108

2.10 Pseudo Karst Aquifers 113

Chapter 3 Important Karst Springs of the United States: West and Midwest 120

3.1 Missouri and Arkansas 120

3.2 Texas 171

3.3 Montana, Wyoming, and South Dakota 196

3.4 Utah, and Nevada 231

3.5 Arizona, and Colorado 263

3.6 New Mexico, and Oklahoma 293

3.7 Wisconsin, Minnesota, and Iowa 304

3.8 Illinois, and Michigan 326

References and Select Readings 341

Introduction

Groundwater in karst environments has long intrigued scientists and non-scientists alike due to its many fascinating facets. It feeds the world's largest springs, many of which enabled the establishment of the first urban centers in human history and continues to serve as a reliable source of water supply to the present day; it creates a mysterious and sometimes frightening underground world of caves (Fig. 1); it supports living creatures often unique to specific locations (Fig. 2); it behaves unpredictably as it can sometimes rise tens of feet after heavy rains in the matter of hours (Fig. 3) giving life to numerous temporary springs and increasing the flow of permanent springs 1000-fold.



Figure 1 Arrival of Charon on a boat from a cave believed to be an entrance to Hades, the Greek underworld. Illustration by Gustave Doré in the Pollard and Moss 1885 edition of Dante Alighieri's Inferno, Plate 9, Canto III—Charon.



Figure 2. Endangered and Federally protected Devil's Hole pupfish, one of the world's rarest fish. It lives only in one cave in Nevada, in the Ash Meadows National Wildlife Refuge which is under constant threat from mining interests. Photo by O. Feuebacher, U.S. Fish & Wildlife Service.

Using karst aquifers and springs for water supply was mastered by the Romans who preferred their water to any other source. In virtually all Mediterranean countries that were part of the Roman Empire, there are remains of numerous Roman aqueducts originating at karst springs, some of which are preserved in almost full glory (Fig. 4.)



Figure 3 Encounter with a summer storm in Bat River Cave, Minnesota Cave Preserve. *Upper left*: Cave passages were dry before the storm. *Upper right*: After sudden intense storm on August 18, 2007, a waterfall occurred where a dry ceiling existed before (4:03 pm); *Lower Left*: The water begins to rise and becomes turbulent (4:17 pm); *Lower right*: Seconds after this photo was taken it was almost impossible to stand in the passage without getting swept away (4:59 pm). The cavers escaped by the ladder extending from the newly drilled vertical shaft to the cave. Photographs courtesy of John Ackerman.



Figure 4 *Left*: Vrelo Jadar is a first magnitude karst spring (average discharge rate is 9.8 m^3/s or 346 cfs) utilized for the water supply of Split, the largest Dalmatian port in Croatia (photo by Zoran Stevanović). This spring was used for millennia and was tapped by the Romans for the water supply of the Palace of Emperor Diocletian. Water was conveyed to the palace by an aqueduct, preserved to this day (*top right* photo by Ivo Eterović). *Bottom right*: Reconstruction of the Palace in its original appearance upon completion in 305 AD by the French town planner Ernest Hébrard, published in 1912 (E. Hébrard and J. Zeiller: Spalato, le Palais de Dioclétien, Paris, 1912.) Diocletian retired to this palace near his birthplace Salona (Solun in Croatian). It is the largest and best-preserved example of Roman palatial architecture in the world. It is now the core of the Split's Old Town, designated a UNESCO World Heritage Site in 1979.

The custom of building handsome fountains at the end point of aqueducts that brought karst water to Rome was revived in the fifteenth century, during the Renaissance. The Trevi Fountain (Fig. 5) is the largest Baroque fountain in the city and one of the most famous fountains in the world, fabled for its water clarity. The city of Rome still relies on large karst springs for its water supply (Fig. 6).



Figure 5 The Baroque Trevi Fountain in Rome designed by Nicola Salvi and finished in 1762. Water for the fountain is brought via Aqueduct Vergine, the revived ancient underground aqueduct Aqua Virgo, from the 13 km distant karst spring Salone. According to a legend, thirsty Roman soldiers asked a young girl for water. She directed them to the springs that later supplied the aqueduct. The source was named the Aqua Virgo after her.



Figure 6 Outflow from the pumping station at karst springs of Peschiera which are in the Velino River valley, a few miles upstream of Cittaducale in the province of Rieti. The springs are the largest in central Apennines with an average flow rate of 17 m^3/s (600 cubic feet per second, cfs). They are collected and diverted to supply Rome and much of the surrounding area.

Karst aquifers are present on all continents and numerous oceanic islands with outcrops on over 20% of the land surface, with more being covered by insoluble rocks. As the importance of water resources and their sustainability is becoming critical worldwide, karst aquifers are receiving rapidly increasing attention due to the many challenges related to their characterization, management, and protection.

Introduction

A warning of what may soon be happening to many karst regions around the world, some of which have been experiencing sinkhole hazards for generations already, is the unfolding catastrophe in Turkey described in an article published by The Guardian on February 28, 2026 (‘I live in constant fear’: surge in giant sinkholes threatens Turkey’s farmers | Sinkholes | The Guardian). Below are some excerpts and photographs from this alarming article:

“Every night I pray before I go to bed and when I wake up I pray again,” said Sik. “I live in constant fear that a sinkhole will take my house.”

Konya, part of the once-fertile Central Anatolia region, gave life to ancient civilisations, including what is believed to be the world’s first agricultural society, at Çatalhöyük, in about 8,000BC. It is dotted with the remnants of water cults, Hittite sacred springs and Roman aqueducts, and once offered vital watering holes to traders on the Silk Road.

Now, though, the land is drying up. Turkey is on the brink of a major drought crisis, with almost 90% of the country at risk of becoming desert.

Sinkholes are appearing in farmland in the region at an increasing pace. Experts say there are now almost 700, causing uncertainty and devastation for the farmers who live and work there.

Konya has always been geologically prone to sinkholes because much of the region lies on bases of limestone and other soluble rocks, but in recent decades intensive agriculture has led to heavy groundwater extraction for irrigation. As water tables drop, underground cavities lose the support that once held them up.



Figure 7 *Left*: Sinkholes in a farmer’s field in Karapınar, Turkey. *Right*: The number of sinkholes and related shallow collapse structures in Turkey’s Konya Plain is approaching 3,000. Experts attribute the dramatic rise to regional drought and unsustainable groundwater pumping for agriculture. Photographs by Emre Çaylak. From The Guardian, February 28, 2026.

While no state in the United States with a significant presence of karst is immune to sinkhole development due to anthropogenic influences (e.g., see Fig. 8), the problem is nowhere more drastic than in Florida (Fig. 9). The exploding population and acreage of irrigated agricultural land are putting enormous pressure on groundwater extraction from the Floridan Aquifer, resulting in dropping aquifer levels and increasing sinkhole hazards.

According to the United States Geological Survey (USGS), karst aquifers in the United States, covering roughly 18-20% of the land surface, provide about 40% of the nation's groundwater used for drinking (Karst Aquifers | U.S. Geological Survey). These highly productive, soluble limestone/dolomite systems, including the Edwards Aquifer in Texas and the Floridan Aquifer, both of which are among the most prolific in the world, are crucial for public water supply, agriculture, and the industry.




Figure 8 *Left*: A sinkhole collapse that occurred early Sunday morning August 13, 2006, in Nixa, Missouri is responsible for the disappearance of a garage and the Chevy Cavalier once parked inside. Photo courtesy of Doug Gouzie. *Right*: Golly Hole sinkhole in Alabama, one of the largest in the United States, suddenly opened on December 2, 1971, in northern Calera, Shelby County. Photo by George Sowers, printed with kind permission of Francis Sowers.




Figure 9 Winter Park sinkhole occurred during the spring of 1981. It severed two city streets, truncated a municipal swimming pool, swallowed a house, and damaged an automobile repair shop together with several automobiles parked behind it. The final hole diameter was more than 300 ft (90 m) with the cone-shaped opening approximately 50 ft (15 m) deep through clayey sands. Below was a 60 ft (18 m) diameter cylindrical opening in a soft, porous limestone, possibly an old collapse sinkhole in the limestone. Movement was apparently triggered by wide fluctuations in the groundwater level from well pumping and possibly aggravated by leaking of the swimming pool. Photographs by George Sowers, printed with kind permission of Francis Sowers.

Examples of the catastrophic influence of aquifer pumping on karst water resources that sustain aquatic life and the environment in general are numerous in the United States and around the world. One such example is Comanche Springs in Fort Stockton, Pecos County, Texas:

Introduction

These artesian springs, issuing from a Comanchean limestone groundwater reservoir, formerly flowed as much as 66 ft^3/s, and served the Comanche and other Indians for uncounted hundreds of years. From 1875 on the springs were the basis for an irrigation district which supplied water to 6,200 acres of cropland. Heavy pumping of the aquifer lowered the water table so that the spring discharge began to fall off in May 1947. The irrigation district sought an injunction in 1954 to restrain pumping which interfered with the normal flow of Comanche Springs. The injunction was denied by the courts, and the springs ceased to flow in March 1961. (Gunnar Brune, 1975: Major and Historical Springs of Texas, Texas Water Development Board Report 189, Austin, TX.)



Figure 10 *Left*: Aerial view of Comanche Springs pool in 1938, at the time a popular tourist destination in Fort Stockton, Texas. The springs ceased to flow due to excessive groundwater pumping for irrigation. Photo courtesy of Fort Stockton Historical Society. *Right*: The main spring at Comanche Springs looks like this most of the year. Once a perennial, strong spring, it now flows as an intermittent trickle when irrigation pumping stops and there is enough recharge from above-average rainfall in the recharge area. Photo by Brian Hunt.

The following excerpts are from a press release by Texas State University entitled "Texas springs in crisis: New study highlights overwhelming increase in dry springs":

New research from The Meadows Center for Water and the Environment at Texas State University uncovers a concerning decline in the state's groundwater resources, finding that the number of dry springs has nearly tripled since the early 1980s.

Co-authored by Robert E. Mace, Ph.D., Meadows Center Executive Director, and Nohemi Galaviz, a Texas State undergraduate research assistant, the study revisits and builds upon the seminal work of Gunnar Brune's 1975 report, "Major and Historical Springs of Texas," and his subsequent 1981 book, "The Springs of Texas, Volume 1." As Texas' acknowledged expert on springs, Brune's publications cataloged the state's major and historical springs, forming the foundation of our knowledge today.

"For the first time in nearly half a century, we looked at the status of major and historical springs across the state to see if more springs had gone dry. Indeed, more—many more—have failed over the past 50 years," Mace said.

"The data speaks volumes about the growing demands on our groundwater resources and some of the consequences of meeting those demands," Mace said. "As groundwater use in the state continues to rise, the resulting decrease in springflow jeopardizes not only our groundwater and surface-water supplies but also the ecosystems that rely on these springs."

Jayme Blaschke | February 28, 2024

The disappearance of springs due to large-scale groundwater withdrawals from aquifers is not unique to arid or semi-arid regions with low rainfall and low natural aquifer recharge such as Comanche Springs. Fig. 11 shows historic Kissengen Spring in Florida, located near the city of Bartow in Polk County. This once mighty spring served as a popular recreational area for decades and had an average flow of about 29 ft^3/s (19 million gallons per day, MGD). Kissengen Spring was included in the anthological publication by Oscar Meinzer of the USGS titled *Large Spring in The United States*, published in 1927, as one of the largest in Florida. The cessation of flow, both at Kissengen Spring and at many minor springs in the upper Peace River basin, was related to the regional lowering of the potentiometric surfaces of the intermediate aquifer system and the Upper Floridan aquifer from 1937 to 1950 for as much as 60 feet, due to heavy groundwater pumping for phosphate mining (Peek, 1951). Permanent cessation of the spring flow occurred in April 1960. The spring never flowed again and today only a small mud pot marks the location where, for many years, it provided joy to countless visitors (Fig. 11).

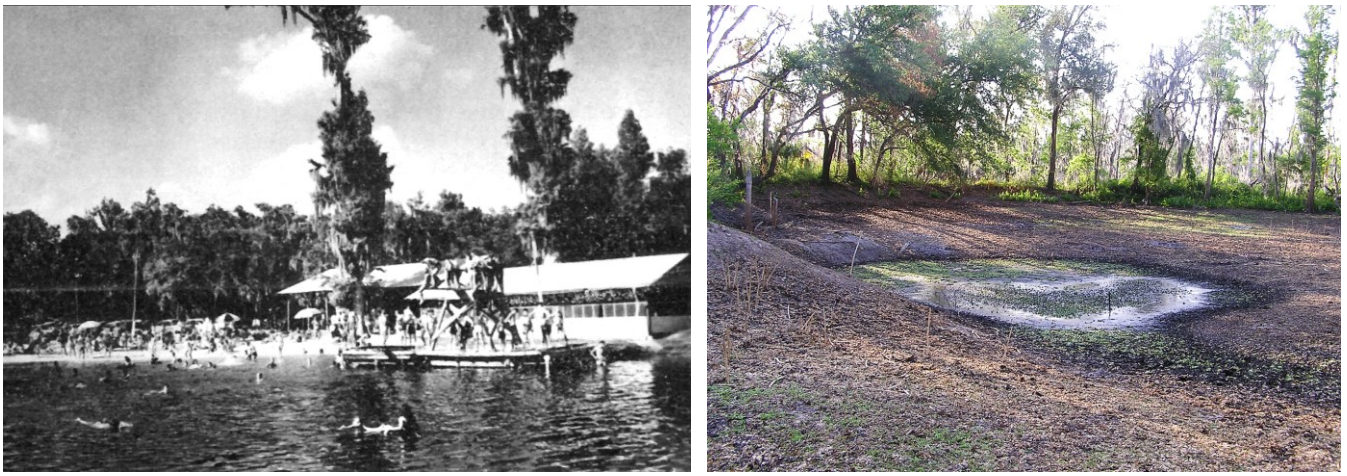


Figure 11 *Left*: Historic photo of Kissengen Spring which remains dry to the present day after all flow ceased in 1960. (From Peek, 1951.) *Right*: Location of former Kissengen Spring in 2006. Courtesy of United States Geological Survey.

Fountain of Sorrow: The Deplorable Fate of Kissengen Springs

From People for Protecting Peace River Inc.—3PR; https://protectpeaceriver.org/

No one person is responsible for the loss of Kissengen Springs, but all signs point to the culpability of an industry and a kind of mass hypnosis that not only allowed the destruction of the springs, but has never held the perpetrators accountable for what happened. On the contrary they accepted their blood money. It's obvious, isn't it? When you walk through miles of useless degraded land, overgrown with invasive vegetation where there are still old cast-off rusty pipes and wires strewn around and huge mine pits gouging the earth a stone's throw from where the springs once prospered, it's obvious what happened to the springs. Yet just down the road, there lies "Mosaic Peace River Park" with its historical brass plaque that says, "The spring ceased to be a tourist destination after the groundwater was captured for other uses." The historical society obviously dared not utter the word "phosphate" since it was the mother of all phosphate mines and fertilizer production that donated the land to the public. The industry mined it and effectively destroyed the natural order that had prevailed there since the dawn of time, and then returned it to the public as a gift, and people bowed their heads and expressed their gratitude.

Introduction

Under relentless pressure from Florida citizens, activists, and many different organizations concerned with the catastrophes facing beautiful Florida springs, the State Legislature in 2016 enacted the Florida Springs and Aquifer Protection Act which specifies 30 "Outstanding Florida Springs". In Statue 373 one can read:

The Legislature finds that the water quantity and water quality in springs may be related. For regulatory purposes, the department has primary responsibility for water quality; the water management districts have primary responsibility for water quantity; and the Department of Agriculture and Consumer Services has primary responsibility for the development and implementation of agricultural best management practices. Local governments have primary responsibility for providing domestic wastewater collection and treatment services and stormwater management. The foregoing responsible entities must coordinate to restore and maintain the water quantity and water quality of the Outstanding Florida Springs.

It remains to be seen if this act will have an effect. In the meantime, those concerned with the sustainability of groundwater resources and the destiny of Florida springs remain concerned, some are skeptical, and some are raising their voices disillusioned with what they see happening as illustrated in the following excerpts from a blog by Dr. Robert Knight, Director of The Howard T. Odum Florida Springs Institute entitled "The Future of Florida Springs" which contains his remarks delivered at a panel discussion held at the University of Florida on April 1, 2023:

But the documented and worsening ecological health of Florida's priceless springs provides ample evidence that Florida's political leadership is not working to fix these problems. Much of the tax dollars being directed to "springs protection" are, in fact, spent on projects that have had no measurable spring benefits. Springs health impairments continue to worsen with every passing year.

This is an unacceptable fate for Florida's most unique and precious natural resources and presents a clear danger for the sustainability of the Floridan Aquifer. The future livability of the Springs Region is not just threatened – it is already substantially diminished. (https://floridaspringsinstitute.org/springsblog/)

The Santa Fe River, fed by karst springs, is a particularly sad story about the mismanagement and demise of Florida's water resources:



Figure 12 *Left*: Foam floating on dark, tannin water of the Santa Fe River in December 2023, just yards away from the Sink pool. *Right*: Santa Fe Sink was choked with green algae, other organisms, white foam, and various debris including pieces of construction lumber and other human-made objects. Photographs taken in December 2023.

In 1983, the Suwannee River Coalition described the Santa Fe River as one of Florida's "most valuable rivers ecologically" as it was in excellent condition. Although they reported problems with exotic vegetation like hydrilla or water hyacinths, it was still classified as "one of the cleanest, healthiest, and least disturbed rivers of its size in the southeastern United States." In the past the Suwannee River Coalition similarly described the St. Johns River as a popular place for Floridians to fish, sail, and swim before pollution had gotten so bad and the dissolved oxygen content so low that fish populations diminished, and the unappealing aesthetics drove visitors away. The 1983 coalition report included an eerie warning to avoid the degradation of the Santa Fe River: "To permit a similar fate for the Santa Fe River would be an irresponsibility of the first magnitude. We must protect this river and it is essential that we designate the Santa Fe River system an Outstanding Florida Water so that it will be enjoyed by generations to come." (Florida Springs Institute, 2020a.)

And yet, the State of Florida, Potentially Responsible Parties (PRPs), and indeed the citizens of Florida all failed miserably. They let down the Santa Fe River, its springs, and its Rise, as attested by the photographs in Fig. 12 taken exactly 40 years after the eerie warning.

This book builds on the materials presented in *Notable Springs of The United States* (Kresic, 2024) which is free for download from various websites such as Groundwater Academy (www.groundwateracademy.org); Karst Waters Institute (KWI, https://karstwaters.org/2024/10/new-open-access-book-on-springs-of-the-us-available-by-kwi-board-member-neven-kresic/); and HydroGeoCenter: (https://hydrogeocenter.com/downloads-and-resources/). The two books are the result of the author's work, as the coordinator for North America, on project MIKAS (Most Important Karst Aquifer Springs; https://mikasproject.org/) by Karst Commission (KC) of The International Association of Hydrogeologists (IAH).

The main goal of this project was to create the first list of the most important karst springs at the global, as well as national levels, representing world and national natural heritage for each country. The importance of each spring is based on historical, aesthetic, economic, scientific and ecological criteria, at the same time recognizing that the five criteria may not be present at all springs and that specific circumstances of each country should be taken into consideration ("Something that is important in one country does not have to be important in others."). Hundreds of karst springs from 30 States in the United States were compiled for the Karst Commission project by the author. Their descriptions are included in this book, most of them accompanied by photographs. General primers on karst, types of karst springs, and major karst aquifers of the United States are also included.

The author is grateful to many colleagues and institutions that provided photographs and information on the karst springs. Special thanks go to Alan Cressler, John Moran, Chuck Sutherland, Phil Lukas, Gregg Eckhardt, Dave Bunnell, Arthur Palmer, Katie Fite, Joseph Honings, Calvin Wolter, Sam Panno, Jake Stacken, Martin Larsen, John Barry, Greg Brick, Gheorghe Ponta, Ryan Maurer, Joel Maynard, William Orndorff, Bill Jones, Calvin Alexander, Nenad Marić, Zoran Stevanović, Alex Mikszewski, John Burtner, David Kreamer, Gretchen Baker, Bethany Neilson, Hyrum Tennant, David Rhea, Todd Kincaid, Hannah Schacht, Tom Provezis, Ben Estes, Katarina Kosič Ficco, Mike Ficco, Amy M. Hourigan, Alan Fryar, Donna Leight, Andrew Wagner, Judith Judee Lyon, and Lydia Sizemore.

Chapter 1 Water in Karst

1.1 Karst and Karstification

Karst is a scientific term named after the geographic district between Slovenia and Trieste, Italy, which has a very distinct landscape. It is a Germanized word for "carso" (in Italian) and "kras" (in Slovene); all three words are derived from the Indo-European word "kar" or "karra," which means rock. The following definitions of karst and karst topography are based on the 1960 edition of the American Geological Institute Dictionary of Geology.

In karst, the limestone rocks are honeycombed by caverns and openings dissolved by groundwater. Most of the drainage is underground and the land is left dry and relatively barren. The landscape is dotted with sinkholes, interspersed with abrupt ridges, and irregular protuberant rocks. Dry valleys are common, and valleys containing streams often end abruptly where the latter plunge into underground channels and caves, sometimes to reappear as large springs miles away. These features are so characteristic of the Karst district of the eastern Adriatic coast that the name karst has been applied to them elsewhere.

Some common words from the Slavic languages of the Karst region have become international scientific terms describing karst features, mainly thanks to the Serbian geomorphologist Jovan Cvijić who in 1893 was the first to defend a doctoral thesis and publish a scientific monograph exclusively devoted to karst: Das Karstphänomen (The Karst Phenomenon). These include *doline* (sinkhole), *ponor* (sink), *jama* (pithole), and *polje* (large, closed depression with flat bottom characteristic of the Dinaric karst). During the following two decades, Cvijić laid the theoretical foundation of many past and current developments in karst geomorphology and hydrogeology by authoring over 40 publications devoted entirely or mostly to karst phenomena. Much of his original work and numerous drawings illustrating the origin of various karst forms and the nature of groundwater flow in karst terrains have since been republished, copied, or otherwise reaffirmed by many. Examples of typical hand drawings made by Cvijić in support of his concepts and illustrating karst features analyzed on his extensive journeys throughout the Balkans and Europe are provided in Figures 1.1 and 1.2.

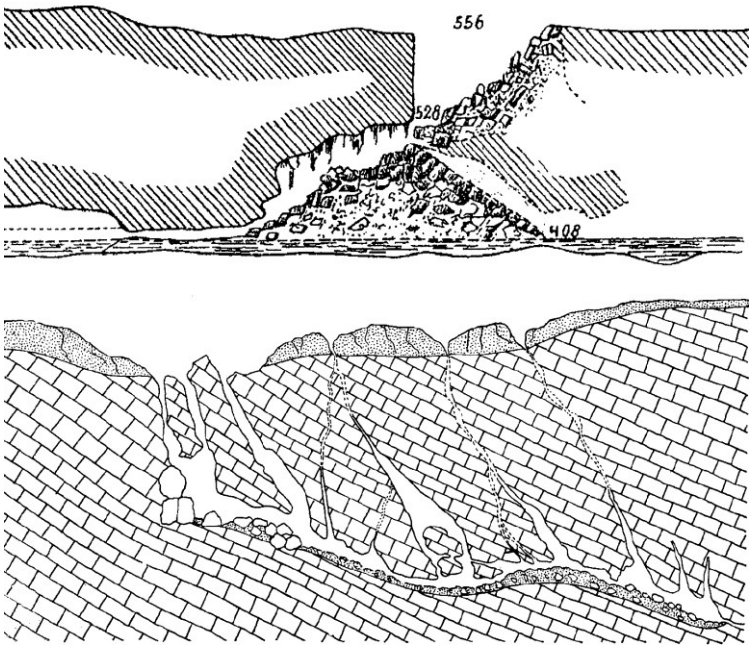


Figure 1.1 Cvijić's drawings accompanying his explanations of the morphogenesis of different karst landforms. *Top*: Collapse sinkhole. *Bottom*: Alluvial sinkholes aligned over karst cavities in underlying limestone. From Cvijić, 1893.

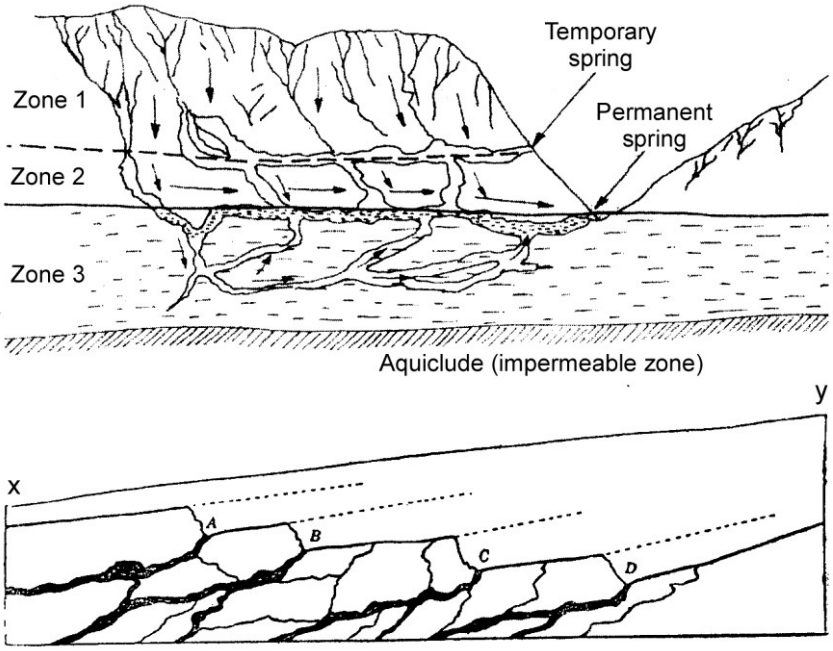


Figure 1.2 Cvijić's sketches illustrating concepts of groundwater flow in karst. *Top*: Schematic section of a karst aquifer with three zones: (1) dry zone with percolation only; (2) transition zone of varying level with both percolation and saturated groundwater flow during high water table; (3) permanently saturated zone, including conduits with ascending flow (English annotation added). *Bottom*: successive lowering of ponors (sinks) of streams in karst creating blind valleys; x-y denotes pre-karstic valley. (From Cvijić, 1918; also available in Stevanovic and Mijatovic, 2005.)



Figure 1.3 *Left*: Underground flow of the Reka River, which sinks into the world-famous *Škocjan Caves* in Slovenia and empties into the Gulf of Trieste, can be accessed in a few deep pits-caves. The longest of them is *Kačna jama* cave near Divača, Slovenia, which impresses with its picturesque abyss entrance 186 meters deep, which in the lower part extends into a huge hall 60 meters high. Photo by Peter Gedei. *Right*: Rokina Bezdana Cave in Croatia. Photo by Dinko Stopić.

Photographs in Figs. 1.3 through 1.11 illustrate some of the spectacular features of the classic Dinaric karst (many more photographs are available in a free-for-download book entitled *The Land of Karst, A Visual Story* (Kresic and Stevanovic, 2023); download is available at www.groundwateracademy.org; and the academic Research Gate site https://www.researchgate.net/publication/383145893_The_Lands_of_Karst_A_visual_Story.





Figure 1.4 *Left*: *Prerast* in Vratna River Gorge, one of many natural bridges in eastern Serbia. Photo by Aleksandra M. Stevanović. *Right*: Karst spring waterfall "Skakavica" on Grabovica river in Montenegro. Photo by Dobrislav Bajović Bajone.



Figure 1.5 Underground canyon of *Škocjanske jame* (Škocjan Caves) in Slovenia. It is criss-crossed by precipitous paths and was fully explored already in the 19th century. After sinking, the Reka River penetrates deep into the cave and descends over rapids and falls along its entire length, until it reaches an unpassable syphon. The vertical drop from the swallow hole to the syphon exceeds one hundred meters. Note lit tourist path and bridge in the upper left for scale. The cave was designated World Natural Heritage site by UNESCO in 1986, one of only four caves in the world to receive this status. Photo by Borut Lozej.

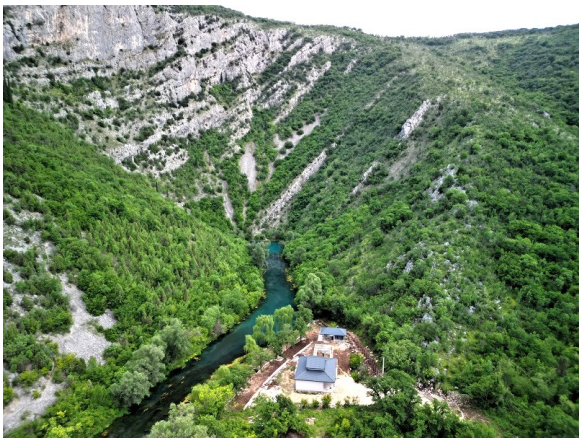


Figure 1.6 *Left*: Ombla Spring (*Vrelo Omble*) is the largest individual karst spring in the world, with the recorded long-term average discharge rate of 24.4 m^3/s (862 cfs). Before diversion of some of the flow of the sinking Trebišnjica River in the spring's vast watershed for hydropower generation at HE Dubrovnik, the average discharge rate was 33.8 m^3/s (1194 cfs). The multi-decadal continuous daily measurements are officially conducted by the Croatian Water Management Agency. The spring is of ascending type, issuing at contact between Mesozoic limestones and impermeable Eocene flysch on the Adriatic coast. It is used for the water supply of Dubrovnik. *Right*: Aerial view of Ombla Spring and the river (Dubrovačka Rijeka) originating from it. Photo courtesy of Antun Maškarić.



Figure 1.7 Buna Spring (*Vrelo Bune*) near Mostar, Bosnia and Herzegovina is a deep siphonal (vauclusian-type) spring and the second largest individual karst spring in the world with a recorded long-term average discharge rate of 23.7 m^3/s (837 cfs). Multi-decadal continuous daily measurements are officially conducted by the Water Management Agency of Bosnia and Herzegovina. The spring issues at contact of Cretaceous limestone and Eocene flysch, along a major regional fault which is marked by dissolution cavities at various elevations – evidence of the lowering of locations of discharge. Elevation of the siphon in the submerged cave is 68 m (223 ft) below discharge level. The spring is one of only three springs included on the list of World Geoheritage Sites by the International Union of Geological Sciences. The other two are also karst springs, Wakulla Spring in Florida, and Fontaine de Vaucluse in France.
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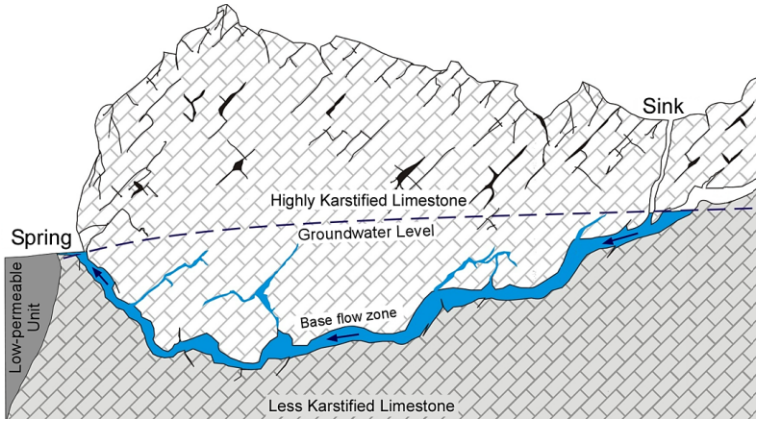


Figure 1.8 Vauclusian-type Little Buna Spring (*Vrelo Bunice*) is the third largest individual karst spring in the world with the long-term average discharge rate of 20.25 m^3/s (715 cfs) and a whopping mean maximum rate of 207 m^3/s. (7310 cfs). Multi-decadal continuous daily measurements are officially conducted by the Water Management Agency of Bosnia and Herzegovina. Two sister springs, *Vrelo Bune* and *Vrelo Bunice* are only 2.5 miles apart and jointly constitute the largest "spring group" in the world if Florida standards were applied (44 m^3/s or 1554 cfs combined long-term average discharge rate). This extraordinary flow creates a spectacular site at the confluence of the Buna River and the Neretva River, the largest tributary of the eastern Adriatic (Fig. 1.9). For comparison, the two largest spring groups in the United States are Malad Gorge Springs and Thousand Springs in Idaho with average discharge rates of 1229 cfs and 1198 cfs respectively. They both issue from young basalts, often referred to as *pseudokarst* (see Volume 2 of this book and Chapter 4 in Kresic, 2024). Photographs by Robert Oroz, courtesy of *Snaga lokalnog*, a five-year program of the former American Agency for International Development (USAID) that worked to empower local communities in foreign countries. Conceptual cross-section from Saša Milanović.

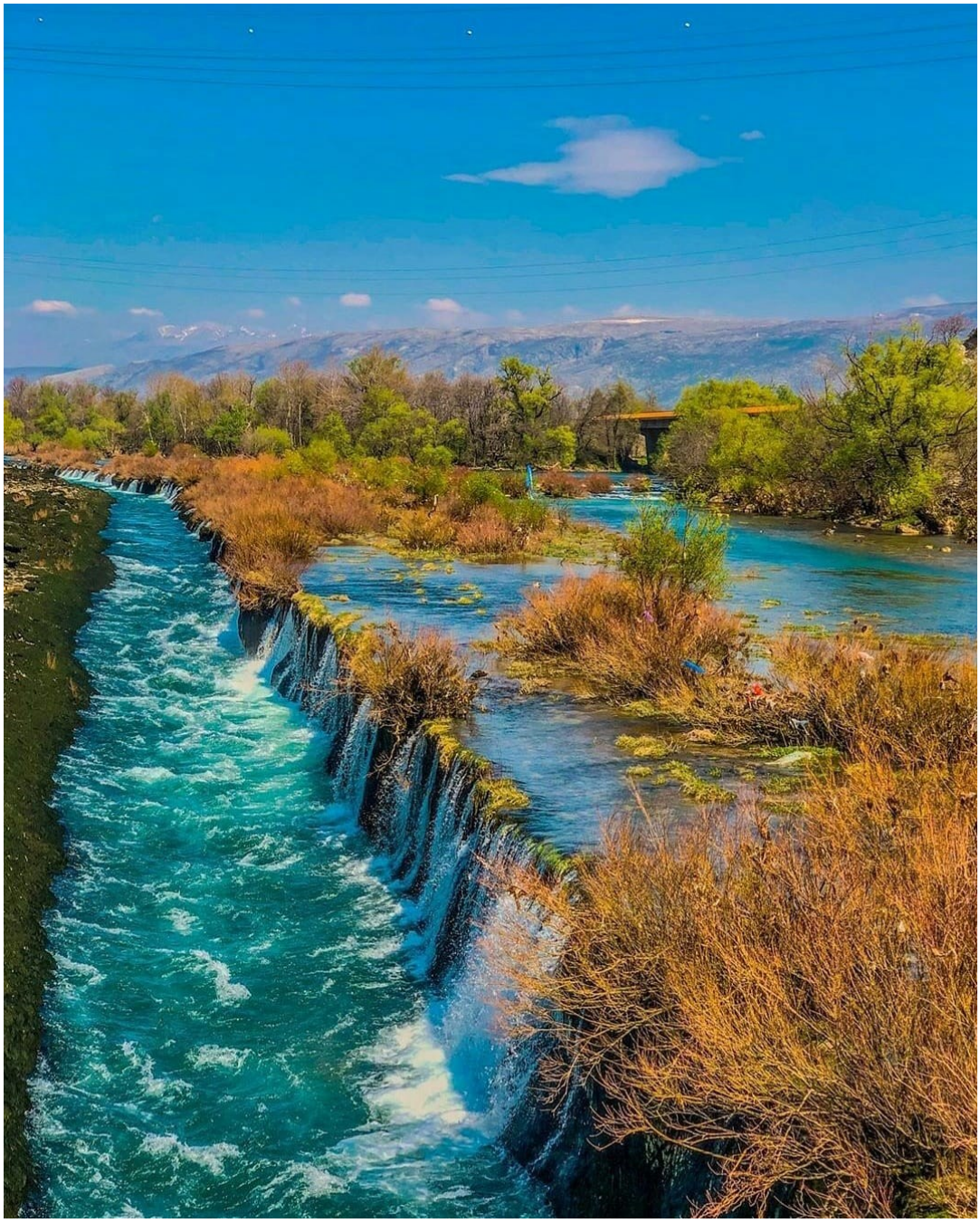


Figure 1.9 Spectacular confluence of the Buna River (on the right, formed by waters of Buna and Little Buna Springs) and the Neretva River (on the left), 9 miles south of Mostar, Bosnia and Herzegovina. “Nera-Etwa" is believed to be an ancient Illyrian name for the Neretva River. It is commonly translated to mean "Flowing Goddess", or "Heavenly Water". Photo available at https://www.facebook.com/visitherzegovina/photos/

What sets karst aquifers apart from any other hydrogeologic environment is the nature and dynamics of surface water-groundwater interactions, and their unique porosity, often referred to as triple porosity, which consists of the rock matrix (initial, primary) porosity, the porosity of fractures (secondary porosity), and the porosity of solution openings (tertiary porosity: enlarged fractures, karst channels, conduits, and all cavities in general). The solution

openings are created by the chemical and mechanical erosion of water percolating into the rock from the land surface and through unsaturated soils, and its enrichment with dissolved carbon dioxide along the way.



Figure 1.10 Glavaševo Vrelo, headspring and the largest spring of the Cetina River near the village of Cetina in Dalmatia, Croatia. This vauclusian-type spring is exceptionally deep, with divers descending 150 meters (490 feet) without reaching the bottom. Average discharge rate is 3 m^3/s (106 cfs). The spring is used for water supply and swimming is prohibited. In recent years, the spring acquired a new name "Eye of the Earth", thanks to a campaign of savvy tourism officials. Photo courtesy of Branimir Jukić.



Figure 1.11 Vrelo Sopot in Boka Kotorska Bay, Montenegro is the largest intermittent karst spring in the world, exceeding 300 m^3/s (10,600 cfs) when flowing after heavy rains. Most of the year the spring is dry and the vast karst aquifer of the Orjen Mountain is drained by a nearby submarine spring. The Orjen Mountain has the highest precipitation rate in Europe (close to 5,000 mm/year). Photo courtesy of Saša Milanović.

The interaction between water and soil rich in carbon dioxide creates a weak carbonic acid which accelerates dissolution of soluble carbonate rocks: primarily limestone, dolomite, marble, and gypsum. The following equations express the general reactions between calcium carbonate and the CO_2–water system:

$$CO_2 + H_2O = H_2CO_3 = H^+ + HCO_3^-$$
$$HCO_3^- = H^+ + CO_3^{2-}$$
$$CaCO_3 + H^+ = Ca^{2+} + HCO_3^-$$



Figure 1.12 Cavities of varying size discovered at a deep (seen dewatered here) construction site in Hartsville, Tennessee. Consolidated old Paleozoic limestones have low matrix porosity of less than 5% but high secondary and tertiary porosity as they are intensely fractured and karstified. Photographs by George Sowers; printed with kind permission of Francis Sowers.

The density, size, and geometry of cavities created by carbonate dissolution in karst aquifer networks can vary by many orders of magnitude depending on site-specific characteristics such as mineral composition, stratigraphy, tectonic fabric, recharge mechanisms, and position of the erosional base. One of the most common errors made by hydrogeologists not experienced in karst is to rely solely on investigative borings and monitoring wells when developing a conceptual model for their site. Declaring that the aquifer (site) is non-karstic based on three, five or even ten borings that did not encounter a large "void" is a typical example. Similarly, a non-karst hydrogeologist may conclude that the aquifer is not karstic because test wells cannot produce "large" yields that, somehow, are expected from a karst aquifer. Fig. 1.12 illustrates these points. One can easily imagine a boring advanced into the large cavity shown on the left photograph or the smaller cavity on the right photograph. In such cases hardly any hydrogeologist would argue that the limestone is not karstified (an inexperienced driller not equipped for drilling in karst may even lose some of his/her equipment when this happens). One could also imagine another boring completed in solid rock adjacent to either cavity that could be used, by some, as "evidence" of non-karst.

The situations shown on Fig. 1.12 can happen at any depth within a karst aquifer including hundreds or thousands of feet below ground surface and below the water table. The depth to which karstification has progressed, the so-called base of karstification, is not uniform and varies from site to site. The degree of karstification and fracture density generally decrease with depth but can locally reach much greater depths. As the erosional basis for karst groundwater discharge is not necessarily at a static elevation and may be constantly lowered due to surface stream incision, sea level regression, or other factors, the depth of karstification and the depth to water table would consequently increase in time as well. For example, in some areas of the Dinaric carbonate platform in Europe the depth to the water table exceeds 1000 meters, while cavities and voids of varying size have been discovered with investigative drilling at depths of several thousand meters below sea level.

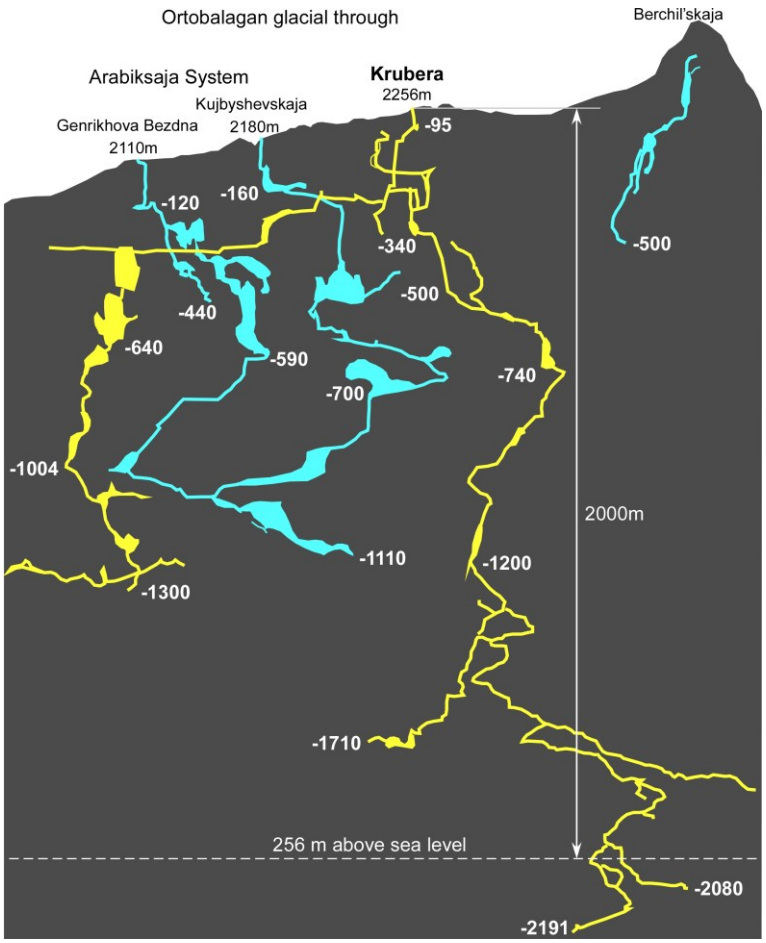


Figure 1.13 Combined cave profile along the axis of the Ortobalagan valley, Western Caucasus, breakaway republic of Abhazia (formerly in Georgia). Krubera Cave is shown in yellow (bottom of the cave is 2191 meters below ground surface), other caves in blue. Simplified from Klimchouk, Samokhin, and Kasian, 2009. Printed with permission from the authors.

Even more impressive are examples of deep karstification on the Arabica Karst Massif, part of the Western Caucasus mountains bordering the Black Sea. The Krubera Cave (Fig. 1.13) in the breakaway republic of Abhazia (formerly in Georgia) is the second deepest cave in the world: 2191 meters or 7188 feet as of May 2022 (the nearby Veryovkina Cave is the world's deepest at 2212 meters or 7257 feet). Krubera has a very steep profile and reveals an enormous thickness of the vadose zone. The lower boundary of the vadose zone (the top of the phreatic zone) is at an elevation of about 110 m above sea level (asl) at low flow. Submarine discharge along the Black Sea coast is reported at depths up to ~400 m below sea level (bsl). A large, closed submarine depression is revealed at the sea floor next to the Arabika Massif, with its deepest point at ~400 m bsl. These facts point to the possibility that the main karst system in Arabika could have originated in response to the Messinian salinity crisis (5.96–5.33 Ma) when the Black Sea (Eastern Paratethys) could have almost dried up, as did the adjacent Mediterranean where a dramatic sea-level drop is well established. Further development of the thick vadose zone and the super-deep caves have been caused by subsequent uplifts during the Pliocene-Pleistocene (Klimchouk et al., 2009.)

Similarly, along the coasts of Florida in the United States, Yucatan in Mexico, and the Caribbean islands there are numerous submarine springs and submerged deep sinkholes and cave passages testifying that the sea level in the past was also lower (Figs. 1.14 and 1.15).

Paleokarst is a term usually used to describe karst features (voids, cavities) at any depth explained by the burial of carbonate sediments below non-soluble rocks. The carbonates were karstified at earlier times when exposed at the surface or when in near-surface settings. These paleo cavities may subsequently be filled with clastic deposits

or may remain open. In some cases, the existence of karstification at great depths (several miles or so) can be explained by the fact that the same general area with carbonate rocks may have been subject to multiple periods of karstification depending on geologic and tectonic history. In thick karst aquifers developed in active geologic environments it is often possible to find very permeable zones, together with large karst conduits, at greater and varying aquifer depths. These zones may indicate the position of a paleo water table where karstification intensity was high.

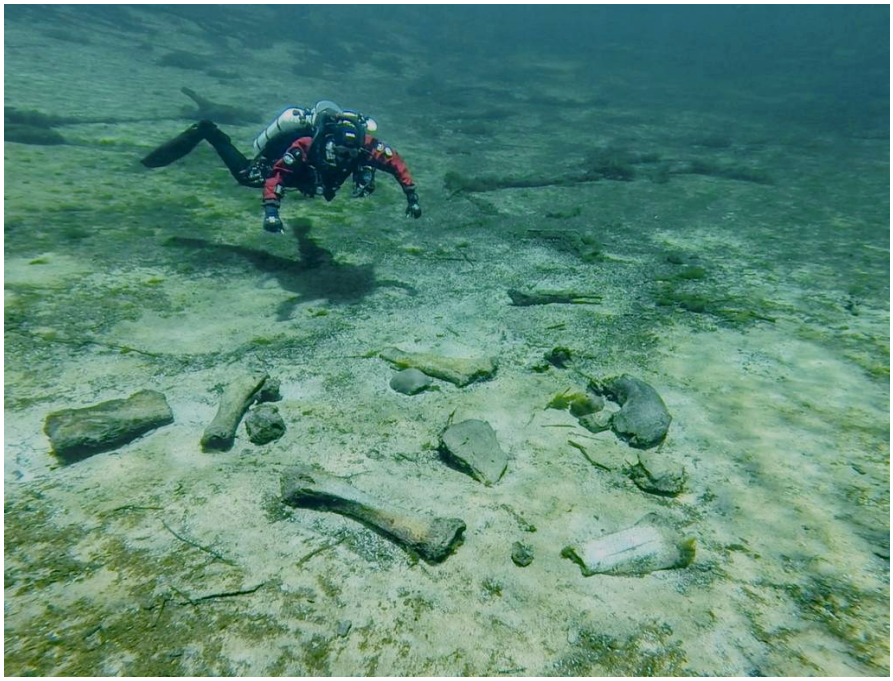


Figure 1.14 Diver Casey McKinlay swims over mastodon fossils in the Wakulla Spring vent. In a West Florida State University documentary Finding the First Floridians, one can learn that fossils are well preserved in springs and sinkholes. One can also learn about Florida's lower aquifer levels during the last ice age, when humans and mastodons coexisted. Photo by David Rhea, Woodville Karst Plain Project.



Figure 1.15 Blue hole on East Caicos is a deep submerged sinkhole. The limestone in which numerous caves in the Turks and Caicos Islands are formed is young, Pleistocene-age biogene reef limestone, roughly 120,000 to over 195,000 years old. The primary porosity of such reef limestones is generally high, with values often ranging from 30% to over 40% (e.g., Diaz et al, 2019). For more information see Lace, et al., 2016, Coastal Cave Survey Times of the Islands Caving In for Science - Times of the Islands. Photo courtesy of Caves, Blue Holes and Sinkholes | Visit Turks and Caicos Islands.

In any case, open karst cavities at great depths may be preserved even where pressure increases to lithostatic values, due to the high rigidity of limestone and dolomite. Numerous such examples from around the world are provided by Andreychouk et al. (2009) including from North America where data reveals that Mesozoic and Paleozoic carbonate rocks may exhibit enhanced secondary porosity, up to 15-20%, at depths up to 10 km. As concluded by the authors, available data unequivocally indicate that karst can develop throughout the entire section of the Earth's crust that is accessible by drilling. The distribution of karst voids at different depths is irregular.

Relatively large cavities, with dimensions exceeding meters, occur in the top few kilometers. Smaller pores, vugs, and enlarged fractures prevail deeper. The authors note that this spatial heterogeneity can be exaggerated, at least partly, by the technical difficulties of identifying large cavities at great depths.

In a series of publications, Palmer (1991, 2007), Klimchouk (1992, 1994, 2011) and Klimchouk et al. (2000) established the concept of hypogenic (endogenic) karst, previously largely ignored by karst hydrogeologists trained in traditional karst settings. As mentioned earlier, the well-established concept, applicable to many open hydrogeologic structures, explains the karstification to be the result of carbonate rock dissolution by infiltrating (descending) waters enriched with soil and atmospheric CO_2, the so called *exogenic* or *epigenic* karst. In contrast, *hypogenic* karst is created by ascending waters that, in addition to CO_2, may carry upwards other acidic fluids such as sulfuric acid and dissolved H_2S, the origin of which is deep (endogenic) due to metamorphism, magmatic, and tectonic processes. Dissolution of carbonates by ascending groundwater is enhanced in semi-confined and confined hydrogeologic settings where cross-formational diffuse flow of the acidic fluids can take place over large areas and create karst cavities by slowly "eating away" the rock mass (Fig. 1.16). This process can result in the formation of very convoluted and extensive karst cavities along previously weakened portions of the rock (such as due to fracturing and faulting) as well as in relatively competent rock. Hypogenic karst does not have to be deep and exclusively the result of dissolution by endogenic fluids such as sulfuric acid. The sufficient triggers for its development are an ascending groundwater flow and some degree of the carbonate aquifer confinement with less permeable rocks. In many cases there will be an overlap between the ascending and descending flows (i.e., the hypogenic and exogenic karst respectively), especially at the locations of regional groundwater discharge.

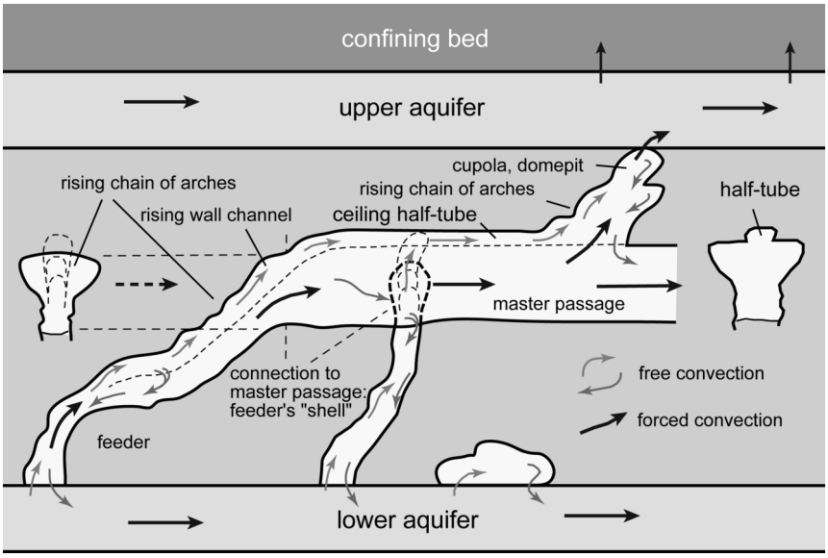


Figure 1.16 Morphologic suite of rising flow, diagnostic of hypogenic transverse origin of caves. Modified from Klimchouk, 2011; printed with original permission from the author.

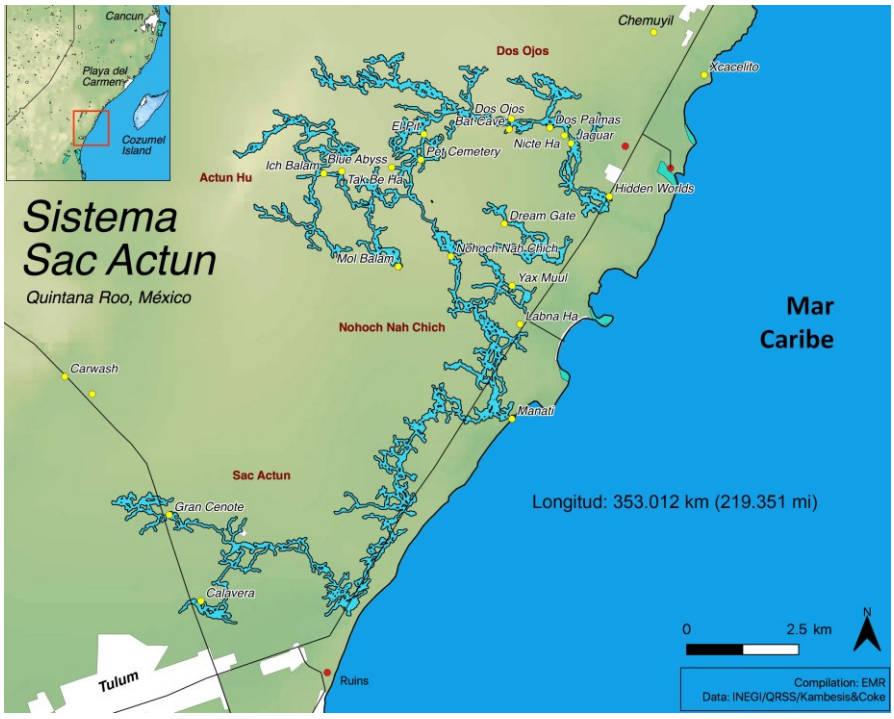
In summary, the only relevant characteristics of an aquifer to be qualified as karstic is that a portion of groundwater flow in it is taking place through a network of interconnected channels, conduits, and various other voids created by dissolution of the rock. These voids can be the result of several physical and chemical processes involving descending (infiltrating) water, as well as ascending groundwater that may also carry acidic fluids such as dissolved H_2S, sulfuric acid, and CO_2 from greater depths. Regardless of the actual mechanism involved, the relative size and the number of interconnected karst voids are irrelevant. If they are present, at any scale, the aquifer is karstic.



Figure 1.17 Lechuguilla Cave in New Mexico is the most impressive example of a hypogenic cave in the United States. It is situated within the 265–259 million-year-old Capitan Reef Complex in the Guadalupe Mountains. The sulfuric acid process left behind unusual mineral deposits not typically found in other caves, such as impressive gypsum chandeliers shown here in The Chandelier Ballroom. Because of its extreme fragility and scientific value, access to the cave is strictly limited to approved researchers and exploration teams; it is not open to the public. Photo courtesy of Art Palmer.

Unfortunately, there still are some misinterpretations of the true nature of karst which may lead to an unwanted propagation of conceptual errors that become entrenched in professional practice. One example is the following quote from Zerga (2024) which is an apparent citation from the book by Ford and Williams (2007): "*Karst cannot be produced solely by a significant amount of rock solubility. Rock structure and lithology are also significant factors. Karst is often poorly developed in soluble rocks with extremely high primary porosity (30–50 %).*"

It of course goes without saying that the above statement is far from the truth. Some of the most spectacular and longest caves, deep sinkholes, and largest karst springs in the United States and around the world are developed in *soluble rocks with extremely high primary porosity (30–50 %);* for example, see the following discussion and Figs. 1.14 and 1.15, Figs. 1.18 through 1.21, Fig. 1.28, Fig. 1.37, and Fig. 1.46.



In a paper entitled "Cave Development on the Caribbean coast of the Yucatan Peninsula, Quintana Roo, Mexico", Smart et al. (2006) explain a number of conditions for development of one of the most fascinating collections of surface and subsurface karst landforms in the world.

Figure 1.18 Map of submerged cave passages of *Sistema Sac Actun* in Mexican state Quintana Roon on Yucatan Peninsula north of Tulum. Courtesy of Dr. Emiliano Monroy-Ríos, Northwestern University. This map and many more maps, photographs, visuals, and descriptions of the fascinating Yucatan karst features are available at https://sites.northwestern.edu/monroyrios/

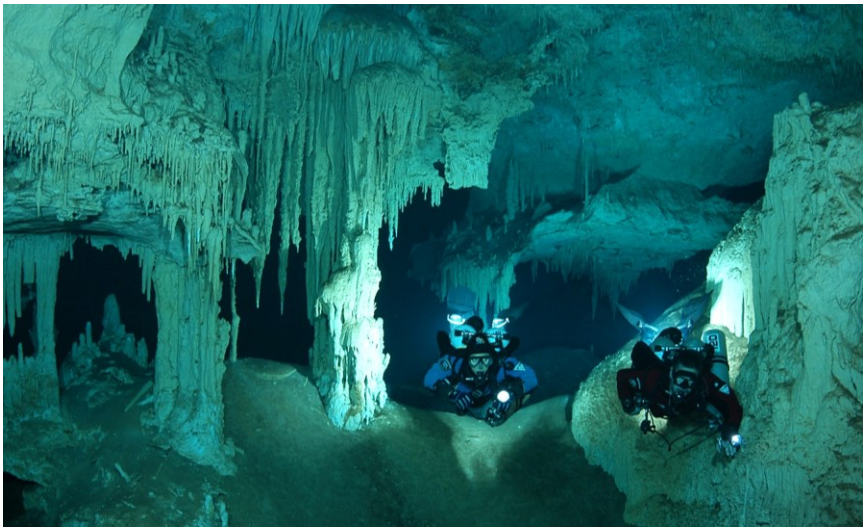


Figure 1.19 *Nohoch Nah Chich*, situated north of Tulum in the state of Quintana Roo, Mexico was once a standalone, famous cave system; it was connected to the *Sistema Sac Actun* in 2007, significantly expanding the overall length of the underwater network. The abundant speleothems were formed before the cave was submerged due to rising sea level after the last major glaciation. Photo courtesy of David Rhea, Global Underwater Explorers.

Below are couple of excerpts from this excellent work as related to the previously discussed incorrect statement that "Karst is often poorly developed in soluble rocks with extremely high primary porosity (30–50 %)."

The limestones may still be diagenetically immature, with a high primary porosity and presence of metastable carbonates, while the ubiquitous fracture sets of telogenetic karst are generally absent."…"*However*, *across most of the study area, the preferred cave inception flow pathways are probably associated with linked vug porosity and developed by random linkages between dissolution-susceptible fragments and existing voids.*"

The area of Quintana Roo in Yucatan, Mexico features hundreds of miles of submerged cave passages formed in young Pleistocene limestones which have a primary porosity ranging from 29% to 50%. This includes the longest submerged cave system in the world, *Ox Bel Ha* (the name "Ox Bel Ha" is Yucatec Maya for "Three Paths of Water".) As of early 2026, it has a surveyed length of approximately 336.6 miles (541.7 kilometers).

The title of the world's longest submerged cave has frequently changed hands between *Ox Bel Ha* and *Sistema Sac Actun* (White Cave), now the second longest with 228.7 miles (368 km) of surveyed passages as of 2026.



Figure 1.20 Cenote Oxman, about 25 miles east of Chichén Itzá, Yucatan. Photo by Carlos Rosado van der Gracht, courtesy of Yucatan Magazine. The word "cenote" comes from the Mayan word "dz'onot", which means "cavern with water". It describes a natural pit in limestone terrain, formed by the collapse of the top of a cavern. These large water holes played a very important role in the life of the Maya. The Yucatán has virtually no surface rivers because all the water drains directly into the porous limestone. For the ancient Maya, cenotes were therefore not just scenic wonders; they were the indispensable source of life—the only reliable year-round water supply in a dense jungle.



Figure 1.21 Devil's Ear Spring, part of the Ginnie Springs cave system in Gilchrist County, Florida. This submerged cave system, approximately 5.7 miles long, is developed within the Ocala Limestone, which is an Eocene-aged limestone, making it approximately 34 to 38 million years old. The primary porosity of the Ocala Limestone at Ginnie Springs is remarkably high for a rock of its age, typically averaging around 30% to 40% (Florea and Vacher, 2009). Unlike older "telogenetic" limestones (such as those in Kentucky's Mammoth Cave), which often have primary porosities of only 2–3%, the Ocala Limestone acts like a massive, hard sponge. Ginnie Springs achieved an unrivaled fame among scuba divers of all skill levels when Jacques Cousteau dove here in 1974 and praised the crystal-clear water (72°F year-round) as "visibility forever." It is the international hotspot for scuba diving. Photo by John Moran; all right reserved, printed with permission.

Cave passages in soluble carbonate sediments provide an invaluable insight into the role geologic fabric plays in the development of karst and the control of groundwater flow directions. Such passages, developed by dissolution along initial discontinuities in the rock mass, are literary evidence "written in stone". An illustrative excerpt from an excellent publication by Swezey et al. (2017) follows below. It illustrates the impact of geologic fabric on karst development and inckudes Figs. 1.22 through 1.24.

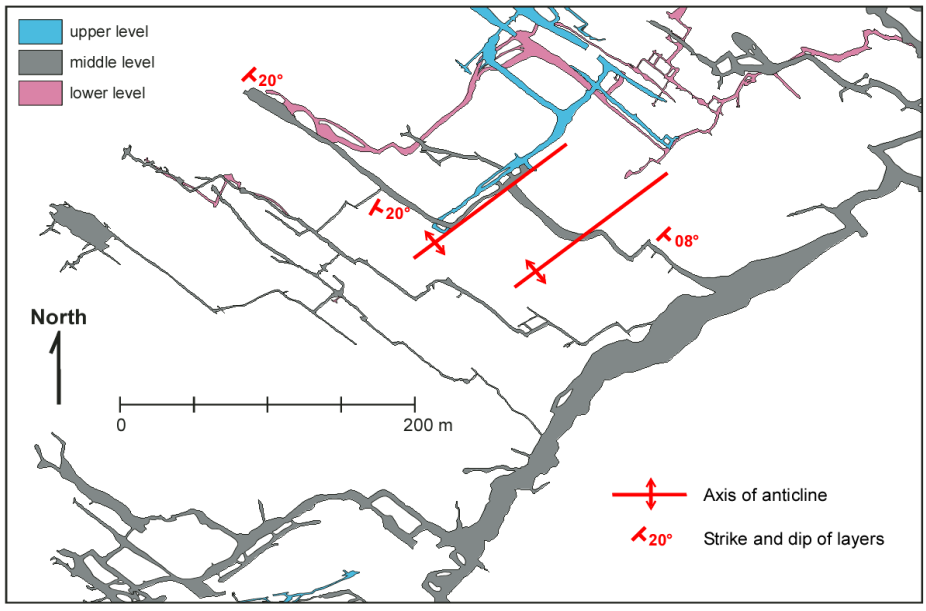


Figure 1.22 Section of map of Butler Cave – Sinking Creek Cave System, Bath County, Virginia showing mapped cave passages. Modified from Swezey et al., 2017, *Geologic controls on cave development in Burnsville Cove, Bath and Highland Counties, Virginia*. From the Blue Ridge to the Beach: Geological Field Excursions across Virginia. The Geological Society of America Field Guide 47; originally modified from White, 2015.

Collectively, the structural and stratigraphic setting of Burnsville Cove influences cave passage morphology. For example, many of the cave passages are tall and narrow, and follow joints. Other cave passages are relatively wide and have flat ceilings and (or) floors. Such passages typically form where the cave ceiling and (or) floor follows a bedding plane. Yet other cave passages have arched ceilings where the passages follow folds in the strata."…"In some instances, both bedding planes and joints exert strong controls on passage morphology, leading to a trapezoid passage profile with a narrow inverted V-shaped cavity at the top of the trapezoid shape.



Figure 1.23 Examples of fissure control on development of passages in caves of Highland and Bath Counties, Virginia. All photos courtesy of Phil Lucas.



Figure 1.24 Lost Cove Cave twilight, Crow Creek, Mr. and Mrs. Harry Lee Carter Natural-Scientific State Natural Area, Franklin County, Tennessee. The cave passage is developed along bedding planes. Photo courtesy of Chuck Sutherland.

Faults are one of the key structural elements that influence the development of preferential flow paths in karst aquifers and groundwater flow directions at various scales. They often form hydraulic boundaries for groundwater flow in both consolidated and unconsolidated rocks. They may have one of the following three roles: (1) preferential pathways for groundwater flow, (2) storage of groundwater due to increased porosity within the fault (fault zone), and (3) barriers to groundwater flow due to decrease in porosity within the fault. The following discussion by Meinzer (1923) illustrates these points:

> *Faults differ greatly in their lateral extent, in the depth to which they reach, and in the amount of displacement. Minute faults do not have much significance with respect to ground water except, as they may, like other fractures, serve as containers of water. But the large faults that can be traced over the surface for many miles, that extend down to great depths below the surface, and that have displacements of hundreds, or thousands of feet are very important in their influence on the occurrence and circulation of ground water. Not only do they affect the distribution and position of aquifers, but they may also act as subterranean dams, impounding the ground water, or as conduits that reach into the bowels of the earth and allow the escape to the surface of deep-seated waters, often in large quantities. In some places, instead of a single sharply defined fault, there is a fault zone in which there are numerous small parallel faults or masses of broken rock called fault breccia. Such fault zones may represent a large aggregate displacement and may afford good water passages.*

As discussed by Hatcher (1995), joints frequently form adjacent to brittle faults. Movement along faults commonly produces a series of systematic fractures in which spacing decreases closer to the fault zone, and the number of sets of joints increases. It is for this reason that fault zones in carbonate sediments and other rocks that do not have significant fine-grain fraction (sandstones are an example) do act as preferential flowpaths and can collect and transfer significant quantities of groundwater. Some of the largest springs in the United States and worldwide owe their existence to faults. For example, San Marcos and Comal Springs of Central Texas are the largest springs in the southwestern United States. The artesian springs emerge at the base of the Balcones Escarpment, a series of large-displacement faults that force the water of the karstic Edwards aquifer to discharge at he land surface (Figs. 1.25 and 1.26). Other good examples of fault-controlled springs are springs in the classic Dinaric karst of former Yugoslavia, some of which are the largest in the world (see Fig. 1.7).

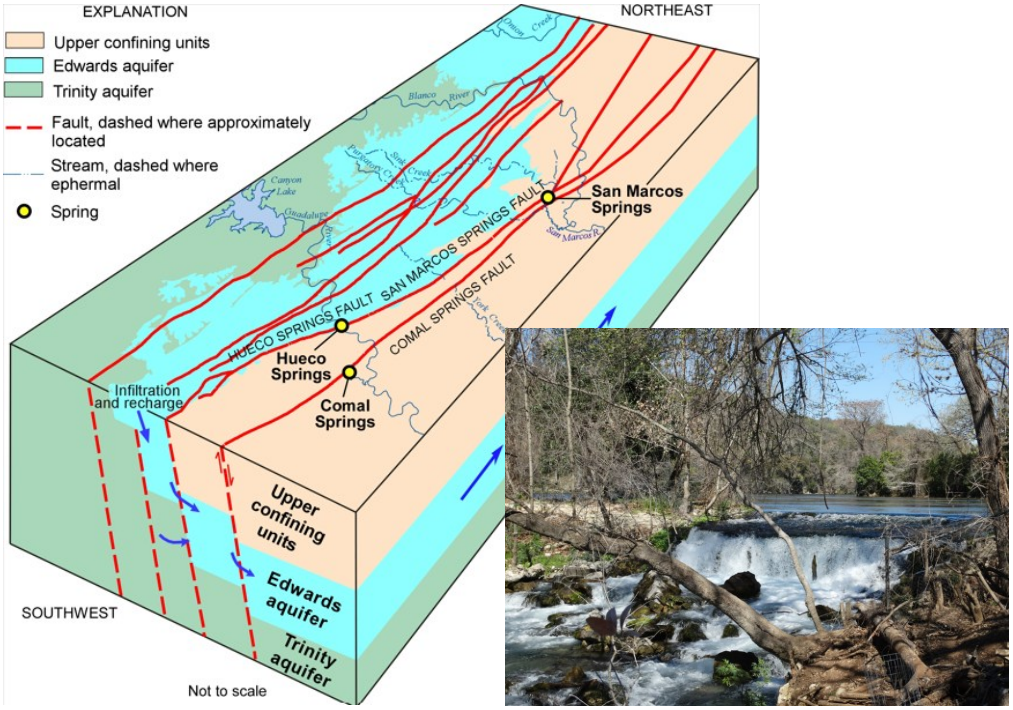


Figure 1.25 *Left*: Idealized block diagram of the Edwards aquifer in the vicinity of Comal, Hueco, and San Marcos Springs, south-central Texas. Modified from Musgrove and Crow, 2012. *Right*: Eastern overflow from the discharge basin of San Marcos Springs, Texas.



Figure 1.26 *Left*: Barton Springs Fault, Austin, Texas. On the left is Georgetown Formation, and on the right side is the Edwards Group. Photo by Brian Hunt (available at https://www.austingeosoc.org/geo-photos/). *Right*: Typical fault in the Balcones Fault Zone, Edwards Aquifer, Texas. Photo courtesy of Gregg Eckhardt.

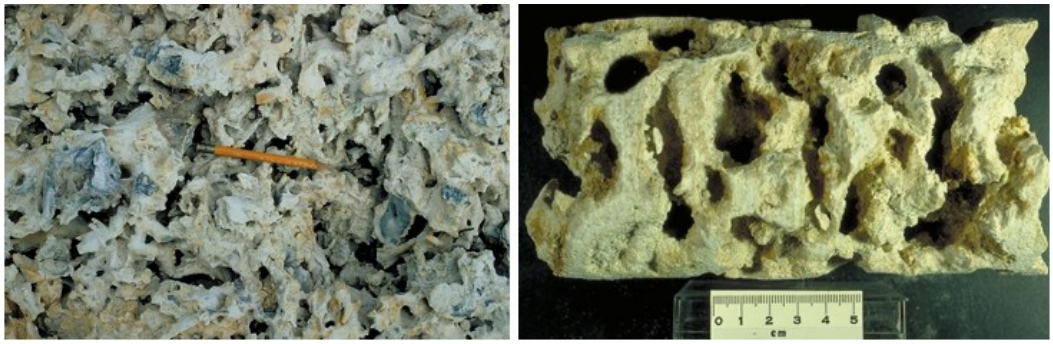


Figure 1.27 Young Florida limestones such as Miami oolite of the Biscayne aquifer have very high primary porosity (left photo) which, together with the dissolution-enhanced secondary ("vuggy") porosity (right photo), makes them one of the most prolific karst aquifers in the world. High-capacity municipal wells commonly produce more than 1,000 gpm with drawdown less than 1-2 feet. Photographs by George Sowers, printed with kind permission of Francis Sowers.

In young Eocene limestones such as those of Floridan and Biscayne Aquifers in Florida, development of cavities and preferential flow paths is seemingly chaotic and much less influenced by fractures and faults compared to old, consolidated Mesozoic and Paleozoic carbonates. This is illustrated in Figs. 1.27 and 1.28 where both the scale of drilling core samples and the cave images resemble sponge-like structures. Because of the very high primary porosity (porosity formed by lithogenesis, i.e., during rock formation), which is enhanced by dissolution, these karst aquifers are the most prolific in the United States.



Figure 1.28 *Left*: Map of the Weeki Wachee Springs Cave System, Florida. Courtesy of Karst Underwater Research. *Right*: Screenshot of a diving video at Radium Springs, Georgia, October 31, 2015, taken by Guy Bryant with divers Kelly Jessop and Peter Buzzacott. Note high "sponge" ("vuggy") porosity of the upper Eocene Ocala limestone due to dissolution. Video available on YouTube (CDroHO56F-Y)

One very important aspect of karstification is that the conduit/channel network continues to expand both vertically (Fig. 1.29) and laterally. Consequently, groundwater connections, i.e., flow directions, can often develop across topographic drainage divides when carbonate deposits are sufficiently thick and extensive. This is true for both local and regional scales of interest. The two most obvious results of this inconsistency between topographic and groundwater divides in karst terrains are sinking and dry streams (Figs. 1.31 through 1.33) and large springs.

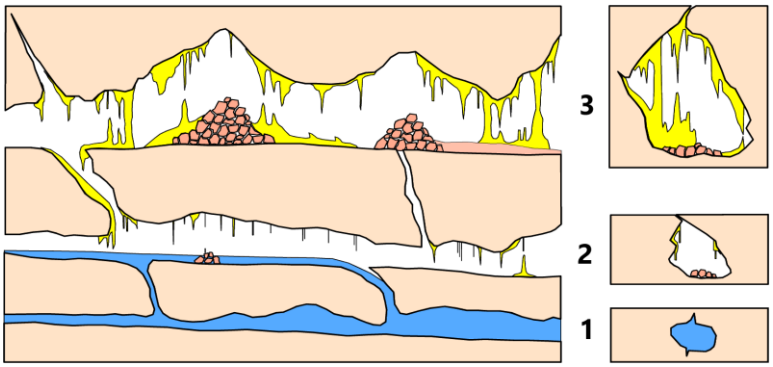


Figure 1.29 Main phases in cave development. (1) Initial stage: phreatic conduit being developed below water table by dissolution. (2) Intermediate stage: cave passage with permanent to intermittent flow on the floor, vertical erosion and dissolution of rock transfers most of the flow to the lower level; speleothems starting to form, occasional rock fall. (3) Last phase of cave development—permanently dry passages with a variety of abundant speleothems. From Kresic, 2013.



Figure 1.30 Relict of an initial tubular (phreatic) conduit in Mammoth Cave, Kentucky, now rarely active and reshaped by downcutting towards lowered karstification base. Photo courtesy of Nenad Marić.

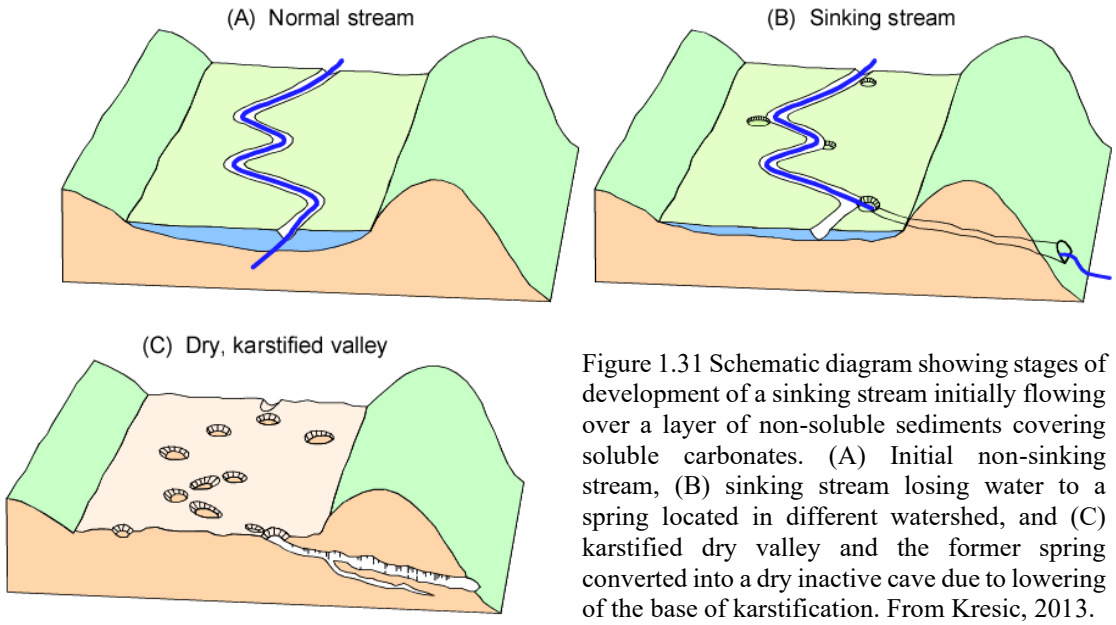


Figure 1.31 Schematic diagram showing stages of development of a sinking stream initially flowing over a layer of non-soluble sediments covering soluble carbonates. (A) Initial non-sinking stream, (B) sinking stream losing water to a spring located in different watershed, and (C) karstified dry valley and the former spring converted into a dry inactive cave due to lowering of the base of karstification. From Kresic, 2013.



Figure 1.32 *Left*: Recharge into Kiwi Sinkhole, near Driftwood, Texas. Photo courtesy of Brian Hunt. *Right*: Whirlpool over Crippled Crawfish Cave on the losing Onion Creek when flowing, near Austin, Texas. Photo courtesy of David Johns.

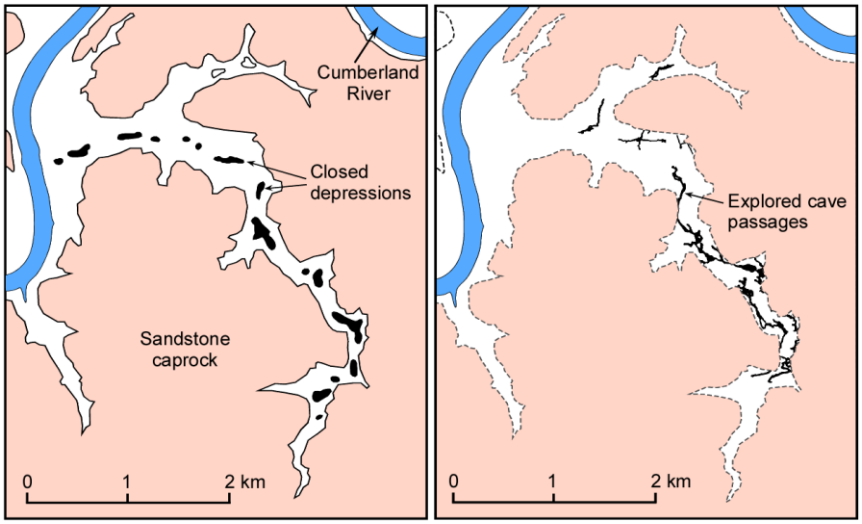


Figure 1.33 *Left*: Closed surface depressions in the Cave Creek dry valley, Pulaski County, north-central Kentucky. Stream erosion breached low-permeable caprock of sandstone and shale and exposed underlying limestone in which these depressions are developed. *Right*: Three-dimensional network of cave passages totaling 17 km in length closely follows the valley axis. Modified from Ewers, 1985; cave information provided courtesy of Louis Simpson.

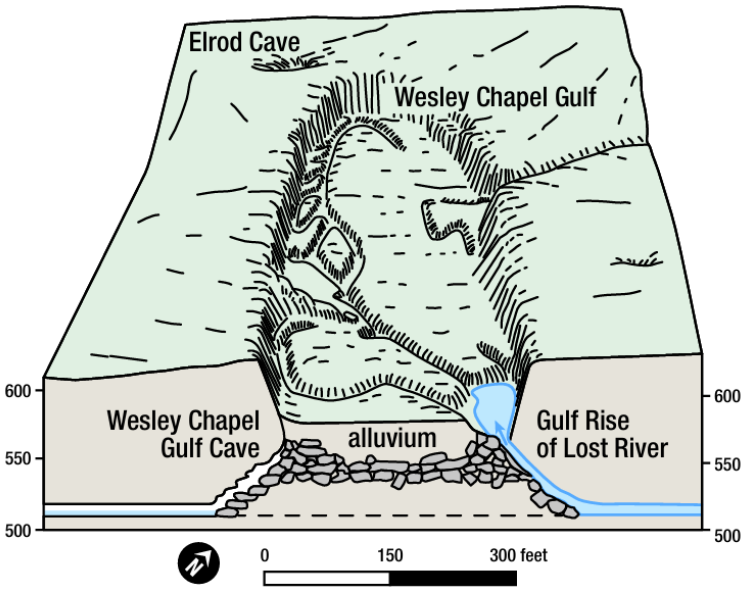


Figure 1.34 Idealized block diagram showing subterranean drainage at Wesley Chapel Gulf, Lost River watershed, Indiana. From *Lost River Field Trip for the Indiana Water Resources Association* by Richard L. Powell. Wesley Chapel Gulf is owned by the U.S. Forest Service and was declared a *National Natural Landmark* in 1972. The steep perimeter of this collapse sinkhole encloses about 8.3 acres and contains an alluvial floor of 6.1 acres. Nearly the entire floor of the gulf is an alluvial flood plain created by deposition of sediment from storm waters discharged from the Gulf Rise of Lost River located in the southeast end of the gulf. A submerged cave passage with a diameter of about 3 feet descends about 160 feet to a large passage about 10 feet high and 30 feet wide for a distance of at least 200 feet (S.D. Maegerlein, 1983, personal communication). During the time that the gulf was mapped in 1931, Malott (1932, p. 294) counted around 100 swallow holes in the floor of the gulf.

To emphasize, the three most important characteristics of groundwater flow in karst aquifers are: (1) natural groundwater divides often do not coincide with surface water (topographic) divides as a result of losing, sinking, and dry streams; (2) groundwater velocities in conduits could be extremely high, on the order of hundreds of meters or more per day; and (3) aquifer discharge is often localized through large karst springs, which have subsurface drainage areas commonly larger than the topographic ones.

A large permanent karst spring usually represents the end point of an integrated karst water system, developed both at the land surface and in the subsurface. Such a system is sometimes referred to as *springshed*, the term commonly used in Florida (Fig. 1.35).

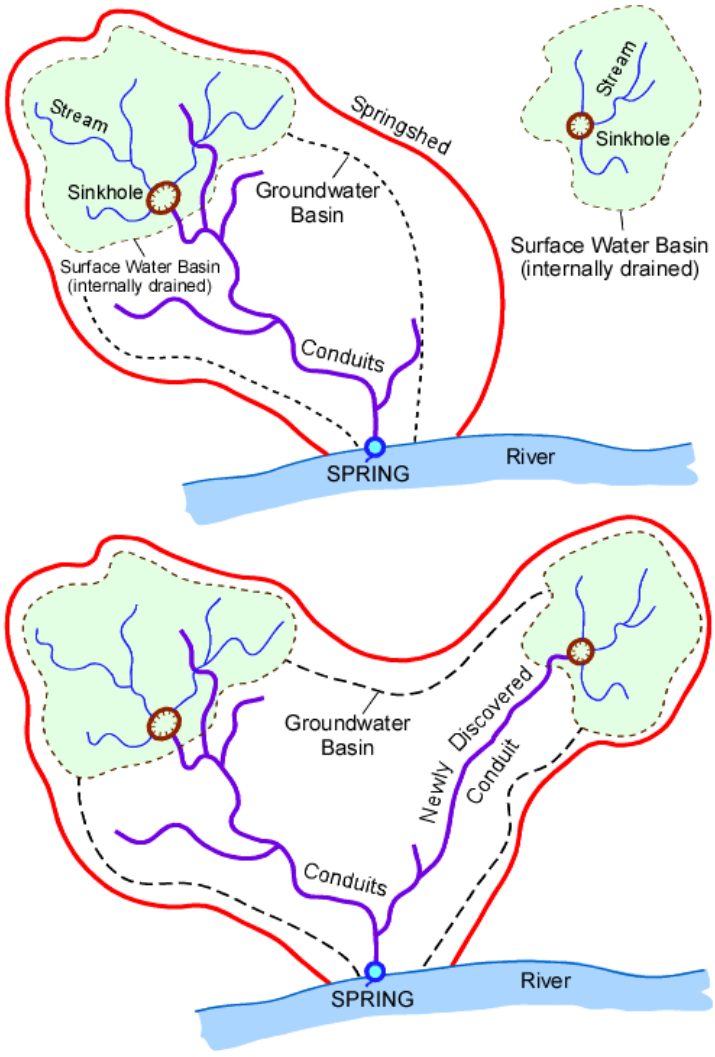


Figure 1.35 *Top*: Springshed boundary depicting surface water and groundwater basins. *Bottom*: Revised springshed boundary based on results of cave diving exploration and tracer tests that confirm the presence of conduits connecting a nearby surface water basin to the springshed. Modified from Arthur et al., 2002.

Dye tracing, the only direct method for determining drainage areas of karst springs and karst groundwater drainage areas in general, often results in unexpected connections (Fig. 1.36). These connections may vary in time depending on the season and major recharge events so that dye tracing tests must be repeated with different dyes and for different hydrologic conditions to accurately determine groundwater flow directions and delineate surface and subsurface drainage areas of interest. Assessing temporal variation of groundwater flow directions and hydraulic heads in the aquifer is a critical step for any project in karst for which a conceptual model is being developed.

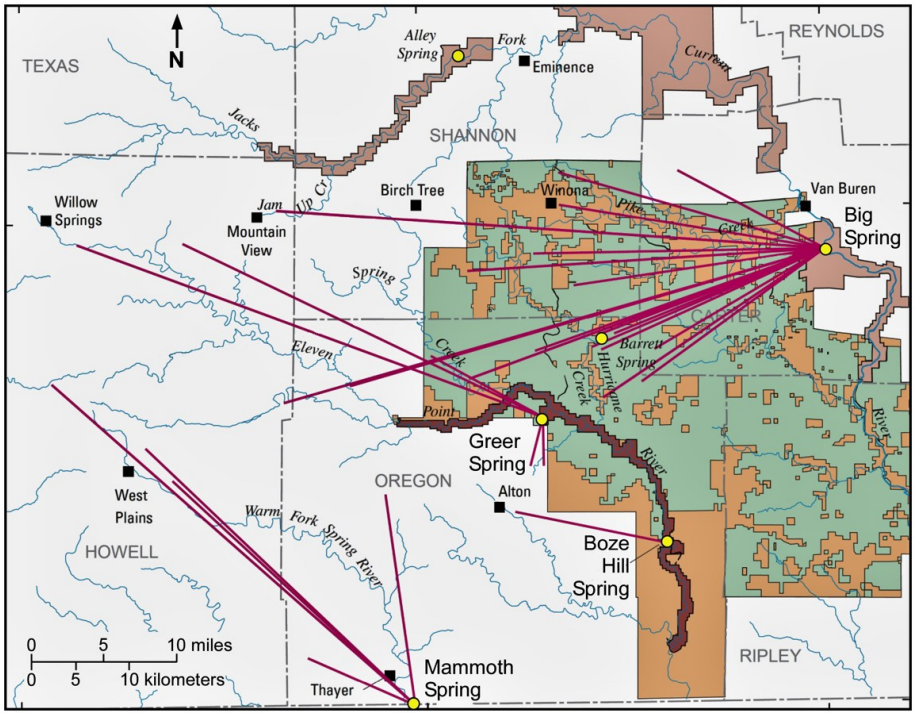


Figure 1.36 Results of dye-tracing investigations that successfully identified a hydrologic connection between a sinkhole or a losing stream reach and a major spring in south-central Missouri. Dye traces are in dark pink, different color polygons represent various land uses. Note how different dye traces cross different watersheds, sometimes intersect, and can be more than 30 miles long. Modified from Imes et al., 2007.

1.1.1 Sinkholes

Sinkholes, together with large springs and caves, are the most obvious signs that the underlying and host rocks are karstified. They can also reveal potential presence of preferential flowpaths in the aquifer below when occurring in linear patterns (see Fig. 1.37 and Fig. 1.1). Being closed depressions, sinkholes also present areas of focused, enhanced aquifer recharge as they interrupt overland precipitation runoff over larger areas and are internally drained downwards. This is one of the reasons why, compared to other aquifer types, karst aquifers generally have higher rates of recharge.

As discussed by Ann Tihansky of USGS (https://pubs.usgs.gov/circ/circ1182/pdf/15WCFlorida.pdf), new sinkholes have been correlated to land-use practices. Induced sinkholes are conceptually divided into two types: those resulting from groundwater pumping and those related to construction and development practices. Modified drainage and diverted surface water commonly accompany construction activities and can lead to focused infiltration of surface runoff, flooding, and erosion of sinkhole-prone earth materials. Manmade impoundments used to treat or store industrial- process water, sewage effluent, or runoff can also create a significant increase in the load bearing on the supporting geologic materials, causing sinkholes to form. Other construction activities that can induce sinkholes include the erection of structures, well drilling, dewatering foundations, and mining.

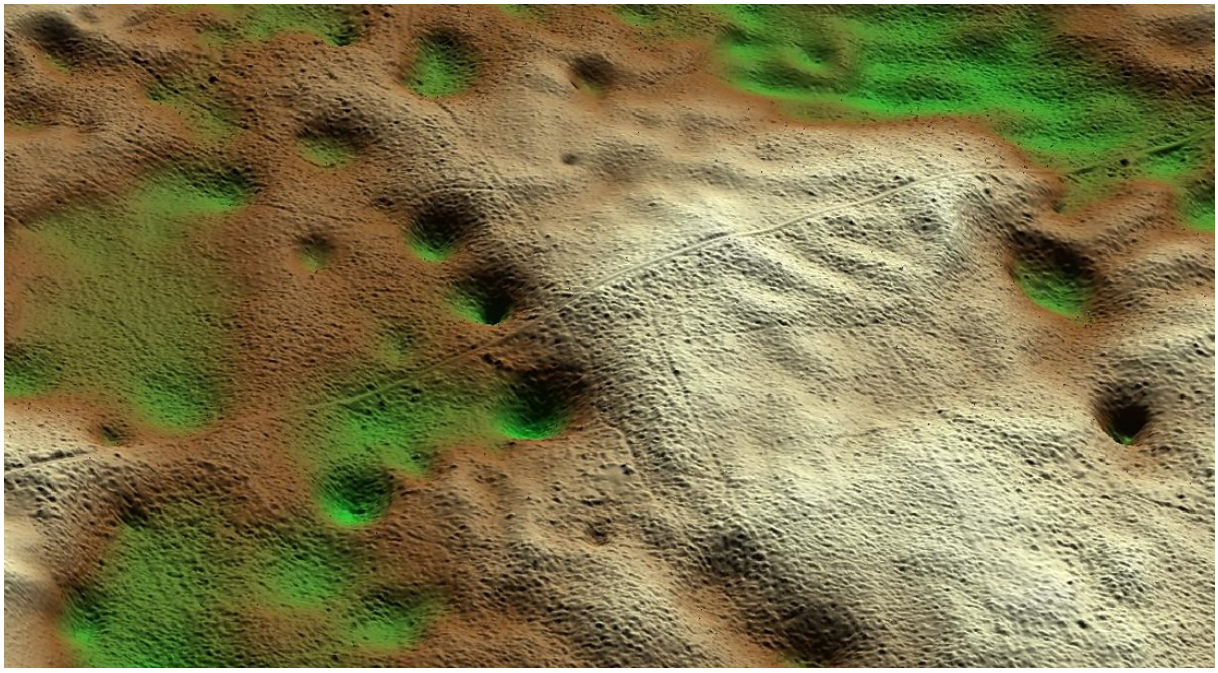


Figure 1.37 Digital elevation model created from high-resolution data collected by airborne LiDAR (Light Detection and Ranging) instrument. Lines of sinkholes like these near Ocala, Florida, developed in surficial sandy sediments, indicate possible preferential flow paths in the underlying karst aquifer. Note the extraordinary detail obtained by the LiDAR technology. Depth of the wider sinkhole by the road is about 20 ft (from Kresic, 2013).



Figure 1.38 Sinkholes in Monroe County, West Virginia. Photo courtesy of William Jones.

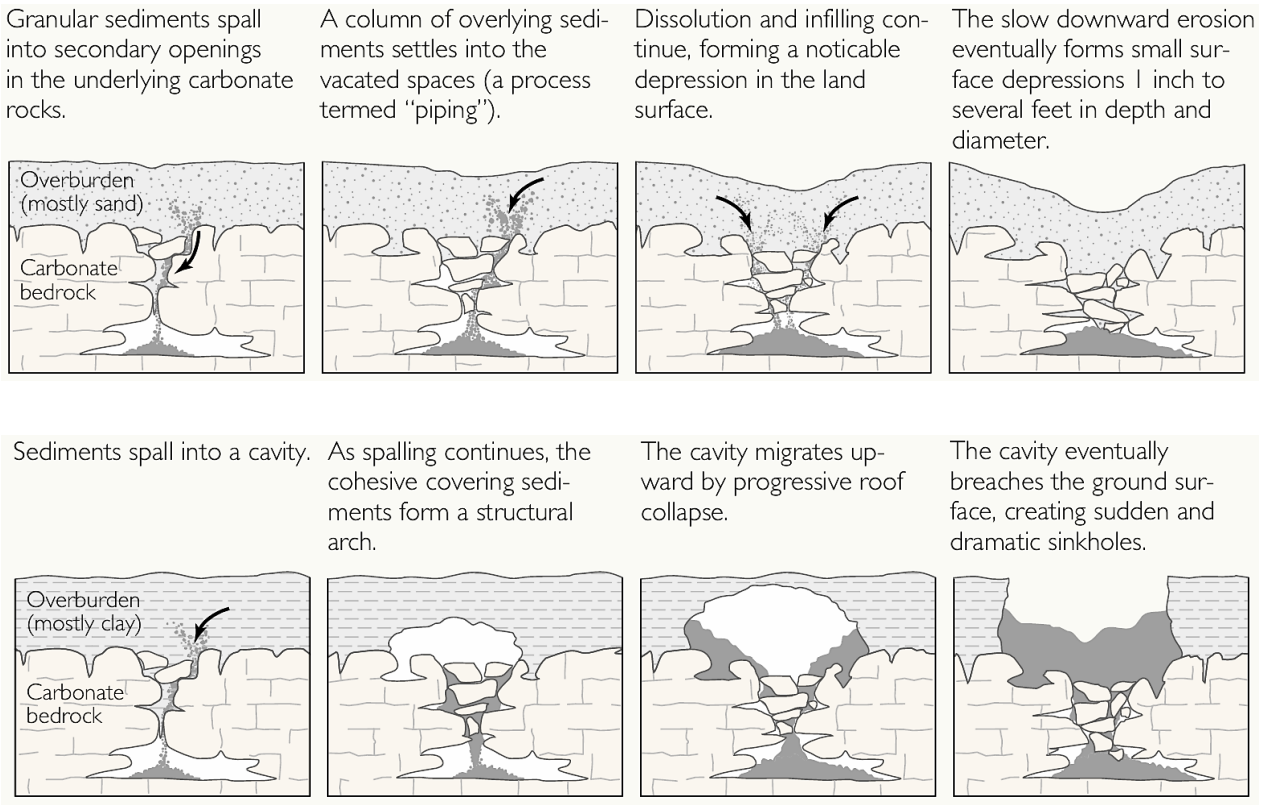


Figure 1.39 *Top*: Cover-subsidence sinkholes tend to develop gradually where the covering sediments are permeable and contain sand. In areas where cover material is thicker or sediments contain more clay, cover-subsidence sinkholes are relatively uncommon, are smaller, and may go undetected for long periods. *Bottom*: Cover-collapse sinkholes may develop abruptly (over a period of hours) and cause catastrophic damages. They occur where the covering sediments contain a significant amount of clay. Over time, surface drainage, erosion, and deposition of sediment transform the steep-walled sinkhole into a shallower bowl-shaped depression. From: *Sinkholes, West-Central Florida, A link between surface water and ground water* by Ann B. Tihansky, U.S. Geological Survey, Tampa, Florida https://pubs.usgs.gov/circ/circ1182/pdf/15WCFlorida.pdf

Excessive irrigation can also cause sinkhole development as the surficial sediments become oversaturated. The increased weight or load of the saturated sediments can contribute to water ponding by causing some subsidence. In turn, the increased load can trigger the development of cover-collapse sinkholes.

The overburden sediments that cover buried cavities in the aquifer systems are delicately balanced by groundwater fluid pressure. In sinkhole-prone areas, the lowering of groundwater levels, increasing the load at land surface, or some combination of the two, may contribute to structural failure and cause sinkholes.

Aggressive pumping can induce sinkholes by abruptly changing groundwater levels and disturbing the equilibrium between a buried cavity and the overlying earth materials. Rapid declines in water levels can cause a loss of fluid-pressure support, bringing more weight to bear on the soils and rocks spanning buried voids. As the stresses on these supporting materials increase, the roof may fail and the cavity may collapse, partially filling with the overburden material. Prior to water-level declines, incipient sinkholes are in a marginally stable stress equilibrium with the aquifer system. In addition to providing support, the presence of water increases the cohesion of sediments. When the water table is lowered, unconsolidated sediments may dry out and coarser-grained sediments, in particular, may move easily into openings (Tihansky, 1999).

Induced sinkholes are generally cover-collapse type sinkholes and tend to occur abruptly. They have been forming at increasing rates during the past several decades and pose potential hazards in developed and developing areas of west-central Florida. The increasing incidence of induced sinkholes is expected to continue as our demand for groundwater and land resources increases. Regional declines of ground-water levels increase sinkhole occurrence in sinkhole-prone regions. This becomes more apparent during the natural, recurring periods of low annual rainfall and drought (Tihansky, 1999).



Figure 1.40 A 40-foot sinkhole at the National Corvette Museum in Bowling Green, KY, swallowed eight of the sports cars on display inside the facility's iconic Sky Dome early Feb. 12, 2014. Photo courtesy of National Corvette Museum. https://www.corvettemuseum.org/sinkhole

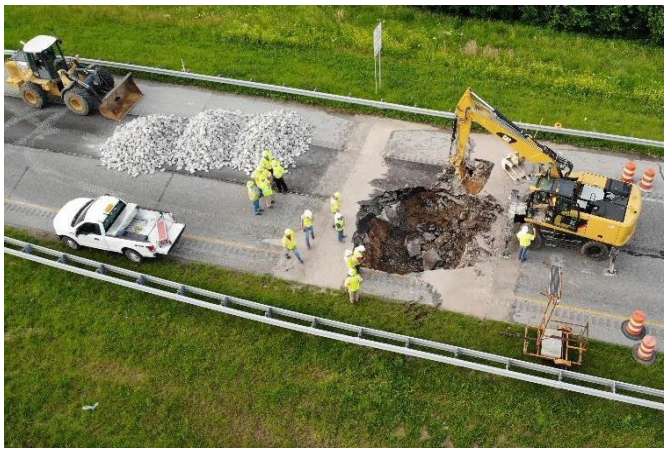


Figure 1.41 Sinkhole repair. From "Sinkhole closes I-265 East in southern Indiana for 12 hours" by Fallon Glick: *A sinkhole opened up on I-265 East near Charlestown Road….Two people were killed in the backup caused by the sinkhole*. Image credit: WDRB Sky Cam. (Sinkhole closes I-265 East in southern Indiana for 12 hours | Local News | wdrb.com)



Figure 1.42 Barry County sinkhole. Photo courtesy of Missouri Department of Natural Resources. Sinkholes | Missouri Department of Natural Resources

1.2 Types and Characteristics of Karst Springs

A spring is a location at the land surface where groundwater discharges from the aquifer, creating a visible flow. This discharge is caused by difference in the elevation of the hydraulic head ("groundwater level") in the aquifer and the elevation of the land surface where the discharge is taking place. The opening through which groundwater discharges is called spring orifice. Springs issuing from consolidated rocks usually have a well-defined orifice, as opposed to springs in unconsolidated sediments, although the latter ones may also have a clearly visible orifice.

A fracture (or fissure) spring refers to concentrated discharge of water from bedding planes, joints, cleavage, faults, and other breaks in the consolidated fractured rock. In karst, fracture springs are in many cases modified by dissolution and often referred to as cave springs (Fig. 1.43).

Secondary springs issue from locations located away from the primary spring discharge, which is covered by colluvium (rock fall fragments) or other natural debris and therefore not visible. Although karst is best known for its spectacular cave springs (Fig. 1.43-*Right*), the water seldom runs entirely from a single opening at any given location. Often near large cave springs, there are scattered smaller springs issuing from a variety of openings at different elevations, some of them active only after heavy rains.



Figure 1.43 *Left*: Fracture enlarged by dissolution at the Montezuma Well outlet spring in Arizona. *Right*: Rick's Springs in Wasatch-Cache National Forest, 15.7 miles from Logan, Utah, up Logan Canyon Scenic Byway (Utah 89). Photo courtesy of US Forest Service.

What is called a single spring in one locality may be equivalent to what in another locality is regarded as a group of springs, each of which has an individual name. Some springs are designated in the singular and some in the plural—for example, Silver Spring (such as main headspring of what is now called Silver Springs in Florida) and Comal Springs in Texas. The idea that underlies this usage is that if the water issues from a single opening or from several openings that are close together, it forms a spring, whereas if it issues from a number of openings that are farther apart, it forms springs. However, local terminology is so variable in this respect that there is no consistent distinction between "a spring" and "springs" and often there is no uniformity in usage of terms even for the same group of openings (Meinzer, 1927).

All springs (except some associated with young volcanism and hydrothermal activity where discharge is driven by gases and temperature gradients) ultimately discharge at the surface because of the force of gravity. Regardless of the rock type, they are divided into two main groups based on the nature of the hydraulic head ("aquifer water pressure") in the underlying aquifer at the point of discharge:

- **Gravity springs** emerge under unconfined conditions where the water table intersects the land surface. In the case of well-defined cave springs in karst, this is equivalent to a free flowing, predominantly horizontal or gently sloping cave, where the cross-sectional perimeter of the cave is not fully submerged most of the time. Such springs are also called descending springs (e.g., see Figs. 1.44 and 1.45).
- **Artesian springs** discharge under pressure due to confined conditions in the underlying aquifer. They are also called ascending or rising springs. The spring orifice is completely submerged, and so are the karst cavities (conduits) supplying water to the springs (Fig. 1.46). The main feeder conduit(s) of these springs is often vertical or quite steep and, if visible and large enough, presents a continuing challenge to brave cave divers who want to see where it goes. Ascending springs are also called **vauclusian springs**, named after Fontaine de Vaucluse in southern France (Fig. 1.47).

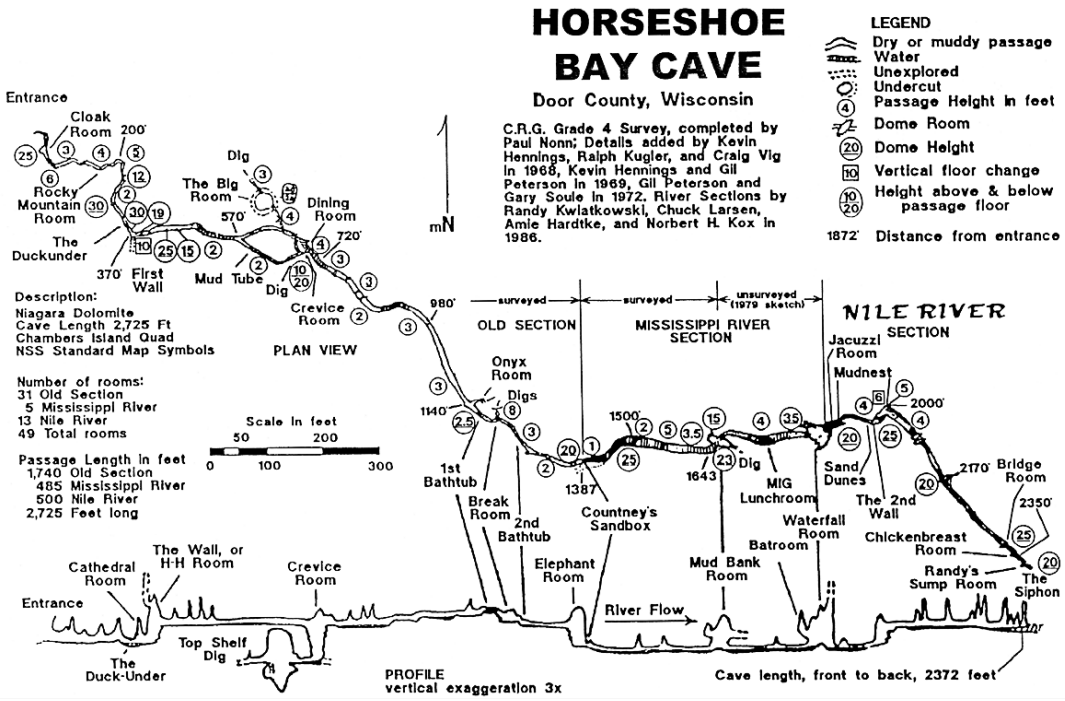


Figure 1.44 Map and profile (bottom) of Horseshoe Cave, a simple flow-through cave. There are both dry passages and passages with water flowing after heavy rain and snowmelt. Horseshoe Bay Cave is one of the longest caves in northeastern Wisconsin, located near the shore of Green Bay in Door County. Courtesy of Wisconsin Speleological Society.



Figure 1.45 Entrance to Horseshoe Bay Cave in Wisconsin. The cave's existence has been known since at least 1896, when two hunters noticed water emerging from the bluff. From 1959 to 1963, cavers explored and mapped 1,740 ft (530 m) of passage, much of it low, wet, and cold. In 1978 cavers dug through a sand and gravel filled crawlway at the end of the mapped passage and entered a watery stream crawlway leading to about another thousand feet of passage. Cave divers have gone even farther. The cave is unusual in Wisconsin for containing flowing water—most Wisconsin caves are high above the current water table and developed before valleys were formed by erosion. Many "rooms" in the cave are formed along the passage by solutionally enlarged joints. Occasionally water flows out of the cave, when snow melts rapidly enough. In September 2014 heavy rain caused the entrance to flood. Photo by Eric Aleson, from Zachariasen, 2016.

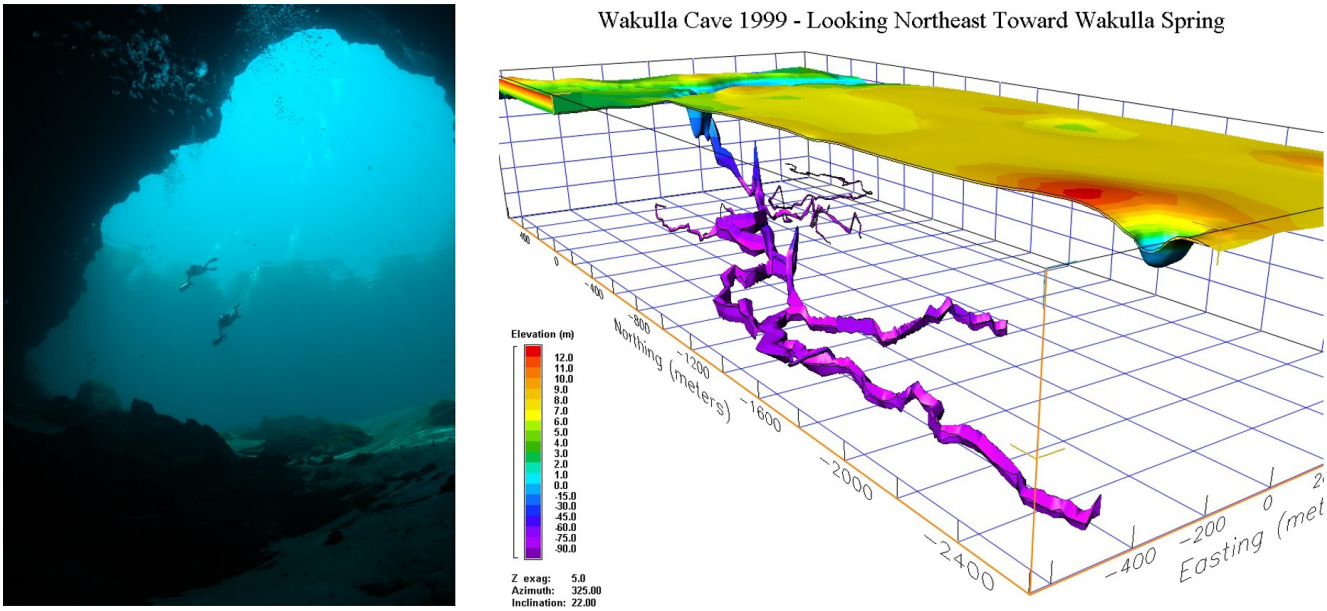


Figure 1.46 *Left*: View of the Wakulla Spring outlet from a cave diver's perspective. This first magnitude spring in Wakulla County, Florida, one of the two largest in the country (together with Big Spring in Missouri) has miles of explored submerged cave passages. The spring and the cave are formed in Oligocene-Miocene limestones with very high, primary intergranular porosity of up to 55% (e.g., see Brooks, 1967). Several of the early *Tarzan* movies starring Johnny Weissmuller were filmed in the Wakulla Springs State Park. Photo copyright David Rhea & Global Underwater Explorers 2006, printed with permission. *Right*: Computer-generated view of the submerged Wakulla Cave passages based on the cave diving surveys. The passages all converge toward the Wakulla Spring which is one of only three springs in the World included on the list of World Geoheritage Sites by the International Union of Geological Sciences The other two are also karst springs, Buna Spring in Bosnia and Herzegovina (Fig. 1.7), and Fontaine de Vaucluse in France (Fig. 1.47). 3D view courtesy of Dr. Todd Kincaid.





Figure 1.47 *Left*: Upper outlet of the Fontaine de Vaucluse during high discharge (64.5 m^3/s or 2278 cfs). Photo courtesy of E. Simon, available at https://iugs-geoheritage.org/geoheritage_sites/fontaine-de-vaucluse/. *Right*: The main outlet is dry when the spring discharges only at a lower level, through colluvium in a narrow valley downgradient. Photo available at https://uk.islesurlasorguetourisme.com/page/when-to-come-see-the-source+5638.html. The main steep/vertical feeder conduit (Fig. 1.49) is explored to depth of 315 meters (1033 ft) by *Spéléonaute*, a remotely operated vehicle (ROV). The spring is considered *locus typicus* (prototype) for all such springs worldwide ("vauclusian type spring"). The spring discharge has been recorded since 1878: the average is 17.3 m^3/s (611 cfs), the minimum is 3.1 m^3/s (109.5 cfs), while the maximum exceeds 80 m^3/s (2825 cfs). Photo courtesy of E. Simon, available at https://iugs-geoheritage.org/geoheritage_sites/fontaine-de-vaucluse/.

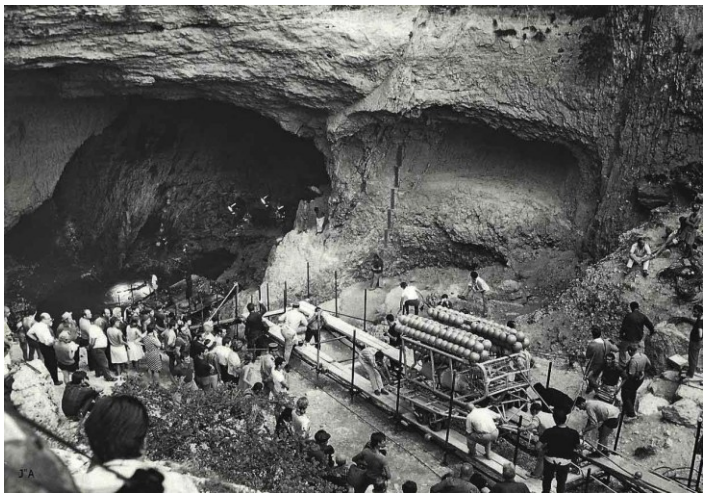


Figure 1.48 Launching of *Telenaute* ROV, 1967. Photo by Jacques-Yves Cousteau. Courtesy of Syndicat Mixte du Bassin des Sorgues.

Virtually all first magnitude karst springs in the United States (springs with average discharge rate higher than 100 cubic feet per second, cfs) and almost all of the largest karst springs in the world are vauclusian. Their drainage areas extend well beyond the immediate topographic divides, and the thickness and therefore the volume of the aquifers (saturated zone below water table) that supply water to these springs are very large. For example, the drainage area of Fontaine de Vaucluse is estimated at 1,160 km^2 (448 mi^2) while the saturated zone is up to 700 m (2297 ft) thick.

Many large and small deep vauclusian springs worldwide are called "Blue Spring", "Pozo Azul" (Spanish), or "Puits Bleu" (French). Their blue color is due to the selective absorption of light by deep, clear, calcium carbonate-rich water. As sunlight hits the water, dissolved microscopic mineral particles in the water absorb longer, warmer

wavelengths (reds, oranges, and yellows) and scatter shorter wavelengths (blues and greens), which are then reflected to our eyes (Figs. 1.50 and 1.51).

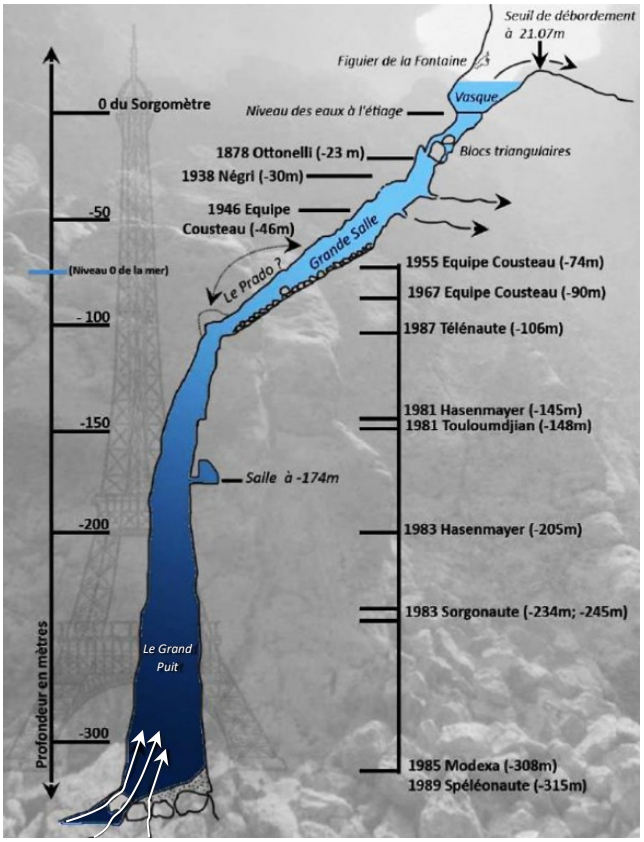


Figure 1.49 Vertical cross-section of Fontaine de Vaucluse, explored to depth of 315 m (1033 ft) in 1989 using *Spéléonaute,* a remotely operated vehicle (ROV*)*. Note the silhouette of Eiffel Tower on the left for scale. Courtesy of Direction Régionale de l'Environnement, de l'Aménagement et du Logement. https://www.fontainedevaucluse.fr/le-gouffre,-la-sorgue



Figure 1.50 Examples of "Blue Holes", ascending karst springs in the Valley and Ridge karst of the Appalachians. *Left*: Ellisons Blue Hole, Crockford-Pigeon Mountain Wildlife Management Area, Walker County, Georgia. *Right*: McBroom Branch Blue Hole, Putnam County, Tennessee. Photographs courtesy of Chuck Sutherland**.**

In many instances, a karst spring represents a combination of the two basic types because of the nature of karst: the karstification process constantly creates new solution cavities (conduits) both vertically and laterally as explained earlier. What ones used to be a "simple" gravity cave spring can evolve into a dry cave spring hanging above the incising erosional level (usually a major stream), while new, lower spring feeder conduits are being

developed trying to catch up with the stream incision. If geological and structural conditions permit, conduits may keep developing deeper below the stream level, and the spring may eventually turn into an ascending one, sometimes discharging directly into the stream without a visible orifice (Fig. 1.52; see also Fig. 1.2).



Figure 1.51 Round Spring in The Ozark National Scenic Riverways, Missouri.

Provided there is enough areal extent and thickness of soluble carbonates, the conduit network feeding the spring will keep expanding including possible underground piracy—the large expanding spring may divert water from the smaller ones which eventually may cease to flow altogether ("big fish eats little fish".)



Figure 1.52 Boiling Spring on Potts Creek, Alleghany County, Virginia. Photo courtesy of Phil Lucas.

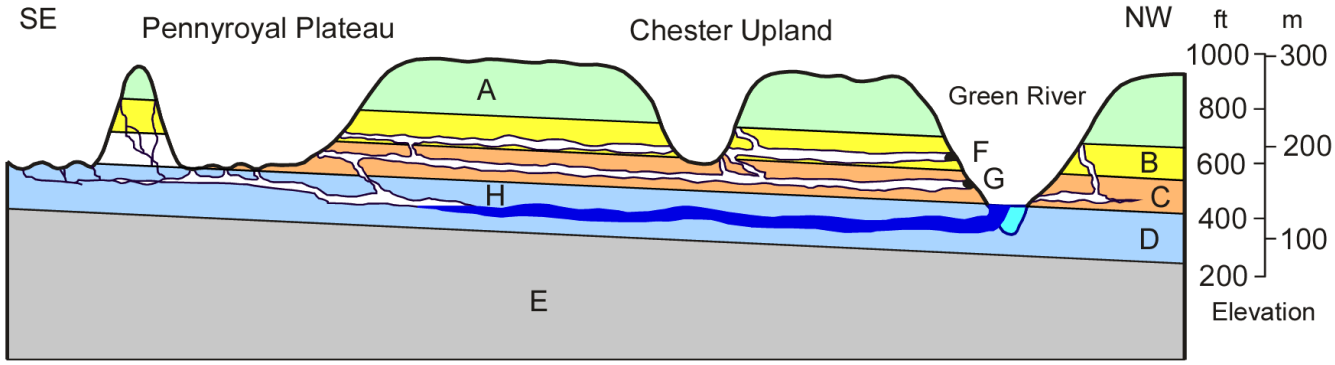


Figure 1.53 Generalized profile through Mammoth Cave (from Palmer, 1985). A=caprock of interbedded sandstone, shale, and limestone (Chesterian Series); B=Girkin Formation; C=Ste. Genevieve Limestone; D=St. Louis Limestone; E=impure limestone and detrital rocks; F=upper cave levels of late Tertiary age; G, H=typical levels of Quaternary age (partial flooding of the lowest level is due to late Pleistocene alleviation of the Green River).

The following points from an educational panel by the Big Spring in Missouri (Fig. 1.54) best illustrate the ever-expanding conduit networks of karst springs. During an average day, Big Spring carries in solution about 173 tons of dissolved dolomite rock. Therefore, in a year's time enough bedrock is removed to form a cave passage 11 feet high by 12 feet high and 1 mile long. Of course, this material is not removed as a unit but is distributed throughout the ramifying system of solution channels that feed the spring. The Big Spring system is continuously

getting larger simply because each year the feeder system grows larger by the removal of rock in groundwater solution and, each year, more rock is exposed to the dissolving action of groundwater.

One piece of practical advice for a novice karst hydrogeologist is that, if a spring is large, behaves like a karst spring (it has notable fluctuations), and appears to have a very significant deficit in its topographic (land surface) drainage area, then it is most likely a karst spring. In other words, even when such a spring discharges from a nonkarstic rock, but there are carbonates somewhere in the vicinity, one should always consider the possibility that the primary aquifer providing water to the spring is a karst aquifer. A very good example is Giant Spring in Great Falls, Montana. This first magnitude spring, one of the largest in the country, was once listed in the Guinness Book of World Records as the world's shortest river called the Roe River that flows into the Missouri River. The spring waters originate as precipitation that falls on the Madison Limestone of Little Belt Mountains over 4,000 ft above the Missouri River. The limestone has all the characteristics of karst, including losing streams and dry caves. At Giant Springs, the Madison Limestone is less than 400 ft below the surface and groundwater from this karstic aquifer escapes upward through a likely fault zone and fractures in the overlying Kootenai sandstone (Fig. 1.55).

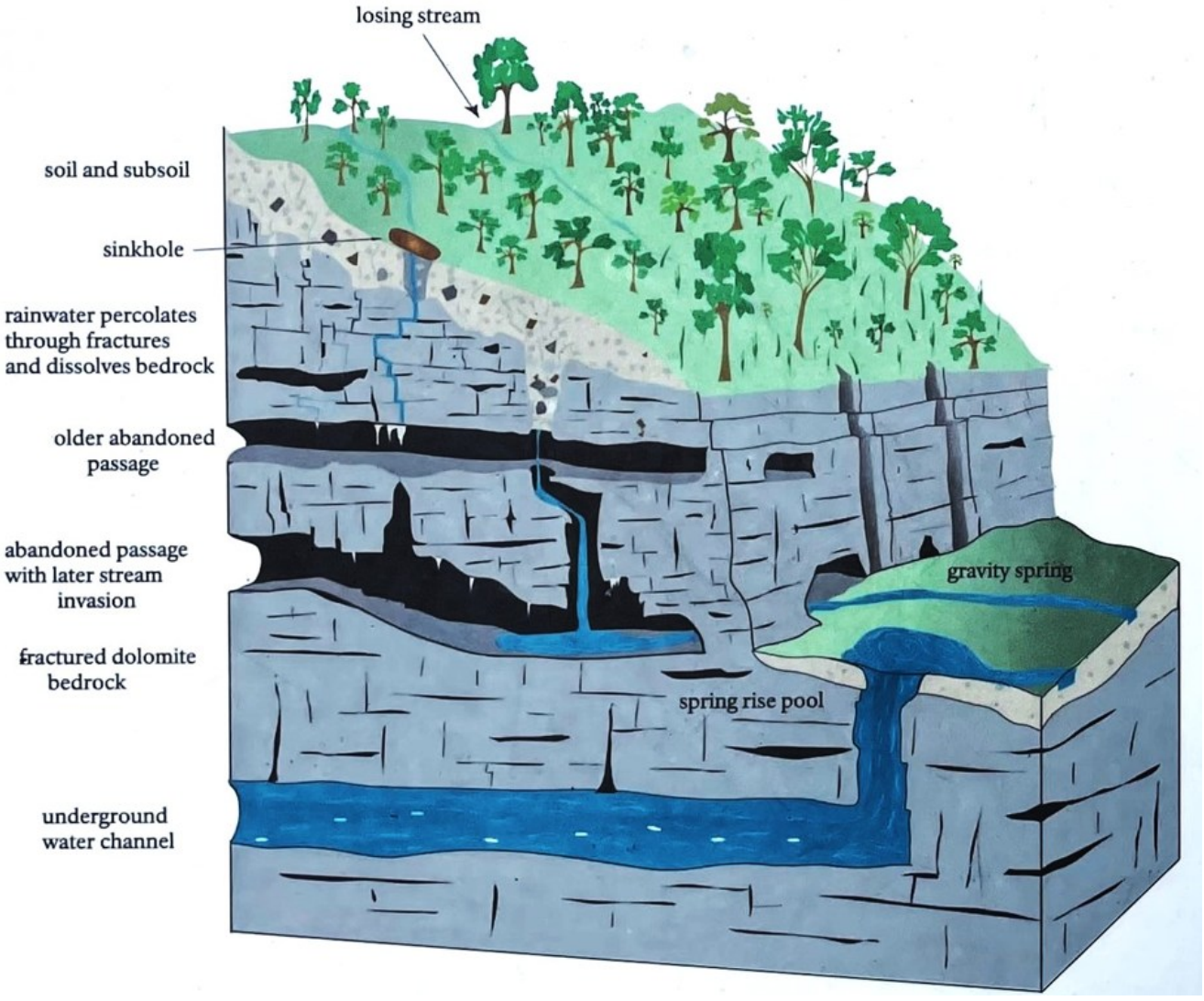
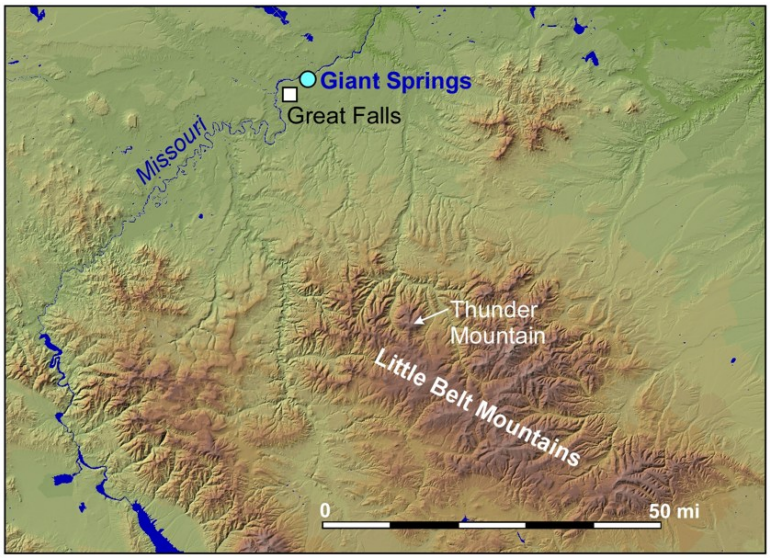
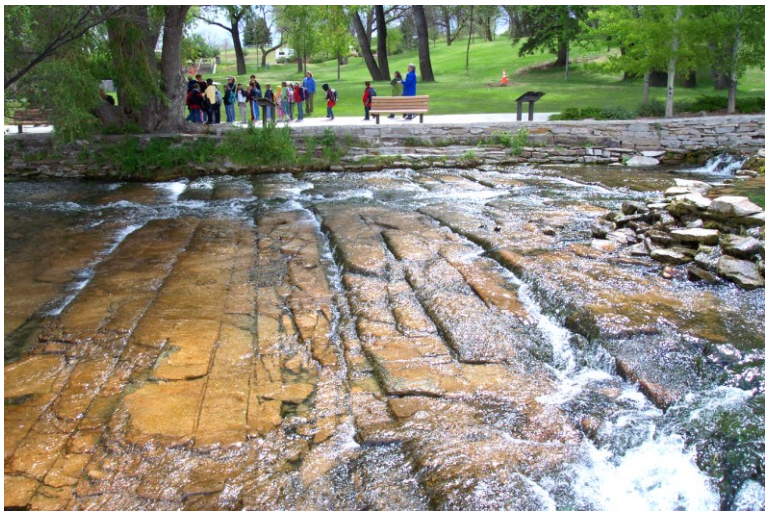


Figure 1.54 Part of a National Park Service informational panel by the Big Spring in Missouri, one of two largest individual karst springs in the country.





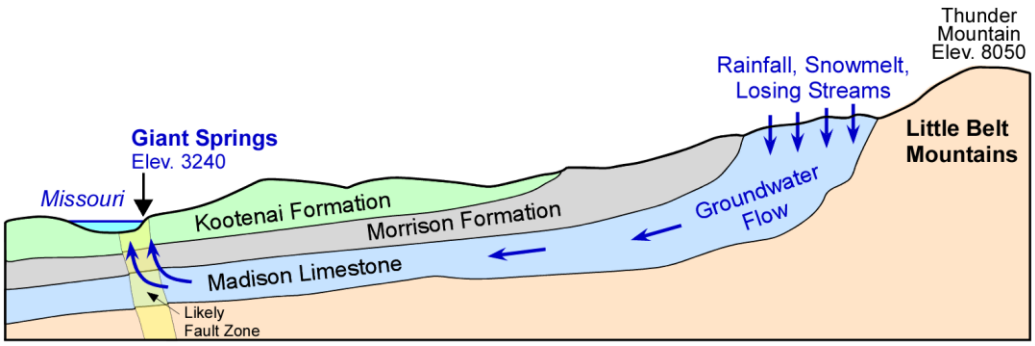


Figure 1.55 *Top*: Main recharge zone for the portion of karstic Madison Limestone Aquifer in Montana, drained by Giant Springs, is in its outcrop areas on Little Belt Mountains *Middle*: The vertical fractures in Kootenai Sandstone at Giant Springs State Park near Great Falls, Montana. Giant Springs is located on the crest of the Sweetgrass Arch—a gentle fold that was tectonically active during deposition of Paleozoic and Mesozoic sediments. Madison Limestone is only 400 feet below the surface. The fractures are extension fractures and are open. They provide an easy conduit for groundwater in the Madison aquifer to rise to the surface and be discharged in and next to the Missouri River. Text and photo courtesy of Dr. David Baker. *Bottom*: Schematic cross-section between Giant Springs and its recharge area in Little Belt Mountains. Modified from a schematic provided by Giant Springs Bottled Water Company.

Estavelle is a spring with a dual function: it acts as a spring during high hydraulic heads in the aquifer, and as a surface water sink during periods when the hydraulic head in the aquifer is lower than in the body of surface water (Fig.1.56). Estavelles are located within or adjacent to surface water features.

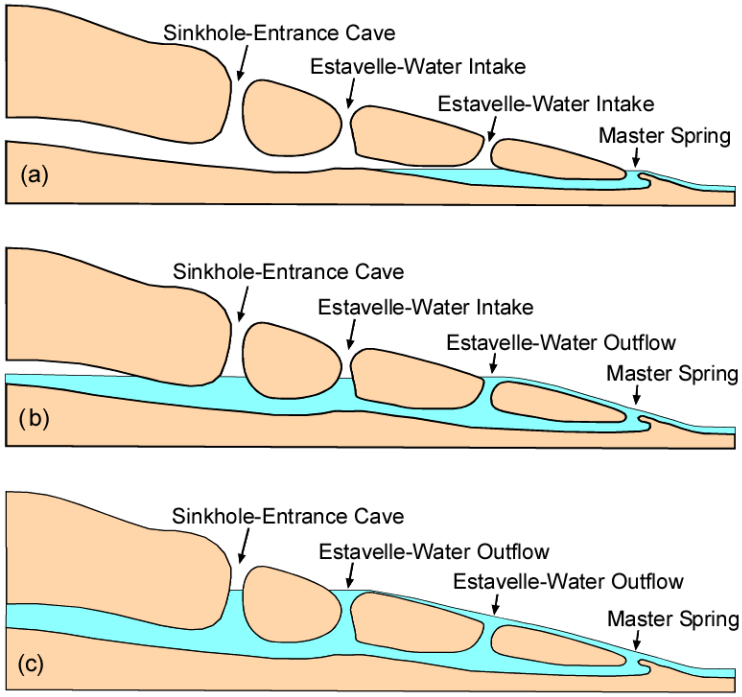


Figure 1.56 Hypothetical stream profile showing sequential reversal of flow in estavelles. (A) Low flow: only lower sections of karst drainage system flooded; estavelles are sinks (swallow holes); (B) medium flow: headward flooding of karst drainage system; lowest estavelle becomes spring (rising); and (C) High flow: storage capacity of karst drainage system exceeded; estavelles become springs, water backs up in headward caves. From Vineyard and Feder, 1974, 1982.

Discharge of *ebb-and-flow springs*, or *periodic springs*, occurs in relatively uniform time intervals (periods) and is explained by the existence of a siphon in the rock mass behind the spring (Fig. 1.57). The siphon fills up and empties with regularity, unrelated to the recharge (rainfall) pattern. The best-known example is Periodic Spring in the Bridger-Teton National Forest near Afton, Wyoming (see also Chapter 3.3).

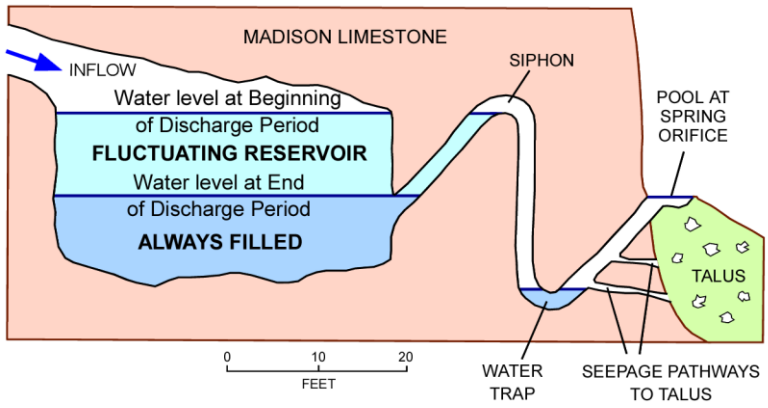


Figure 1.57 Siphon model of the Periodic Spring reservoir. When the reservoir water level reaches the siphon level, the flow commences. Flow continues until the reservoir level drops to base level. From Blanchard et al., 1990; modified from Huntoon and Coogan, 1987.

Karst springs typically have the highest amplitude (oscillation) of discharge rate compared to springs in other aquifer types. This is because of rapid aquifer recharge from precipitation and snowmelt episodes due to karstified rocks, and high transmissive rates of fractures and conduit networks (Fig. 1.58).

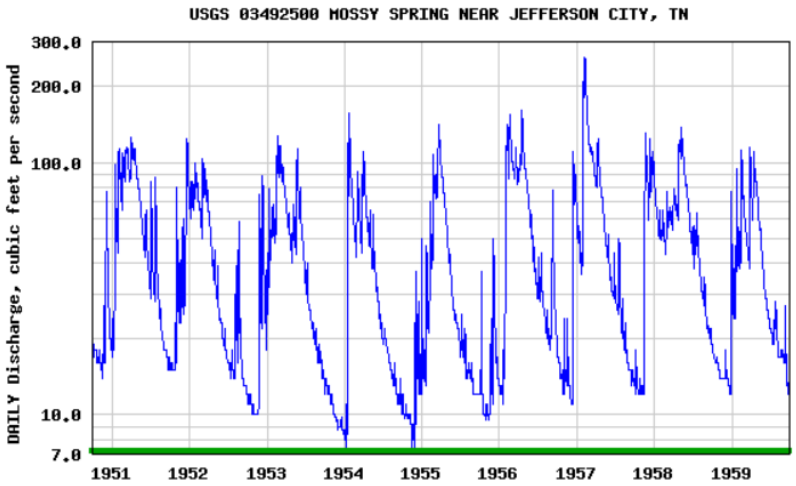


Figure 1.58 Daily discharge rate of Mossy Spring near Jefferson City, the largest spring in Tennessee, for the period of record. Average is 43.8 cfs, minimum is 7.4 cfs, maximum is 261 cfs. Index of Variability is 35.3.

Meinzer's classification of springs based on average discharge (Fig. 1.59) is widely used by the USGS and state agencies. Below are excerpts from the Meinzer publication illustrating the importance of evaluating and classifying springs based on their average discharge (on page 2):

Proposed classification of springs according to discharge

Magnitude	Average discharge
First	100 second-feet or more.
Second	10 to 100 second-feet.
Third	1 to 10 second-feet.
Fourth	100 gallons a minute to 1 second-foot (448 gallons a minute).
Fifth	10 to 100 gallons a minute.
Sixth	1 to 10 gallons a minute.
Seventh	1 pint to 1 gallon a minute. About 200 to 1,500 gallons, or 5 to 40 barrels, a day.
Eighth	Less than 1 pint a minute. Less than about 200 gallons, or 5 barrels, a day.

Figure 1.59 Photo of Oscar Meinzer's table on page 3 of the USGS Water Supply Paper 557 published in 1927. Note that 1 second-feet (cfs) = 449 gallons per minute = 646,000 gallons per day = 28.3 liters per second.

Thus the discharge of a spring can be determined with fair accuracy only by establishing a gaging station and obtaining a continuous record over a period of years. To compare a single measurement of one spring with a single measurement of another spring made at an entirely different time may give a result that is almost as incorrect as would be obtained by a similar comparison of surface streams. The question also arises as to whether a comparison is made according to the minimum, maximum, or average discharge. Thus, if the average discharge of a nearly constant spring is about the same as that of a spring which fluctuates greatly, the constant spring will rank higher with respect to minimum discharge and the fluctuating spring higher with respect to maximum discharge. There is some popular interest in the maximum discharge, but the value of a spring, whether it is used for a public supply, irrigation, power, or other purpose, generally depends more nearly on its minimum discharge, or on its discharge at the time when the most water is needed. In so far as comparisons are attempted in the following discussion, they are based on the average discharge.

As far as this author can tell, it appears that the Florida Geological Survey (FGS) is the only agency in the country that chose not to follow Meinzer's classification of springs based on their long-term average discharge (see Kresic, 2024 for a detailed analysis.) For example, FGS states that *One discharge measurement is enough to place a spring into one of the eight magnitude categories* (Scott et al., 2004. *Springs of Florida*, FGS Bulletin No. 66; emphasis added). This is just one of the reasons why the number of first magnitude springs in Florida grew from 11 (Meinzer's 1927 publication entitled Large Springs in the United States), to 17 in 1947, when Ferguson et al. published their Bulletin No. 31, also entitled *Springs of Florida,* and finally to 33 first magnitude springs in 2004, after various rules and practices in Florida changed.

As emphasized by Meinzer, the classification based solely on average spring discharge, without specifying other discharge parameters, is not very useful when evaluating the potential for spring utilization. For example, a spring may have a high average discharge, but it may be dry or just trickling most of the year. This discharge variability is typically higher in older carbonate rocks with low matrix porosity which cannot store significant volumes of groundwater that would provide relatively sustained flow between rainfall and snowmelt (recharge) episodes. The practice in most countries is that the springs are evaluated based on the minimum discharge recorded over a long period, typically longer than several hydrologic years (hydrologic year is defined as spanning all wet and dry seasons within a full annual cycle). The simplest measure of variability is the ratio of the maximum and minimum discharge, called Index of Variability:

$$I_v = \frac{Q_{max}}{Q_{min}}$$

Springs with the index of variability (I_v) greater than 10 are considered highly variable (see Fig. 1.58), and those with $I_v < 2$ are sometime called *constant* or *steady springs*. Meinzer proposed the following measure of variability expressed in percentage:

$$V = \frac{Q_{max} - Q_{min}}{Q_{av}} \times 100(\%)$$

where Q_{max}, Q_{min} and Q_{av} are maximum, minimum, and average discharge respectively. Based on this equation, a constant spring would have variability less than 25 percent, and a variable spring would have variability greater than 100 percent.

As discussed throughout this book and many other publications on karst, karst aquifers and springs are particularly vulnerable to both natural and anthropogenic contamination. The following discussion by Wolter et al. (2011) nicely echoes this common knowledge:

Losing streams occur where the local water table has dropped below the level of a stream bed, allowing water to flow downward. Karst areas are commonly typified by losing streams, as the development of conduit-flow systems tends to lower the water table in the aquifer, and the fractured and dissolved nature of the aquifer allows for ready downward leakage of surface water. In areas of karst development, some losing streams end in clearly visible sinkholes, while others tend to lose water more diffusely into the stream bed. Under relatively low flow conditions in many karst settings, total loss of stream flow occurs. The varying ways water enters and travels through karst groundwater systems has a profound effect on the quality of the recharge water, and hence on the quality of groundwater discharging back to the surface. Following rainfall or snowmelt, runoff to sinkholes and losing streams deliver recharge water with relatively high concentrations of sediment (see Fig. 1.56), *sediment-attached nutrients such as phosphorus and ammonia-nitrogen, herbicides, organic matter, and bacteria. In contrast, infiltration recharge to the aquifer, passing through the relatively thin cover of soil and surficial materials, delivers relatively high concentrations of soluble, non-adsorbing contaminants, in particular nitrate.*



Figure 1.60 Aqua Spring has the largest flow rate and the largest variation of flow of all springs in Bullpasture Gorge, Virginia. *Left*: Average flow conditions. *Right*: High flow rate after heavy rainfall carrying a heavy sediment load. The boils or mounds of water above the spring are more than 10 feet high. Photographs courtesy of Phil Lukas.

The following excerpts from a report prepared by the USGS in 2014, entitled *Water Quality in the Principal Aquifers of the Piedmont, Blue Ridge, and Valley and Ridge Regions, Eastern United States, 1993–2009* (Lindsey et al., 2014) illustrate well the general vulnerability of karst aquifers and springs to anthropogenic contamination, not just in the East.

Concentrations of nitrate in the carbonate-rock aquifers were among the highest in the Nation.

One of the reasons that the carbonate-rock aquifers are susceptible to nitrate contamination is that they have high-permeability karst features, such as sinkholes and conduits, which allow contaminants from the surface to easily enter the aquifer and to rapidly move through them. As a consequence, much of the water that reaches a well is very young. In addition, the presence of sinkholes favors oxic conditions, making denitrification unlikely. The combination of direct infiltration of contaminants into the aquifer, rapid movement through the aquifer, and oxic conditions within the aquifer, coupled with large nitrogen inputs, explains the high nitrate concentrations in samples from the carbonate-rock aquifers

In the Piedmont, Blue Ridge, and Valley and Ridge regions, total coliform bacteria were detected in all samples from springs and nearly 60 percent of samples from wells. E. coli, an indicator of fecal contamination, was detected in 91 percent of samples from springs and 17 percent of samples from wells. Although detections were frequent in all aquifers and settings, the carbonate-rock aquifers had a higher frequency of detection for E. coli and total coliform, as did areas overlain by agricultural land use.

At least one pesticide was detected in 55 percent of samples, and multiple pesticide compounds were commonly detected in a single sample. Median atrazine concentrations in domestic wells in the Piedmont and Blue Ridge carbonate-rock aquifers and the Valley and Ridge carbonate-rock aquifers were higher than those measured in domestic wells nationally.

Thirty-five of the 47 pesticide compounds measured were detected in at least one sample, and at least one pesticide was detected in 55 percent of samples. The most frequently detected pesticides were atrazine, deethylatrazine, metolachlor, simazine, prometon, tebuthiuron, and dieldrin.

Mixtures of pesticides (two or more pesticides detected in the same sample) also were detected more frequently in the carbonate-rock aquifers than in crystalline- and siliciclastic rock aquifers. In fact, mixtures of pesticides in the Piedmont and Blue Ridge and Valley and Ridge carbonate-rock aquifers were much more prevalent than in aquifers nationwide with five or more pesticides detected in more than 40 percent of wells in agricultural areas and more than 80 percent of the wells in urban areas.

Pesticides with highest use—atrazine and metolachlor—were among the most frequently detected pesticides in domestic and public-supply wells.

All seven of the most frequently detected pesticides were detected most often in samples of groundwater from domestic wells in the carbonate-rock aquifers. Many of these domestic wells are also in agricultural areas.



Figure 1.61 Weathered rocks that make up the carbonate-rock aquifer are exposed at the surface and provide direct pathways for water and contaminants to enter the aquifer without the filtering effects of soils. Voids, or enlarged holes in carbonate rocks, such as the ones shown here, are common in the aquifer. Water moves rapidly through these voids and remains oxygenated. Note geologic hammer for scale. Photograph by William Kochanov, Pennsylvania Geological Survey. Text from Lindsey et al., 2014.

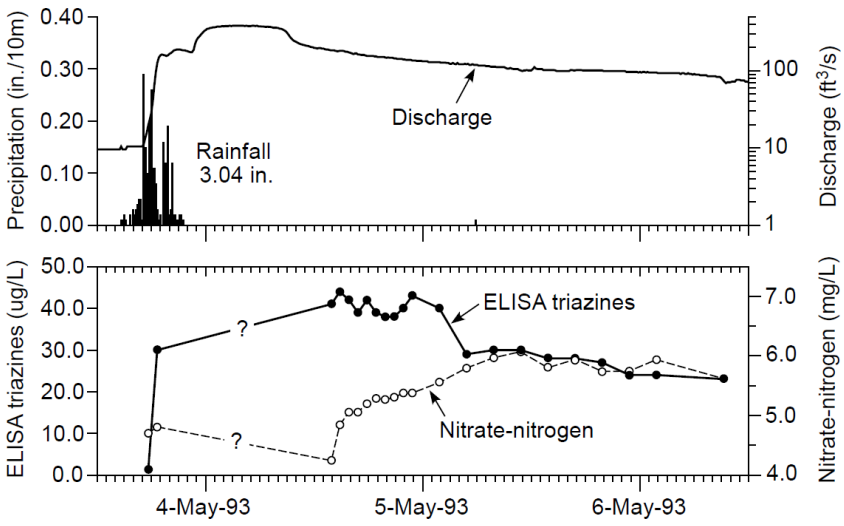


Figure 1.62 Chemographs and discharge hydrograph for a high-flow event at Pleasant Grove Springs in Logan County, Kentucky is an example of non-point source contamination from agricultural activities (Currens, 1999). Concentrations of triazines (including atrazine) and alachlor have exceeded drinking water MCLs during peak spring flows. Flow-weighted average concentrations for 1992–93 were 4.91 µg/L for atrazine-equivalent triazines and 5.0 mg/L for nitrate-nitrogen. In comparison, the maximum allowed concentration of any individual pesticide in drinking water in the European Union is 0.1 µg/L, and of all pesticides combined it is 0.5 µg/L.

In 2003, and again in 2020, the U. S. Environmental Protection Agency (U.S.EPA) allowed continuing use of atrazine. This herbicide has been banned for use in the European Union since 2004. In the U.S., it is one of the most commonly reported contaminants in groundwater and public drinking water. While atrazine is applied to a wide range of crops, it is primarily used on sugarcane, soy, sorghum, and corn; the U.S. Department of Agriculture notes that more than 65 percent of all corn crops in the U.S. have been treated with the herbicide. It is also a weed killer for golf courses, fields, and residential and commercial lawn spaces across the United States (https://usrtk.org/pesticides/atrazine/).

It is mystifying that in its evaluations of risks associated with the continuing use of atrazine and other harmful pesticides, the U.S. Environmental Protection Agency did not even survey or mention private wells and springs used for water supply. A significant body of scientific research suggests that atrazine is an endocrine disruptor and linked to various cancers, premature birth and birth defects (https://usrtk.org/pesticides/atrazine/).

1.2.1 Spring vs. River Rise

As stated repeatedly and shown with many photographs and schematics in this book, what sets karst aquifers apart from any other hydrogeologic environment is the nature and dynamics of surface water-groundwater interactions. This is also the focal point of ongoing polemics about what constitutes a true spring in karst. For most classically trained and professional, practicing hydrogeologists, the two key terms in this polemics, *aquifer* and *spring,* are inseparable. Equally important is the relation between *groundwater* and *aquifer*: groundwater resides within and moves (flows) though an aquifer. It then follows that a spring is a location at the land surface where groundwater discharges from an aquifer—a definition that is common to all textbooks on hydrogeology and scientific or general encyclopedias. However, when it comes to karst, these general definitions and relationships between aquifers, groundwater, surface water, and springs may be blurred ("all is not black and white"). The following points and discussion will, hopefully, shed some light on this sometimes-contentious polemics.

A casual observer or perhaps a karst geographer that describes what her or his eyes see will state that, if water comes out of a cave opening (however small or large), then this is undoubtedly a spring, no further discussion needed. A trained karst hydrogeologist would, in most cases, also ask himself or herself a much more important questions "What is the origin of that water"? before making any final judgment. This difference between the two approaches is embedded in the meaning of words **-graphy**, which refers to describing, drawing, or mapping a subject, and **-logy** which refers to the scientific study or theory of a subject. Put in a different way, a karst geographer sees water coming out of the cave and automatically describes it as a spring; a karst hydrogeologist wonders if at least some of that water comes from the karst aquifer before declaring it a spring, understanding fully that karst aquifers are unique when it comes to surface water–groundwater interactions.

Following are some examples illustrating the usage of terms *spring* and *river rise*, as well as the related hydrogeologic analyses.

The Santa Fe River in Florida flows west across the Northern Highlands to the Cody Scarp, an extensively karstified transition zone between the confined and unconfined parts of the Floridan aquifer (Scott, 1991) where it disappears into O'Leno sink. From O'Leno sink, the river flows underground for roughly three miles to its resurgence at the River Rise. From there, the river flows onto the lowland limestone plains of the Central Highlands and Gulf Coastal Lowlands where the Floridan aquifer is unconfined and precipitation infiltrates directly into the aquifer. Below the River Rise, the Santa Fe River is both a gaining and losing stream. It receives considerable water from several springs, however, less noticeable are the frequent "sucks" where part of the flow is diverted underground for some distance (Ellins et al., 1991; Kincaid, 1994).

Hisert (1994) conducted tracing experiments at O'Leno State Park which demonstrated that, after disappearing into O'Leno sink, the Santa Fe River flows at over 3 mi/day to its resurgence point at the River Rise. Hisert recorded the groundwater tracer at 8 intermediate sinkholes between O'Leno sink and the River Rise which are now known to be karst windows that intersect the underground flow path of the river. Furthermore, he found that over 40% of the resurgence at the River Rise is groundwater that augmented the river flow (Global Underwater Explorers, https://www.gue.com/hydrogeologic-setting.)

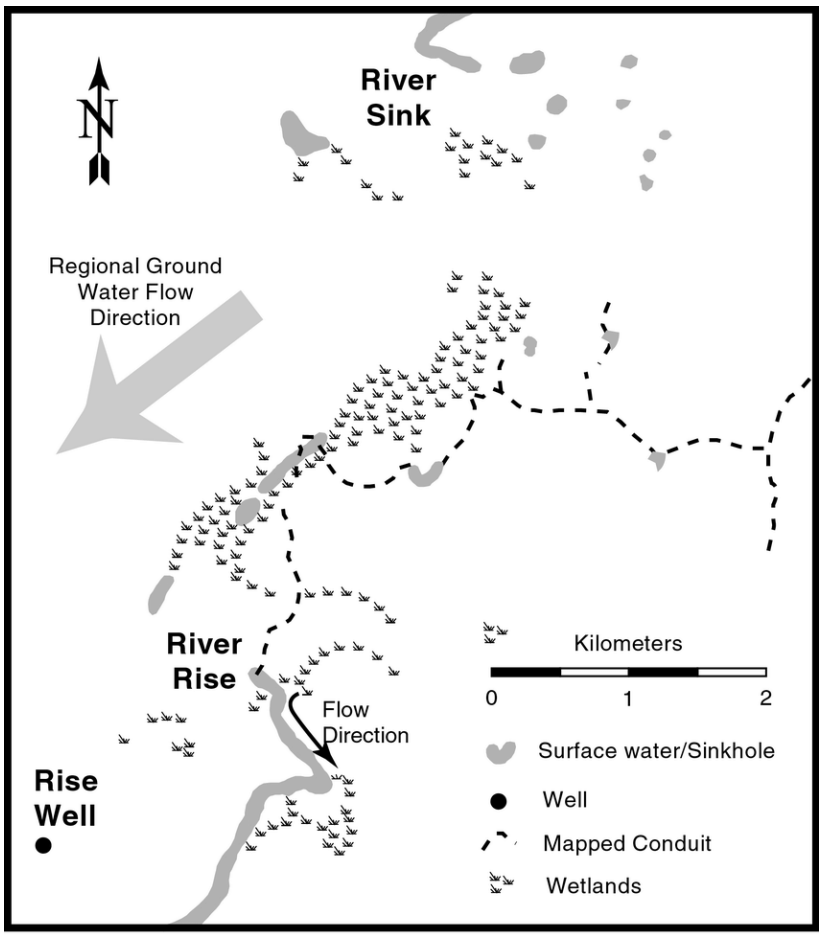


Figure 1.63 Sketch map showing the location of Santa Fe River at the sink/rise system, surface water bodies that commonly represent sinkholes, distribution of wetlands and the extent of the mapped conduit system. The river sinks into the subsurface at the location labeled River Sink, and re-emerges at the location labeled River Rise. Modified from Martin and Dean, 2001 in press; from Martin and Screaton, 2001.



Figure 1.64 The Santa Fe River goes underground in O'Leno State Park and reemerges 3 miles away in River Rise Preserve State Park as a circular pool before resuming its journey to the Suwannee River (https://www.floridastateparks.org/RiverRise). The water upwelling into the Rise pool is virtually always dark due to presence of naturally occurring substances in rivers called tannins, released during the decomposition of wood. Unlike this and other river rises, many true karst springs in Florida are famous for their clarity and near-constant water temperature of about 72 degrees Fahrenheit year-round. Photo courtesy of Suwanee River Water Management District (SRWMD). https://map.mysuwanneeriver.com/district-springs/santa-fe-river-rise-alachua-30/

The average flow rate of the Santa Fe River Rise, based on data made available by SRWMD, was 547.9 cfs as of 2014, the last year of data reporting (https://www.mysuwanneeriver.com/507/Water-Data-Portal). And yet, no government agency or anyone else has ever stated that the Santa Fe River Rise is a spring, or the largest spring in

Florida, even though its average flow rate was higher than that of any other true (hydrogeologically speaking) karst spring in Florida at the time, including Silver Springs and Wakulla Spring.



Figure 1.65 Holton Creek Rise with its characteristic dark color due to the presence of tannins. SRWMD does not consider it to be a true karst spring: *While it is connected to the Floridan Aquifer, Holton Creek Rise is also a resurgence of the Alapaha River. Most of its waters come from swallets in the Dead River Sink, where the Alapaha River disappears. Holton Creek Rise serves as an overflow for the Alapaha River, and when river levels are low or high, the spring flow fluctuates. This is a popular spot for fishing and hiking.* Photo and text courtesy of SRWMD (May 2026). https://map.mysuwanneeriver.com/district-springs/holton-creek-rise-25/

A similar example is described in detail by Intera Geosciences and Engineering (2021). The Alapaha River contains a series of swallets, natural openings in the riverbed, southeast of Jennings, Florida. This is approximately four miles south of the Florida state line. During periods of low flow, noted as below 350 to 500 cubic feet per second (cfs), surface water in the river is entirely captured by the swallets and enters limestone conduits. This water provides recharge to the Floridan aquifer (Upchurch, 2014). The water that enters the aquifer via the swallets at Dead River eventually discharges to the Suwannee River through two resurgences: the Alapaha Rise and Holton Creek Rise. This is about 19 miles away from the location of the sinks.

Ceryak (1977) describes groundwater level fluctuations of up to 20-feet between periods of low and high flow, where high flow is defined as flow greater than 786 cfs. During low flow, Holton Creek Rise discharges clear groundwater. During high flows when the river stage exceeds the elevation of the potentiometric surface of the Floridan aquifer, water from the river is forced into the Floridan aquifer. Discharge from the Alapaha Rise and Holton Creek Rise is typically discolored by organic material during such times. During low flows, discharge from Holton Creek Rise is entirely clear and will dry up during extreme low flows (Intera Geosciences and Engineering, 2021).



Figure 1.66 One of Alapaha River sinks. Courtesy of Suwannee Riverkeeper. https://wwals.net/2014/11/13/alapaha-river-sink/

The above analyses of surface water-groundwater interactions in karst are prime examples of quantitative hydrologic and hydrogeologic approaches to the evaluation of river rises which are discharging a mixture of surface water and groundwater (i.e. true aquifer water) in various proportions depending on surface water input conditions.

Table 1.1 Mean River Water Fractions at Alapaha Rise and Holton Rise based on Alapaha River flow conditions at Jennings. From Intera, 2021.

Flow Condition	Holton Rise River Water Fraction	Alapaha Rise River Water Fraction
< 400 cfs	0.3208	0.2414
400 – 1500 cfs	0.5931	0.5129
> 1500 cfs	0.8157	0.6907

An example of surface water-groundwater interactions in the case of true karst springs, where aquifer water is the most important component of spring discharge, is provided by Kincaid (1994).

The Devil's Ear Spring cave system in Florida is an extensive network of phreatic karst conduits that are directly overlain by the 1.5 km reach of the Santa Fe River between Rum Island and July springs. The cave and the river are situated on a karst limestone plain where the Ocala Limestone is exposed at the surface leaving the Floridan aquifer unconfined. Fluctuations in the stage of the Santa Fe River and the hydraulic head in the Floridan aquifer drive dynamic interactions between the surface water in the river and groundwater in the network of submerged cave conduits. Natural tracers Radon-222 (222Rn) and δ18O and the artificial tracer Sulfur hexafluoride (SF6) were used to quantify these interactions and describe their hydrogeological implications.

The Devil's Ear Spring cave system is comprised of the network of underwater cave passages that proceed upstream from Devil's Ear Spring. Devil's Ear Spring is a second magnitude spring that has gone mostly unnoticed in the scientific community because it is located in the Santa Fe River channel and is not visible from the surface throughout much of the year. Devil's Ear Spring is physically connected to two other springs, Devil's Eye and July where the combined discharge raises the group to first order status. All three springs are considered part of the Ginnie springs group (Wilson and Skiles, 1988). This spring system is famed for its water clarity, achieving an unrivaled fame among scuba divers of all skill levels when Jacques Cousteau dove here in 1974 and praised the crystal-clear water as "visibility forever." It is the international hotspot for scuba diving.

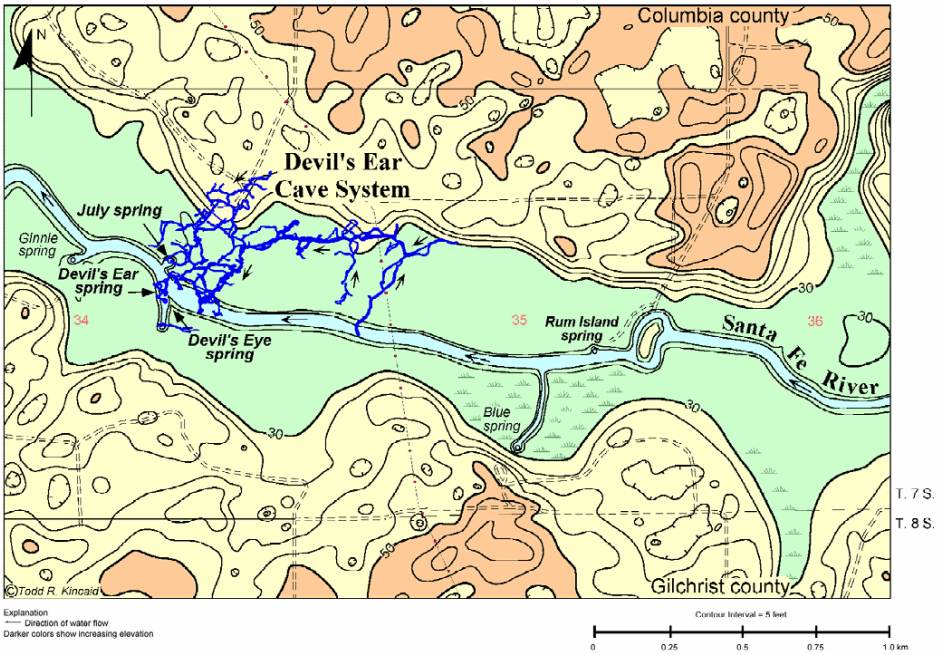


Figure 1.67 Section of the USGS High Springs S.W. topographic quadrangle, 7.5-minute series, showing the drainage features and karst landscape in the field area between Rum Island and July spring on the Santa Fe River near High Springs, north-central Florida. From Kincaid, 1994.



Figure 1.68 Drone photo of The Devil's Eye Spring by Frank Moore. Note the contrast between the crystal-clear spring water and dark tannic water of the Santa Fe River. Available at https://firstmagflorida.com/2020/10/10/devils-spring-system-ginnie-springs/

Fig. 1.69 shows groundwater percentage contour maps for the February 1992 and June 1993 comprehensive cave experiments which included cave diving and collection of water samples. The contour lines show the sources and distribution of pure groundwater and intruded river water throughout this section of the aquifer. Darker areas correspond to sections of the cave containing progressively greater amounts of intruded river water. Both maps show that surface water influx to the cave system originates from the passages on the south side of the system which trend toward and in some cases directly underlie the Santa Fe River.

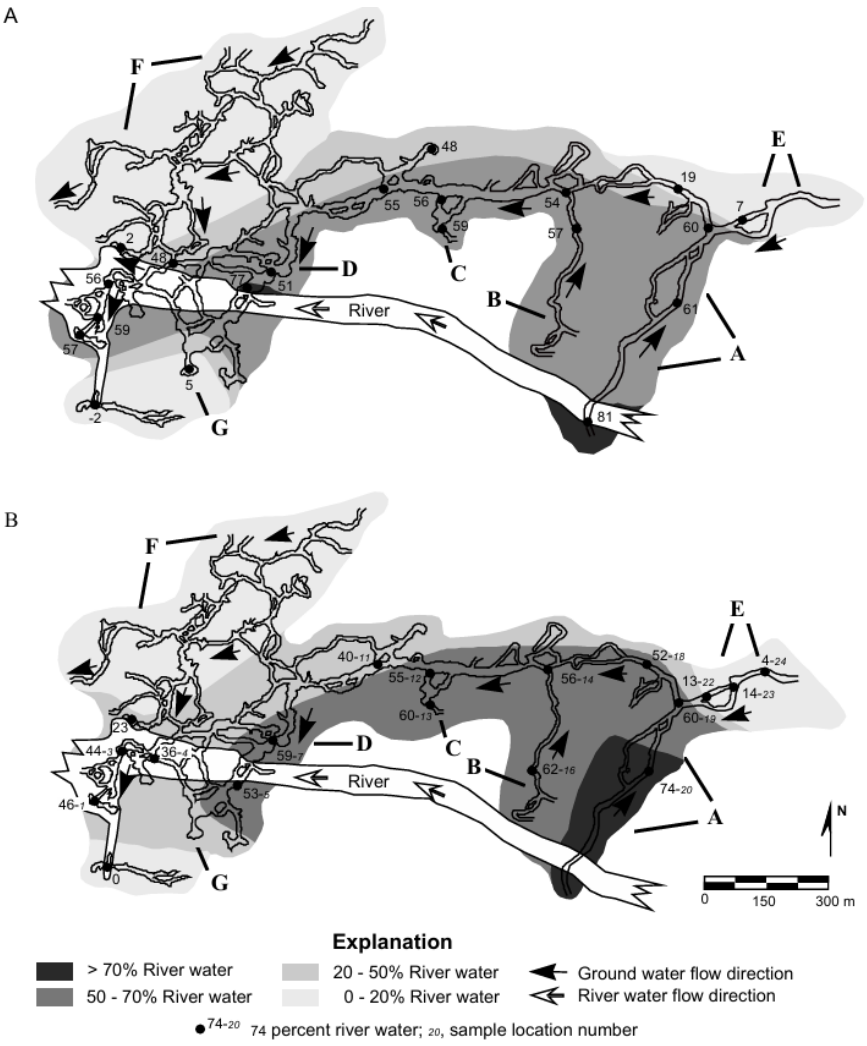


Figure 1.69 Contour maps showing the distribution of intruded river water in the Devil's Ear cave system on the Santa Fe River, north-central Florida during the (A) February, 1992 and (B) June, 1993 cave sampling experiments. Darker regions are the areas of greatest river water intrusion to the aquifer. From Kincaid, 1994.

The sections of the cave system contributing the most river water to the cave system are the: Mainlands region, Split Tunnel, River Intrusion Tunnel, and the Big Room (sections A, B, C, and D on Fig. 1.63). The contour maps indicate that the northern conduits are the major contributors of pure groundwater to the cave system (Kincaid, 1994).

The three presented examples of comprehensive scientific analyses of karst springs and river rises, as they relate to interactions between surface water and *aquifer water*, i.e., *groundwater,* are far from common as they are expensive, time-consuming, and are done by or with some help from experienced groundwater professionals. What they demonstrate is that, in karst, surface water including sinking streams, interacts with groundwater in different ways and to varying degrees, depending on specific circumstances.

However, when an entire surface stream disappears into the ground through a cave opening, and then reappears through a cave opening, after briefly (in terms of both time and length) flowing underground through a cave, the question of river rise vs. spring is not rhetorical anymore. This is illustrated with the schematics in Fig. 1.70. If during its underground journey, regardless of the time and distance involved, the underground portion of the surface stream does not interact with the aquifer surrounding the flow-through cave (conduit) in any appreciative fashion, then that cave outlet is not a spring. In other words, characteristics of the water input to the system (flow/quantity, temperature, concentration of constituents) are the same as of the water output from the system. The *aquifer water*, i.e., the *groundwater* is not involved, and the common hydrogeologic definition of a spring, including in karst, does not apply.

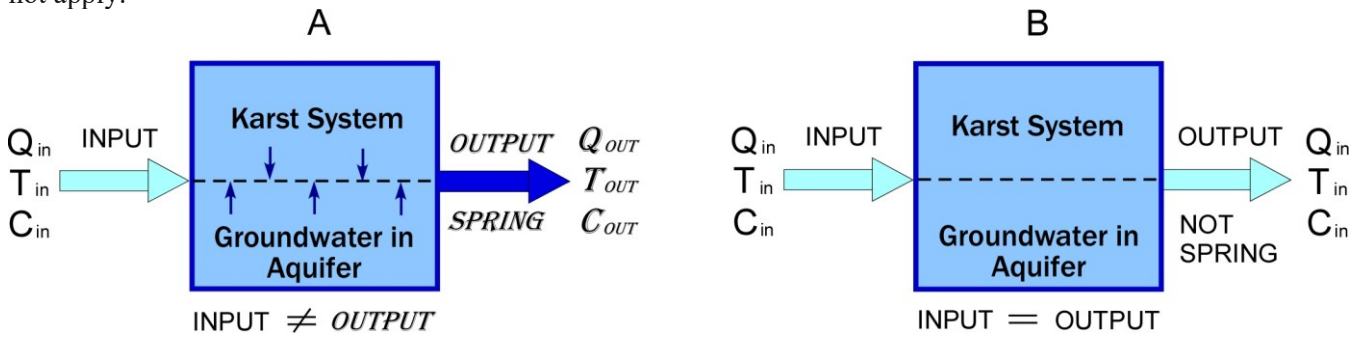


Figure 1.70 General schematic of a karst aquifer (system) which receives water input and has water output though a visible cave opening. A: If all or any of the characteristics of water, such as quantity (Q), temperature (T), or concentration of constituents (C) is transformed between the input and the output, then the output is a spring. B: When none of the characteristics is transformed, i.e. the water residing in the aquifer does not mix with the input water, then the output cannot be considered a spring. Note: There may be, and usually are, multiple water inputs such as sinking and losing streams (so-called *allogenic recharge*) and recharge from precipitation (so-called *autogenic recharge*).

In conclusion, when classifying or evaluating the emergence of water in karst, it is important to apply as many hydrogeologic principles as possible to avoid any misunderstandings. This includes a propagation of possible errors made because of generalizations based on insufficient information. The following excerpts from Gunn et al. (2025) illustrate this point. This paper explains the process of nominating springs to be included in the list of **M**ost **I**mportant **K**arst **A**quifer **S**prings in the world (project MIKAS of Karst Commission of the International Association of Hydrogeologists)

"Initially it was planned to have as much information about the proposed springs as possible. However, the PAB recognised that the amount of information available varies widely from country to country and decided to simplify the Survey Form to facilitate the work of national experts. One of the parameters that was initially thought of as being mandatory is the spring regime. However, it has become apparent that surprisingly few of the springs that meet other MIKAS and NIKAS criteria have their flow measured on a continuous basis and for some the flow

has only ever been estimated as opposed to being measured using recognised hydrometric techniques. A case in point is Tobio in Papua New Guinea which was listed in Ford & Williams (1989, Table 5.6 and 2007, Table 5.9) as being the world's largest karst spring with a mean discharge of 85 – 115m3s-1. The main reference given (Maire, 1981) states that "the huge Tobio emergence has been reckoned by Beck to [sic] 85 – 115 cubic metres per second". No reference is given but the actual source is thought to be Beck (1975) who states that "at Tobiu something like 3000 – 4000 cusecs (85 – 113 cumecs) was thundering from an entrance 55m high by at least 24m in width". The lead author of the present paper was able to contact Beck (pers. comm., 8 January 2025) who provided the following information: "On route to Pulupare the Iaro River which emerges from Tobio is crossed on a bridge about 16 km upstream… our estimate of water flow at the cave had no scientific basis, other than attempting to gauge volume from the banking at this crossing. I would think this would be the only safe place where a proper measurement might be attempted, but it was beyond our remit and abilities at the time. The river could well of course have picked up other unknown tributaries between bridge and resurgence". This is entirely reasonable and Beck is to be commended from at least making an estimate but it illustrates the problems of assembling data on world karst spring regime."

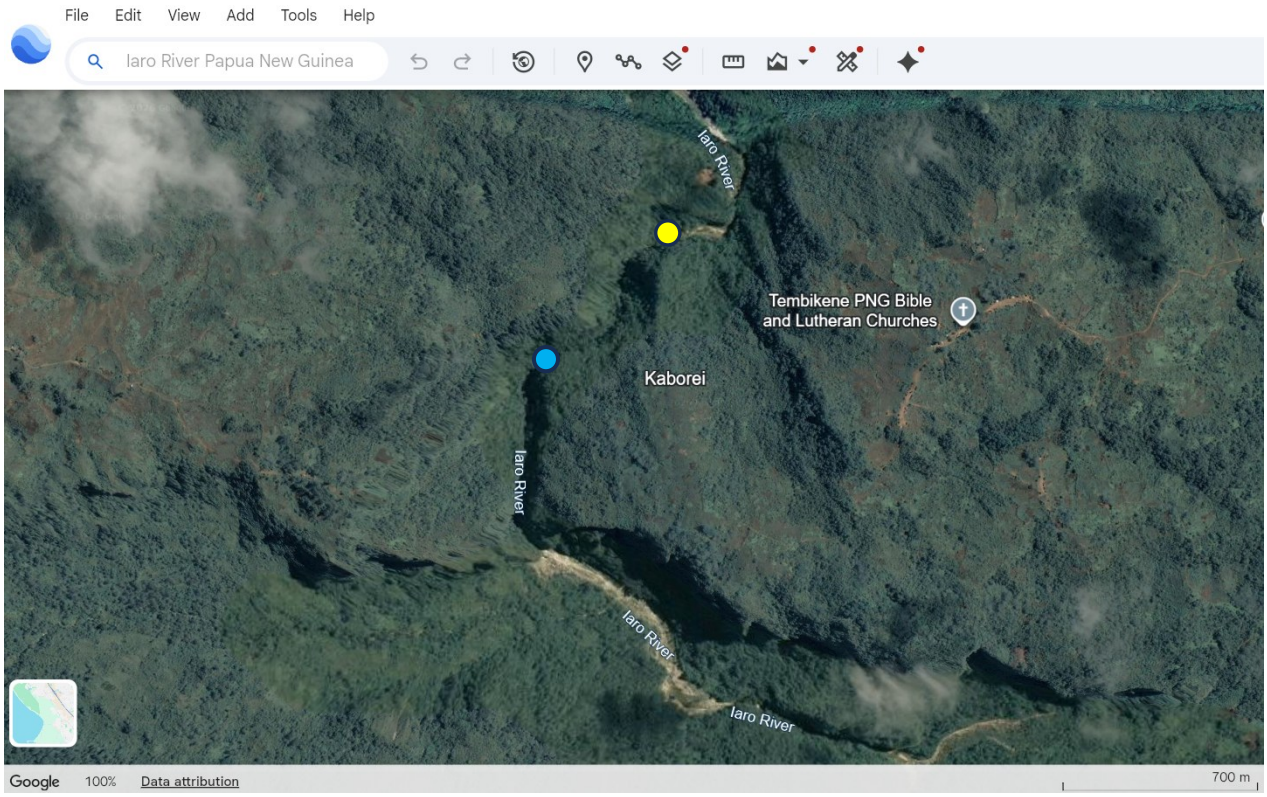


Figure 1.71 Google Erath image of the Iaro River section near Kaborei, Papua New Guinea. Added yellow dot denotes the Iaro River sink, added blue dot denotes the Iaro River resurgence locally called *Tobio*. A curved depression in the relief between the river sink and the river resurgence is an abandoned (former) canyon of the river, now hanging above the underground section of the river; it is about 670 meters (2,200 feet or 0.4 miles) long. Note metric scale in the bottom right. White labels are original to Google Earth Map. Image from Maxar Technologies CNES/Airbus, 4/20/2014 and newer.

By looking at a satellite image (Fig. 1.71) of the general area of Tobio, which is how the feature is called locally (it is not referred to as a spring), it is also entirely reasonable to conclude, without having any other information, that Tobio ("Spring") may be just an exit point of the underground section of the Iaro River, roughly 670 meters

long, which starts at its upgradient entrance to a flow-through cave. It is also possible that the percentage of karst aquifer (groundwater) contribution to the flow of this short underground section of the Iaro River may be minimal or even non-existent. Without any such information, it would be impossible to state that Tobio is a spring, let alone that it is *the world's largest karst spring*. Interestingly, the Google Earth Map layer, if chosen for a background map in Fig. 1.65, shows no break of the Iaro River between its sink and its resurgence called Tobio.

A Google inquiry about the Iaro River Cave yielded the following response (no answer was returned for "Tobio Spring" or any spring near Kaborei):

The large subterranean feature is officially documented in speleological records as the **Iaro River Cave**, located right in the Southern Highlands Province of Papua New Guinea.

First explored and mapped extensively by international caving expeditions in the mid-1970s, it is recognized as a **massive, high-volume river cave** Descent (31) Second-hand - Wild Places Publishing.

Hydrology and Scale of the Cave

- **The Underground River:** The cave serves as a dramatic underground conduit for the Iaro River. The river plunges into the limestone bedrock, carving out enormous chambers and roaring through subterranean channels before reaching its resurgence point Index to Niugini Caver - ANU Open Research.
- **Extreme Water Volume:** Because the Southern Highlands receives intense tropical rainfall, the Iaro River Cave carries immense amounts of water. It is considered a highly active and dangerous environment due to the risk of flash flooding inside the chambers.
- **Unique Ecosystem:** Early collections of cave fauna from the expedition teams documented unique, light-deprived wildlife thriving inside the cave, including specialized cave crickets and unique bat populations NEWSLETTER OF THE TASMANIAN ... - ASF Library.

The Caving History

- Detailed mapping and descriptions of this specific cave were published in early editions of *Niugini Caver*, a historical journal dedicated to documenting the incredible karst systems of Papua New Guinea. A prominent report on this exact system was written by explorer H. Beck in 1975, titled *"Iaro River Cave, Southern Highlands District."* NEWSLETTER OF THE TASMANIAN ... - ASF Library.

Below are excerpts from an article by Rolan Eberhard in the January/February 1990 issue of Newsletter of the Tasmanian Caverneering Club.

A SEASON IN THE HIGHLANDS Part 2: Iaro River

In October I made the long drive from Goroka to Kagua in the Southern Highlands. With me was Tony Wrightson, an expatriate kiwi who is a keen bushwalker but whom I'd managed to get interested in seeing a big river cave. This particular cave - formed where the Iaro River travels underground for some distance - was apparently first seen by white man around 1969 when Neil Ryan was kiap in the area. He returned to the location in 1974 accompanied by Howard Beck and others, and they briefly explored a section of fossil upper level inside the entrance. Some years later Neil Hickson also made short visit to the cave. From all accounts it was a very spectacular feature, with the flow rate of the emerging Iaro River being estimated at circa 100 cumecs.

After an overnight stay at Kagua, Tony and I drove along a steadily deteriorating road towards the village of Pulpare. It was decided to leave the vehicle at a pint where a small landslip had demolished part of the road and walk the final 5km. Our reception at Pulpare was rather cool. The local 'big man' said that the cave was taboo and

that we were not to go there. Long discussions ensued without getting us anywhere. At this point a younger man turned up and declared himself to be 'papa bilong graun' (ie. owner of the cave), and told us that we could go right ahead and have a look at it. Having come so far we decided to take his word for it and so we followed him down the track leading to the cave. On the way another man blocked the path saying that he was 'papa bilong gram true'. It was all getting rather tedious.

Eventually, after more haggling and a descent of several hundred metres down the precipitous sides of the Iaro gorge, we reached the efflux. Locally the feature is known as Tobio. From a gaping cavern in the rock face the Iaro River thundered out into daylight. The speed and force of water were quite awesome and downstream the river could be seen plunging through a narrow ravine reminiscent of the Gordon Splits. Like earlier visitors, we found that it was not possible to proceed upstream for any great distance inside the cave. The walls narrowed to a width of perhaps 12m and through this canyon pulsed the foaming Iaro. The upper level passage that had been explored in 1974 was clearly visible on the left side of the entrance, though we had no time for a proper look ourselves. Various 'papa bilong graunas' had subsequently arrived on the scene and were looking rather annoyed. This fact, combined with the long walk back, ensured that our visit was only brief.

Despite the hassle of local politics I am hoping to get back to the Southern Highlands for a closer look at Tobio. Howard noted the existence of numerous additional entrances in the dry canyon above the efflux and it is quite feasible that fossil upper levels meander over much of the length of the river on its underground course. A large log jammed in the river just inside the efflux indicates a fairly direct route from sink to resurgence, though exploring this at a safe distance above the raging water via upper levels sounds the most sensible option. As far as I am aware the point where the water goes underground - estimated to be about 1.5km away (Ryan, 1974) - has not been seen by anyone other than the locals, though they report that it is not hard to get to.



Figure 1.72 The Nostrils Cave in Sohodol Gorge, Carpathian karst of Romania. This is a textbook example of a flow-through cave with the outlet no one refers to as a "spring". This is simply because both the entrance of the surface stream into the underground ("input"; see Fig. 1.70) and its outlet ("output") can be clearly seen at the same time, by anyone. The cave is also readily accessible through its entire length. A potential argument may develop if the "input" (upper cave entrance) and the output are not both visible at the same time and are farther away from each other. In such case, a geographer or a casual observer may describe it as a karst spring (water is flowing out of a cave). A karst hydrogeologist will consider various elements discussed previously before making any final judgement if the feature is a karst spring, or a river rise, or just the end of an underground flow-through conduit that is not gaining water from the surrounding aquifer and is therefore not a karst spring. Photo courtesy of Zoran Stevanovic.

1.3 Example of Public Health and Environmental Catastrophe

In 2025, the non-profit organization Investigate Midwest, in partnership with the University of Missouri, investigated the link between agrichemicals and cancer in Missouri, finding that many cases were in rural communities that lacked access to healthcare. Investigate Midwest expanded its investigation by analyzing data across the country, along with interviewing more than 100 farmers, environmentalists, lawmakers and scientists as part of a partnership with the Pulitzer Center's StoryReach U.S. Fellowship. (Reporting was also supported by the Fund for Investigative Journalism.) The result was a picture of a nation at a crossroads in dealing with a public health crisis that has not just been ignored by state and federal health officials but aided.

America's farmers and farmworkers, their families and neighbors, are being diagnosed with cancer at rates higher than the national average. A growing body of research indicates that pesticides are partly to blame.

Most of the 500 counties with the highest pesticide use per square mile are located in the Midwest. Sixty percent of those counties also had cancer rates higher than the national average of 460 cases per 100,000 people, according to an analysis of data from both the U.S. Geological Survey and the National Cancer Institute (Felder, 2026).

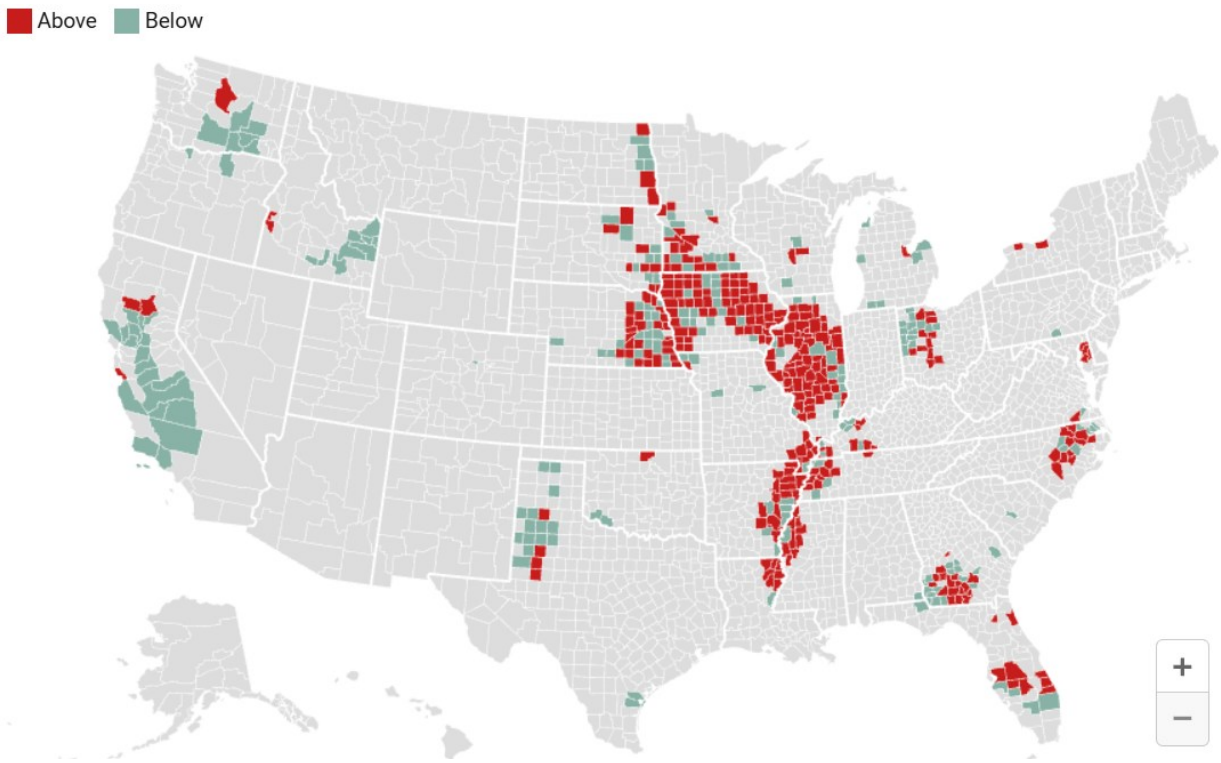


Figure 1.73 Top 500 counties in pesticide usage. Of the top 500 counties for per-square-mile pesticide use, shown on this map, more than 60% also had cancer rates above the national average of 460 cases per 100,000 people. Map: Ben Felder, Investigate Midwest. Source: National Cancer Institute and the CDC, five years of incidence data, from 2017 - 2021. Pesticides: Pesticide National Synthesis Project, from the U.S. Geological Survey. Created with Datawrapper.

Under the Trump administration, the Environmental Protection Agency and U.S. Department of Agriculture have hired dozens of former pesticide executives and lobbyists, some of whom have already pushed for deregulation of their industry. The Department of Health and Human Services has also altered its own reports to downplay the harm of pesticides.

Two states — North Dakota and Georgia — recently passed laws limiting their residents' ability to sue pesticide companies, and at least a dozen other states will consider similar laws in the coming months.

"We've gotten to a point in the U.S. … where we've stopped treating pesticides as if they are dangerous tools," said Rob Faux, who manages a small Iowa farm and has advocated against pesticide liability shield laws. "Instead, these companies tell these stories that these pesticides are completely safe and we are encouraged to use them anytime. We've been convinced that we must use them or we are not going to have enough food to eat."

In Iowa, a state with heavy pesticide use — 53 million pounds last year — and the nation's second-highest cancer rate, doctors and health officials have been sounding an alarm for years (Felder, 2026).

For years, national conversations about cancer have often focused on the so-called "Cancer Alley" in the Deep South or Appalachian states with high smoking rates. But recently, a different state has grabbed headlines. Iowa has the second-highest cancer incidence rate in the country and, together with West Virginia and Utah, is the only state where this rate is rising.

Iowa has become ground zero in the fight to limit the impact of pesticides on health and the environment. Farmers have gathered at the state Capitol to advocate for increased laws and funding to address the rising cancer rate. That advocacy likely helped defeat a bill last year that would have protected pesticide makers from some lawsuits.

Despite being largely hidden from view, Iowans rely heavily on groundwater as a water supply for a variety of domestic, agricultural and industrial uses, including 80% of Iowa residents who depend on groundwater for their drinking water (Prior et al., 2003).

Excerpts below are from "*Water contamination and cancer concerns*" by Sharon Kits Kimathi, Reuters Sustainable Switch, October 2, 2026 (https://www.reuters.com/sustainability/water-contamination-cancer-concerns-2026-10-01/).



Figure 1.74 Typical small-size cattle feeding lot (also called Concentrated Animal Feeding Operation or CAFO) in Iowa. Photo courtesy of Zomer Company.

We'll look at how Iowa's soaring cancer rates have become a campaign issue ahead of November's midterm elections. Many voters fear that the state's worsening water pollution from nitrates and other chemicals may be to blame for a spike in cases.

Iowa's rising cancer rate - the second highest in the US behind Kentucky - has emerged as a major campaign issue ahead of this year's vote.

A Reuters review of campaign materials showed that every major candidate in Iowa's competitive races for Congress and the state governorship has talked about their plans to tackle cancer in speeches and online posts.

Candidates have pledged a range of actions from increasing cancer screening to improving water quality monitoring. (Kimathi, 2026).

Excerpts below are from "*In a "toxic terrain," journalism exposes Iowa's farm pollution problem"* by Keith Schneider, The New Lede, July 27, 2026 (https://www.thenewlede.org/2026/07/in-a-toxic-iowas-farm-pollution-problem/)

Real change occurs in America when civic pressures from outside the centers of influence force insiders to retreat. A compelling example of that type of systemic change is now unfolding in Iowa.

Severe water contamination from crop and livestock production linked to rising high rates of cancer has developed into a key election year issue in Iowa, one of the country's most important farm states. Never before has industrial agriculture and its mammoth toxic discharges been subject in Iowa or any other state to such intense public scrutiny and condemnation.



Figure 1.75 Pollution seeps into an Iowa stream from a nearby farm. Photo by Lynn Betts/USDA.

Our work to fully explore the scandal of farm pollution and its health consequences, and our publishing partners in addition to Circle of Blue – Investigate Midwest, MinnPost, Iowa Capital Dispatch, The Guardian – also proved the influence of non-profit environmental journalism and collaboration in advancing the public interest.

For decades it was common knowledge in academia and government that supersize farms and livestock operations are the largest water polluters in Iowa, and almost every other state. In 2015, for instance, the Des Moines-area regional drinking water utility filed a federal lawsuit to regulate nitrogen discharges from farms upstream of its Raccoon River treatment plant. High concentrations of nitrates, which form when nitrogen unites with oxygen, were contaminating the drinking water of 500,000 residents, the utility said.

The case, though, occurred in an era when the Iowa Farm Bureau and its allies in the grain, hog, chemical, equipment, and seed companies and associations had built an intimidating political empire. Former Republican Gov. Terry Branstad assailed the case as a "war on Iowa farmers" and it was dismissed in 2017 by US District Court Judge Leonard Terry Strand.

With support from The New Lede, as well as the Fund For Investigative Journalism and the Alicia Patterson Foundation, Toxic Terrain dug deeper into the health and environmental damage caused by industrial farm practices, the source of the worst water pollution in the nation's history.

The depth of the reporting, its persistence, and its widespread publication by our non-profit partners validated public concern, and informed and aroused interest from other regional and national media, including recent reports by Politico and PBS. (Schneider, 2024.)

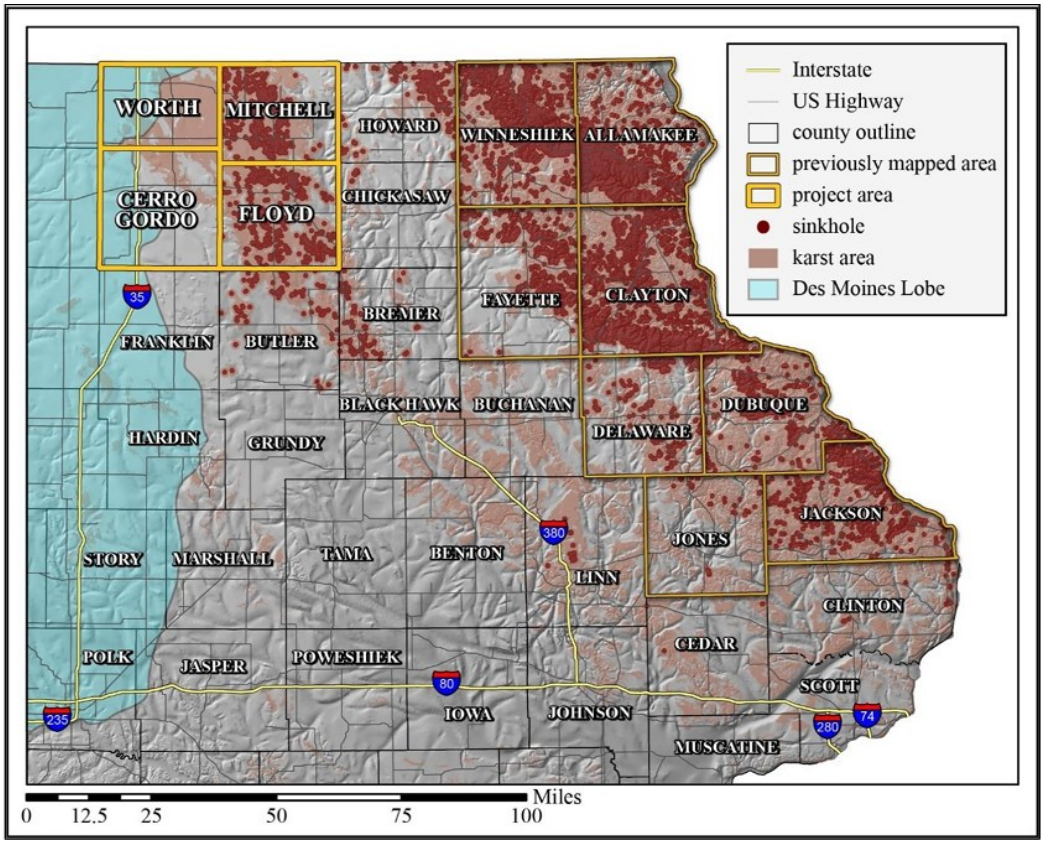


Figure 1.76 Map of northeast Iowa showing the existing sinkholes (red) and karst areas (pink). The previously mapped eight county area is outlined in yellow and the current four county project area is outlined in bold yellow. The Des Moines Lobe is shown in blue. From Tassier-Surine et al., 2021.

Finally, bringing it all together to emphasize the karst aspect of the previous discussions are the following excerpts from an article by Carey Gillam, The New Lede, entitled "*EPA called to address "long-ignored health crisis" in Iowa drinking water"* and published on April 17, 2024 (https://www.thenewlede.org/2024/04/epa-called-on-to-address-iowa-long-ignored-health-crisis-tied-to-farm-pollutants/)

The Iowa Environmental Council (IEC) and 12 other organizations are calling for the Environmental Protection Agency (EPA) to take a series of steps to reduce harmful nitrate pollution found in water supplies in Northeast Iowa.

The petition highlights the fact that for many years, nitrate concentrations found in thousands of private wells and some public water sources have exceeded the federal limit of 10 milligrams per liter for drinking water.

"For decades, Northeast Iowa residents have been exposed to dangerous levels of nitrate contaminated water," Food & Water Watch attorney Dani Replogle said in a press release. "As the state reckons with high cancer levels and ongoing pollution regulation rollbacks, federal action is needed to safeguard the right to clean water. EPA must exercise emergency authority to hold polluters accountable and deliver safe drinking water in Iowa."

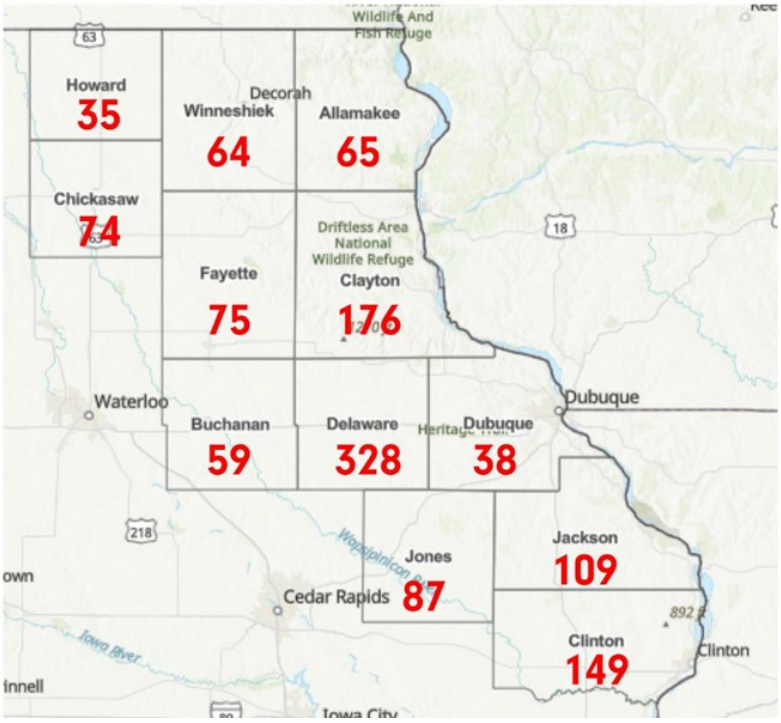
Iowa is one of the nation's largest farm states for crops and livestock, and Northeast Iowa is particularly vulnerable to farm pollutants because of its "karst" terrain, which features porous limestone that allows surface pollutants to more quickly move into groundwater. All of the drinking water in the karst region of northeast Iowa depends on groundwater aquifers.

Despite these facts, only a small fraction of private wells are tested regularly for nitrates, which are known to cause health problems at high levels, especially for infants. Cancer is a specific concern with nitrate exposure. Still, more than 1,200 wells in the 12-county karst region tested above the federal threshold for safety from 2016 to 2023, the petition states.

Residents of the region "have been, and continue to be, ingesting this contaminant. This alone demonstrates that the endangerment is substantial," the petition states. (Gillam, 2024)

The petitioners want the EPA to launch an investigation to determine why Iowa officials have not protected groundwater in the karst region, and determine what enforcement measures should be taken to reduce nitrogen pollution from large, concentrated animal feeding operations (CAFOs) and farms. CAFOs should be blocked from opening, expanding or modifying operations in the area, the petitioners say. They also want free water supplies to residents of the karst region.

Carey Gillam, 2024



All drinking water in northeast Iowa's karst region depends entirely on groundwater. An estimated 230,000 to 290,000 Iowans use private wells, which are unregulated and completely exempt from federal Safe Drinking Water Act mandates.

Figure 1.77 Wells per county in northeast Iowa that tested above 10 mg/L of nitrate between 2016-2023. IA_SDWA_Petition_Complete.pdf

Excerpts below are from an article by Brian Bienkowski, The New Lede, entitled "*New study finds higher cancer rates in counties with more CAFOs*", and published in Little Village on April 17, 2026 (https://littlevillagemag.com/new-study-finds-links-cancer-and-cafos/)

People living near concentrated animal feeding operations, or CAFOs, in California, Texas and Iowa suffer from higher rates of cancer, suggesting that the air and water pollution from the massive farms may be playing a role, according to a new study.

The study, published in the journal Environmental Research, comes just weeks after a years-long report in Iowa found the state's agricultural chemicals and CAFO pollution are, in part, behind its rising cancer rate.

CAFOs are major sources of water pollution from the large amounts of manure that are stored or spread on fields as fertilizer. This can lead to nitrate contamination in water, which can lead to the formation of carcinogens. Nitrate pollution in water has been linked to various cancers including colon, bladder and thyroid cancers. (Bienkowski, 2026).

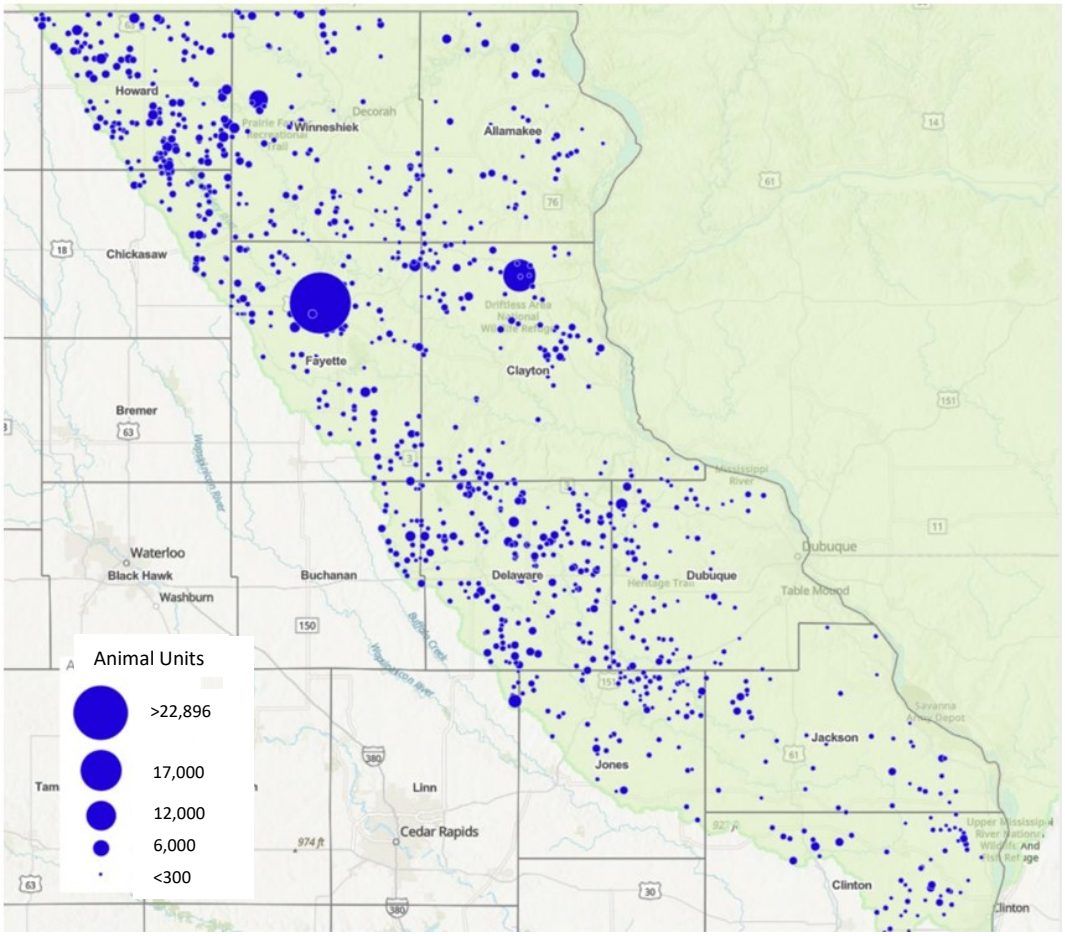


Figure 1.78 Medium and large animal feeding operations (AOFs) in the Driftless area as of 2022. Medium and large AFOs are defined as operations with 300 or more animals. AFO data from Iowa DNR. From IA_SDWA_Petition_Complete.pdf

Excerpts below are from an article by Brian Bienkowski, The New Lede, entitled "*Iowa cancer rates surge — farm chemicals are a key risk, new report find*" and published on March 25, 2026 (Iowa cancer rates surge — farm chemicals are a key risk, new report finds - The New Lede)

Iowa has nearly 87,000 farms, ranking first for corn, pork and egg production. It is also in the top five states for soybeans and raising cattle. Of Iowa's 35.7 million acres of total land, roughly 31 million is devoted to farming and the state has long suffered from excess pesticides, fertilizers and manure contaminating its waterways.

Iowans are suffering from higher rates of more than a dozen types of cancers linked to pesticides and pollutants than the rest of the country, with researchers saying the risk of pesticide exposure alone may rival that of smoking, according to a new report.

"This is impacting every corner of the state. There's no bounds … Democrat, Republican, urban and rural," said Sarah Green, executive director of Iowa Environmental Council (IEC), which published the new report along with The Harkin Institute, a public policy research institute located at Iowa's Drake University.

The new report focused on Iowa's three most-used pesticides: glyphosate, acetochlor and atrazine. The state is routinely one of the five top states for use of the three pesticides, and the report found they are linked to several types of cancers that are increasing in Iowa, including pancreatic, oral, kidney and breast.(Bienkowski, 2026)

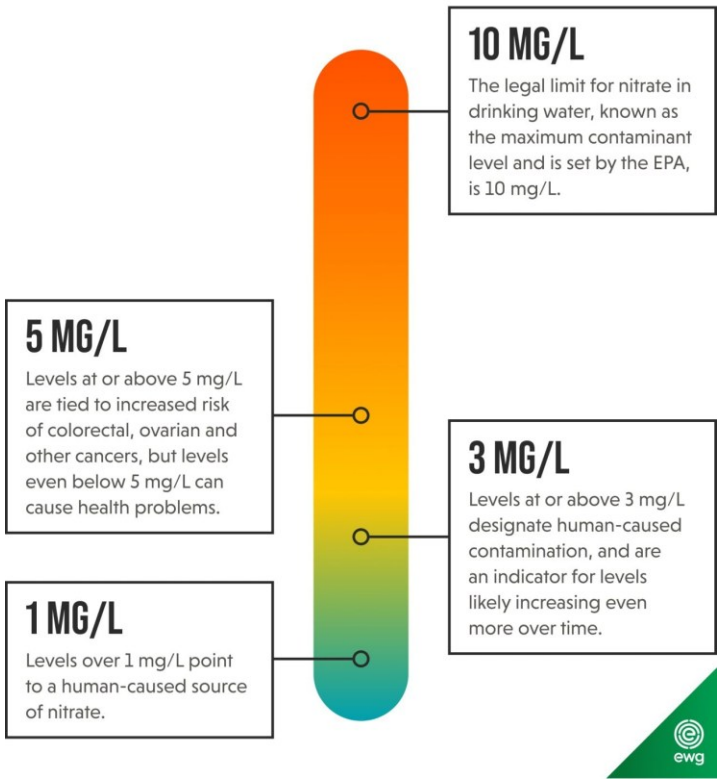
Notably, The European Commission withdrew its approval for atrazine and it has been completely phased out of use since 2004 (https://epha.org/eu-withdrawns-approval-for-herbicide-atrazine/). European Union (EU) banned atrazine due to ubiquitous and unpreventable groundwater contamination. The chemical consistently violated the EU's strict regulatory limit of 0.1 micrograms per liter (parts per billion) for any single pesticide in groundwater.

Despite the active ban, atrazine is very persistent in the environment. According to data tracking, atrazine and its byproducts remain among the most common pesticide contaminants found in European groundwater monitoring sites. The EU's stance stands in direct contrast to countries like the United States, where atrazine remains the second most widely used agricultural weedkiller and is permitted at levels up to 3 micrograms per liter (parts per billion) in drinking water (https://www.theguardian.com/environment/2026/sep/28/atrazine-herbicide-pollution-rivers.)



Figure 1.79 Aerial view of a CAFO (Concentrated Animal Feeding Operation) in Iowa. Courtesy of The Humane League (Communities in Iowa are standing up to factory farms | The Humane League)

Petition for Emergency Action Pursuant to the Safe Drinking Water Act, 42 U.S.C. § 300i, to Protect the Citizens of the Karst Region of Iowa from Imminent and Substantial Endangerment to Public Health Caused by Nitrate Contamination of Underground Sources of Drinking Water (Submitted on Behalf of Petitioners Iowa Environmental Council, Allamakee County Protectors – Education Campaign, Center for Food Safety, Environmental Law & Policy Center, Environmental Working Group, Food & Water Watch, Iowa Alliance for Responsible Agriculture, Iowa Citizens for Community Improvement, Izaak Walton League of America – Iowa Division, Sierra Club Iowa Chapter, Socially Responsible Agriculture Project, Iowa Coldwater Conservancy, and Trout Unlimited – Iowa Driftless Chapter 717; IA_SDWA_Petition_Complete.pdf):



The EPA has broad authority to investigate and remediate threats to public health under the Safe Drinking Water Act (SDWA). "Once EPA determines that action under Section 1431 is needed, a very broad range of options is available" as necessary to protect users of Underground Sources of Drinking Water (USDW). The tools available to EPA include conducting studies, halting the disposal of contaminants that may be contributing to the endangerment, and issuing orders such as mandatory changes to manure generation, handling, and land application practices.

Figure 1.80 Levels of nitrate in drinking water and their associated health risks. From Environmental Working Group http://www.ewg.org/

In fact, "EPA may take such actions notwithstanding any exemption, variance, permit, license, regulation, order, or other requirement that would otherwise apply." EPA should prioritize investigating and abating nitrate contamination in the karst region. EPA has already begun such action on a similar petition in Minnesota filed in April 2023. Groundwater does not follow political boundaries, and as shown above, Iowans are suffering from the same drinking water contamination as Minnesotans. At a minimum, EPA Region 7 must provide the same level of relief to Iowans as to Minnesotans in the Karst Region. Accordingly, EPA must require the state of Iowa to do the following:

1. *Create a communication plan to specify how information and responsibilities will be shared among state agencies, local governments, and any other entities acting on the response to the petition.*
2. *Identify each residence obtaining drinking water from a private well in the region.*
3. *Provide immediate notice and instruction to residents and continue providing such communication as long as contamination persists at or above the MCL (maximum contaminant level), particularly for vulnerable populations.*

4. *Create and implement a plan for drinking water sample analysis for all residents that request it.*
5. *Offer alternate drinking water at no cost to residents for each residence where water tests show an exceedance of the MCL for nitrate in the private well. The state must provide a detailed plan for distribution and timeline and minimize the burden on impacted residents.*
6. *Maintain and regularly publish records that allow the public to understand the scope and severity of nitrate contamination in the karst region and measure progress in implementing the response plan.*
7. *Provide quarterly progress reports to EPA describing (a) actions taken to address immediate health impacts of nitrate contamination; (b) major accomplishments and issues; (c) actions and timelines for the upcoming quarter; and (d) any problems or delays and the solutions implemented to address them.*
8. *Develop and implement a long-term solution to achieve reductions in nitrate concentrations in drinking water supplies.*

Furthermore, Petitioners respectfully request EPA undertake the following measures under its SDWA Section 1431 emergency powers, either by administrative order or through civil action:

***Investigation and Risk Assessment**:*

- *Investigate DNR's CAFO (Concentrated Animal Feeding Operation) permit requirements and Iowa Department of Agriculture and Land Stewardship (IDALS)'s and DNR's best management practices for nutrient management to determine why they have been unsuccessful at protecting groundwater in the karst region;*
- *Investigate other potential sources of nitrate pollution, such as failing septic systems;*

***Planning**:*

- *Determine what enforcement measures should be implemented to effectively reduce nitrogen pollution from CAFO and industrial agriculture sources;*
- *Provide a timetable for implementing a remedy to abate nitrate contamination from identified contaminators;*

***Assistance**:*

- *Order the parties responsible for the nitrate contamination to supply free water testing and ensure a free source of clean drinking water to residents of the karst region whose private wells or PWSs exceed safe limits for nitrate to prevent blue-baby syndrome, cancer, and other adverse health effects;*
- *Provide assistance to private well owners to engage in effective private well management practices;*

***Regulation**:*

- *Prohibit CAFOs from opening, expanding, or modifying operations in the karst region unless and until nitrate concentrations in wells with historically high levels of nitrate consistently fall below the MCL of 10 mg/L;*
- *Impose or require DNR to impose monitoring and discharge requirements in the karst region related to (1) subsurface discharges from manure, litter, and process wastewater storage, as well as (2) discharges from land application, in compliance with Food & Water Watch v. EPA, 20 F.4th 506 (9th Cir. 2021).*
- *Commence civil enforcement actions using authority in section 1431 against entities causing threats to public health by contaminating drinking water supplies.*

In conclusion, the presented case study of large-scale groundwater contamination in Iowa, and specifically in the state's karst region, demonstrates how the public education and the fight by various environmental organizations, investigative journalists, and concerned citizens against special interests, complacent government agencies and elected officials is irreplaceable in protecting public health and the environment.

1.4 Web Resources

Most states with karst have recognized its importance in a variety of ways, commonly addressing its vulnerability and the need to protect the dependent ecological and human environments. This includes sustainability of groundwater withdrawals for public supply, agriculture, and industry, contamination of karst water resources, and threats to aquatic life and animals that depend on both karst groundwater and surface streams in karst. There are many examples of state and local parks established because of notable karst springs, and there are karst drainage areas protected from unchecked development by local and state regulations. Some state agencies have attractive karst-specific educational and outreach programs, while others still lag behind. Most importantly, there are numerous non-profit organizations throughout the country that advocate for karst protection, often including lawsuits against special interests and government entities that choose to ignore the specifics of karst and its water resources.

Examples of recognition of karst as unique and valuable natural environment include Missouri Department of Natural Resources, Virginia Natural Heritage Karst Program, The Nature Conservancy (Ozark Highlands Karst Program), The Watershed Association in Texas, Minnesota Pollution Control Agency, Minnesota Department of Natural Resources, Iowa Department of Natural Resources (Karst Bedrock & Well Contamination), The Edwards Aquifer Website, Florida Springs Institute, Florida Springs Council, Ohio Department of Natural Resources, Karst Waters Institute, and United States Geological Survey, to name a few. Their respective web sites are provided below.

https://dnr.mo.gov/land-geology/geology/karst-missouri
https://www.dcr.virginia.gov/natural-heritage/karsthome
https://www.nature.org/en-us/get-involved/how-to-help/places-we-protect/ozark-karst-program/
https://watershedassociation.org/impactareas/education-and-outreach/karst/
https://www.pca.state.mn.us/air-water-land-climate/protecting-water-in-karst-regions
https://www.dnr.state.mn.us/waters/groundwater_section/mapping/springs.html
https://www.iowadnr.gov/environmental-protection/water-quality/private-well-program/well-testing/karst-bedrock-well-contamination/
https://www.edwardsaquifer.net/
https://floridaspringsinstitute.org/
https://www.floridaspringscouncil.org/
https://ohiodnr.gov/discover-and-learn/safety-conservation/geologic-hazards/karst
https://karstwaters.org/about-kwi/mission-statement-governance-and-goals/
https://www.usgs.gov/mission-areas/water-resources/science/karst-aquifers

Chapter 2 Karst Aquifers of the United States: West and Midwest

There are quite a few USGS maps and reports on principal aquifers, carbonate aquifers, karst aquifers, or areas with karst in the United States. Similarly, there are many reports and maps by the USGS and state geological surveys that feature karst areas and karst aquifers in individual states. However, they are not all necessarily consistent, not only because of the issues of scale. One difficulty when attempting to show the extent of major aquifers is that they may, and often do, extend both at the land surface (are outcropping), and deeper in the subsurface where they may be covered by other non-carbonate aquifers and rocks including aquitards (confining layers). There may also be multiple major karst aquifers below each other, separated by non-karst aquifers and aquitards which complicates attempts to create a single map of aquifers. In the case of karst, things get more complicated when considering areas with evaporate rocks, such as gypsum, which do cover large portions of Oklahoma and Texas for example, but are not considered aquifers because of inadequate water quality for water supply.

When aquifers are deeply buried and extend hundreds of miles beyond their recharge zones, the groundwater may become highly mineralized and not potable along its thousands of years long pathways. A good example of this is Madison Limestone (Early to Middle Mississippian), one of the most extensive karst aquifers in the country. It is the most important part of the "Paleozoic Aquifers" of the Northern Great Plains Aquifer System (Figure 2.1), often referred to as Madison Formation aquifer when not lithologically differentiated. Madison Limestone feeds major freshwater springs including thermal ones and it is utilized by thousands of wells for private and public water supply in the Northern Great Plains (Montana, Wyoming, South Dakota, North Dakota), particularly where surface water is limited. However, in most of North Dakota, i.e., where furthest away from its recharge areas in the Black Hills (S. Dakota and Wyoming), the Bighorn Mountains (Wyoming and Montana), and Little Belt and Big Snowy Mountains (North-Central Montana), the Madison Formation aquifer yields water with a dissolved solids concentration greater than 100,000 milligrams per liter (Driscoll et al., 2002).

The Colorado Plateau physiographic province is a good example of two large karst aquifers, one above the other and separated by confining layers, the sequence of which may be hundreds of feet thick. On its current web page dedicated to karst aquifer in the country, USGS lists Colorado Plateau Karst in Arizona as one of 13 major karst aquifers ("Karst Aquifers: Colorado Plateau Karst"; https://www.usgs.gov/mission-areas/water-resources/science/karst-aquifers-colorado-plateau-karst.) This karst aquifer, Kaibab limestone (Early to Middle Permian), stretches across northern Arizona and southern Utah. It prominently caps the spectacular landscape of the Grand Canyon. However, a much more prolific karst aquifer, (R-Aquifer; Mississippian) which feeds the two largest springs in Arizona and many other springs that sustain water in the Colorado River tributaries, is not mentioned or shown on the USGS map. This equivalent of Madison Limestone is deeply buried hundreds of feet below the Kaibab limestone and the intervening aquitards (see Chapter 2.6).

Major karst areas and karst aquifers in the western and midwestern United States are shown on the map in Figure 2.2, with added annotation. This map is best viewed together with a map of all principal aquifers of the United States compiled by USGS (2003 revised version, The National Atlas of the United States of America, available at https://www.usgs.gov/mission-areas/water-resources/science/principal-aquifers-united-states). The two maps provide a solid overview of karst (carbonate) aquifers in the country.

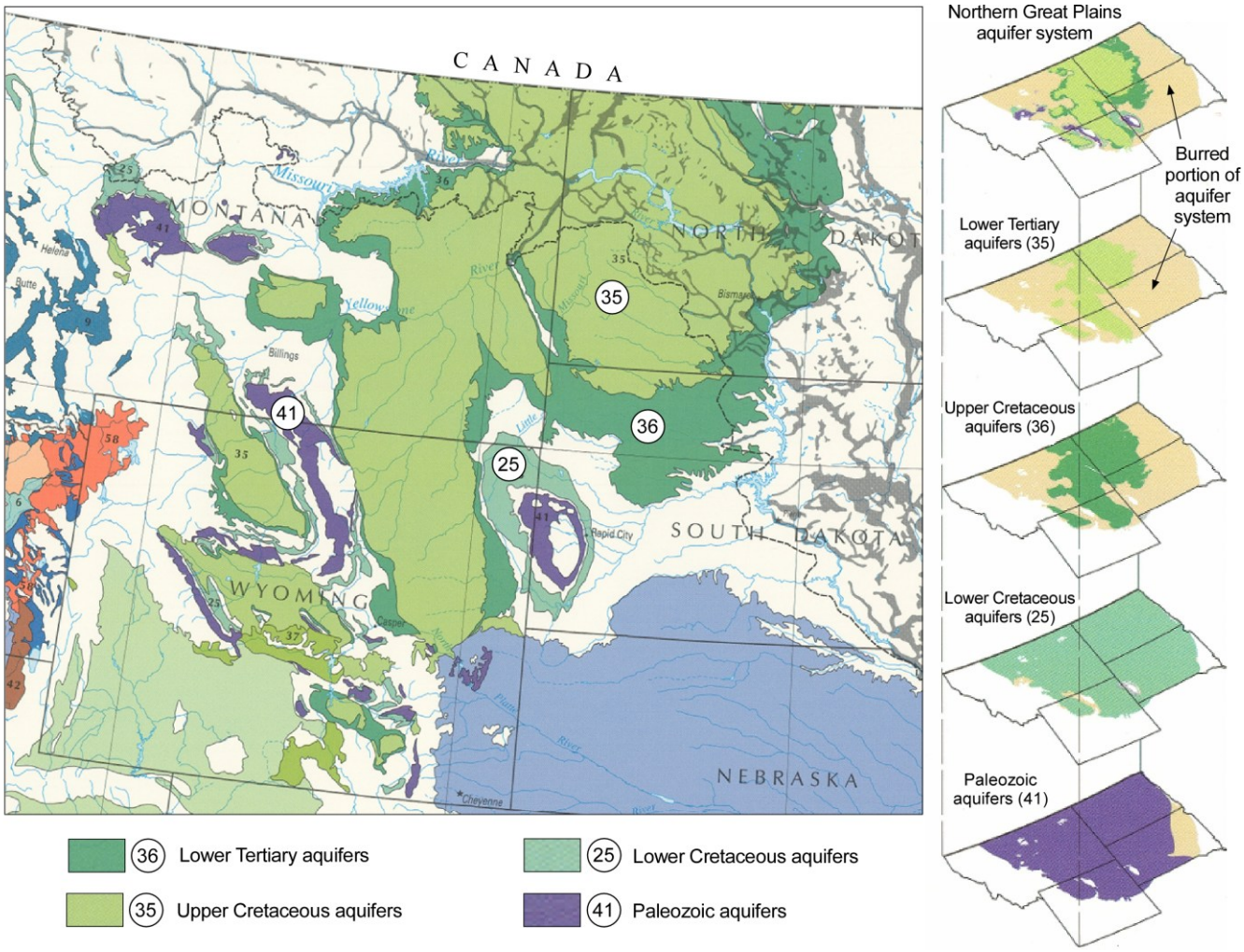
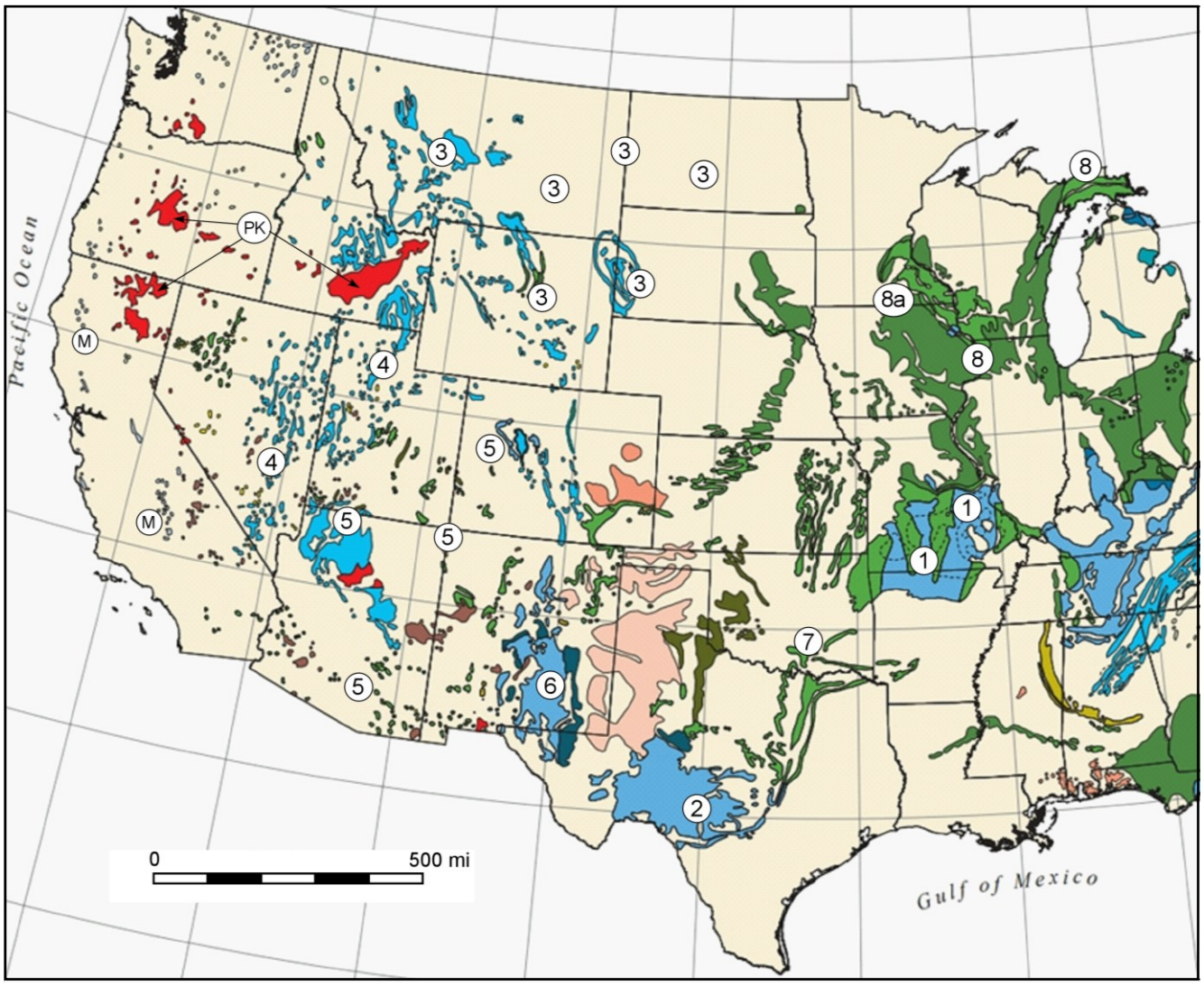


Figure 2.1 Part of a map of principal aquifers in the United States compiled by USGS and available at https://www.usgs.gov/mission-areas/water-resources/science/principal-aquifers-united-states. This 2D map, derived from Ground Water Atlas of the United States data (published as part of the National Atlas in 1998, revised 2003), indicates the areal extent of the uppermost principal aquifers on a national scale. A principal aquifer is defined as a regionally extensive aquifer or aquifer system that has the potential to be used as a source of potable (drinkable) water.

Aquifers shown in Figure 2.1 are composed of the following formations and rocks. **36**: Tertiary Aquifers (uppermost, shallow in many areas): Paleocene Fort Union Formation, and Eocene Wasatch Formation. Primarily semiconsolidated to unconsolidated sandstone, interbedded with siltstone, claystone, and coal/lignite beds. Locally, fractured clinker zones (burned coal) are highly productive. **35**: Upper Cretaceous Aquifers: productive Fox Hills Sandstone and Hell Creek Formation (sandstone beds), with Piere Shale, Lewis Shale, and Steele Shale acting as thick confining layers separating them. **25**: Lower Cretaceous Aquifers (Dakota Aquifer System): Dakota Sandstone, Inyan Kara Group (Lakota Formation and Fall River Formation), and Newcastle/Muddy Sandstone, with Skull Creek Shale and Thermopolis Shale acting as confining units. Predominant rock type in these productive aquifers is highly porous sandstone. **41**: Paleozoic Aquifers, deeply buried: Madison Limestone (or Madison Group) of Mississippian age, Minnelusa Formation (Permian/Pennsylvanian; includes carbonates), Tensleep Sandstone, and Red River Formation (Ordovician). Primary water-bearing units are highly karstified limestone and dolomite.



LEGEND

Fissures, tubes, and caves over 1,000 ft. long; 50 ft. to over 250 ft. vertical extent

- In metamorphosed limestone, dolostone, and marble
- In moderately to steeply dipping beds of carbonate rock
- In gently dipping to flat-lying beds of carbonate rock
- In gently dipping to flat-lying beds of carbonate rock beneath overburden of noncarbonate material 10 ft. to 200 ft. thick
- In moderately to steep dipping beds of gypsum
- In gently dipping to flat-lying beds of gypsum

Fissures, tubes, and caves less than 1,000 ft. long; 50 ft. or less vertical extent

- In metamorphosed limestone, dolostone, and marble
- In crystalline, highly silicious intensely folded carbonate rock
- In moderately to steeply dipping beds of carbonate rock
- In gently dipping to flat-lying beds of carbonate rock
- In gently dipping to flat-lying beds of carbonate rock beneath overburden of noncarbonate material 10 ft. to 200 ft. thick
- In moderately to steep dipping beds of gypsum
- In gently dipping to flat-lying beds of gypsum
- In gently dipping to flat-lying beds of gypsum beneath overburden of nongypsiferous material 10 ft. to 200 ft. thick

Fissures, tubes, and caves generally absent; where present in small isolated areas, less than 50 ft. long; less than 10 ft. vertical extent

- In crystalline, highly siliceous intensely folded carbonate rock
- In moderately to steeply dipping beds of carbonate rock
- In gently dipping to flat-lying beds of carbonate rock

Features analogous to karst

- Fissures and voids present to a depth of 50 ft. or more in areas of subsidence from piping in thick, unconsolidated material
- Fissures and voids present to a depth of 50 ft. in areas of subsidence from piping in thick, unconsolidated material
- Fissures, tubes, and tunnels present to a depth of 250 ft. or more in lava
- Fissures, tubes, and tunnels present to a depth of 50 ft. in lava
- Areas in which extensive historical subsidence has occurred

Figure 2.2 Karst map and principal karst aquifers of the western and midwestern United States (on previous page). From Tobin and Weary, 2004. Annotation with names of karst aquifers added. 1: The Ozark Plateaus Aquifers (including where buried); 2: Edwards-Trinity Carbonate Aquifers, including Edwards Balcones Fault Zone Aquifer; 3: Madison Formation limestone (outcrops denoted with blue color; and with numbers where buried deep; see Figure 2.1); 4: Basin and Range and Bear River Range Carbonate Aquifers (outcrops shown with blue, carbonates also extend deep below non-carbonate basin-fill aquifers); 5: Colorado Plateau Karst Aquifers, both shallow (Kaibab limestone), and deep (Mississippian R-Aquifer and equivalents); 6: Roswell Aquifer; 7: Arbuckle-Simpson Aquifer; 8: Midwest Paleozoic Carbonate Aquifers; 8a: Upper Carbonate Aquifer (Paleozoic); M: Marble karst in California (The Marble Mountain Wilderness in Klamath Mountains, and marbles in Sequoia and Kings Canyon National Parks; not principal aquifers); PK: Pseudokarst aquifers in young basaltic rocks of Idaho, Oregon, and California.

2.1 The Ozarks Plateaus Aquifers

The Ozark Plateaus aquifer system consists primarily of intensely karstified carbonate rocks that are Cambrian through Mississippian in age. It underlies most of southern Missouri and a small part of southeastern Kansas; it also underlies a large area in northwestern Arkansas and a small part of northeastern Oklahoma (Fig. 2.1.1). The Ozarks were formed by a structural dome that has been uplifted during several periods since Precambrian time. The core of the dome is in southeastern Missouri. Widespread extensional fracturing, jointing, and faulting of Ozarks' rocks occurred with uplift (Hudson and Cox, 2003), creating secondary porosity that has provided key starting points for initiation of dissolution of the carbonate rocks and karst development. The karstic bedrock of the region is overlain by weathered regolith that greatly varies in thickness from near zero in many areas to more than 100 ft. The regional dip of the rocks is westward and southwestward.

The aquifer system consists of two aquifers and an intervening confining unit. The uppermost aquifer is the Springfield Plateau aquifer, which consists of the Keokuk Limestone and the Reeds Spring and the St. Joe Limestone members of the Boone Formation. The rocks are mainly chert and limestone with solution-enlarged openings and local cavernous zones. The Springfield Plateau aquifer has a maximum thickness of about 500 feet in Oklahoma and contains water under unconfined to confined conditions. Yields of wells completed in the aquifer range from a few gallons per minute to more than 80 gallons per minute, whereas springs that issue from the aquifer might discharge as much as 3,500 gallons per minute. A confining unit that consists of rocks of Ordovician, Devonian, and Mississippian age and is called the Ozark confining unit separates the Springfield Plateau aquifer from the Ozark aquifers.

The Ozark aquifer consists of dolomites and sandstones within the Roubidoux, the Gasconade, the Eminence, and the Potosi Formations of Ordovician and Cambrian age. The thickness of the Ozark aquifer ranges from 200 to 500 feet. Much of the water in the aquifer is in fractured and highly karstified dolomite and sandy zones where it is under confined conditions. Wells completed in the Ozark aquifer commonly yield from 50 to 250 gallons per minute and locally yield as much as 1,000 gallons per minute (USGS, 2021).

Bedrock in the Ozarks shows evidence of multiple episodes of water movement and carbonate dissolution defining distinct karst development events through time (Stoffell et al., 2008). Widespread dissolution features associated with the lead and zinc ore-bearing fluids that moved from the Arkoma Basin (Leach and Rowan, 1986) are observed across the Ozarks region. Lead and zinc were mined commercially where faults conveyed mineral-saturated fluids up from depth.

Hypogene indicators such as diagnostic calcite isotopic compositions (Brahana et al., 2009), collapse breccias (McKnight, 1935), and isotopic dating (Brannon et al., 1996) show that this episode of fluid movement and dissolution was a separate occurrence predating recent karst development. Exposure of Mississippian strata

associated with the Mississippian-Pennsylvanian unconformity resulted in meteoric diagenesis followed by pedogenesis, regolith formation, and paleokarst development on the Pitkin Limestone (Webb, 1994).

The Ozarks once again showed active karst development as uplift, denudation, and abundant precipitation initiated dissolution of soluble carbonate lithologies. The recent period of dissolution and karst development has had a very visible impact on the modern karst land surface, controlling subsurface flow and ultimately the surface flow in the region.

The hydrogeology of the region is typical of karst with focused flow paths that are well connected to the surface and deliver water from input to discharge points at streams and springs at velocities of tens to thousands of feet per day (Funkhouser et al., 1999; Mott et al., 2000; Hudson et al., 2005). Abundant karst features are apparent—ponors (sinks), losing stream reaches, springs, caves, and sinkholes. Recent karst development often reactivated and followed previous dissolution-enhanced flow paths that originally developed during ancient exposure periods (such as at the Ordovician-Mississippian and the Mississippian-Pennsylvanian unconformities; Webb, 1994) or hypogene episodes such as Mississippi Valley Type ore emplacement (Leach and Rowan, 1986; Brannon et al., 1996; Brahana et al., 2009).

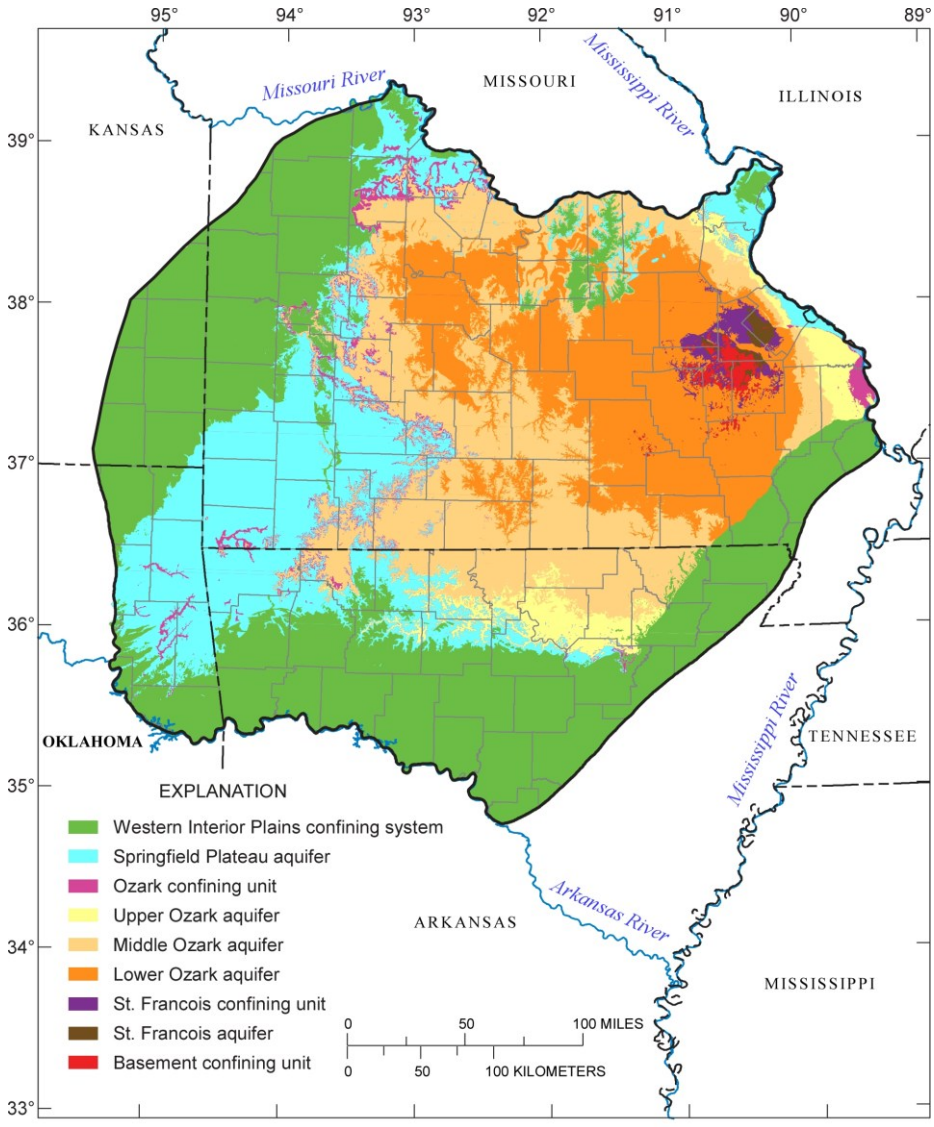


Figure 2.1.1 Aquifers of The Ozark Plateaus. The extent of the Ozarks Plateaus aquifer system is shown with thick black line. Modified from Westerman et al., 2016; USGS.

Another very important phenomenon caused by karst development is interbasin transfer of water. Dye tracing studies and observations of drainage-area and discharge relations show the abundant occurrence and transfer of groundwater across surface-water drainage-basin divides (Mott et al., 2000). This is one of the key reasons why the Ozark aquifer is drained by some of the largest first-magnitude springs in the country and many second and third-magnitude karst springs (see Chapter 3.1).

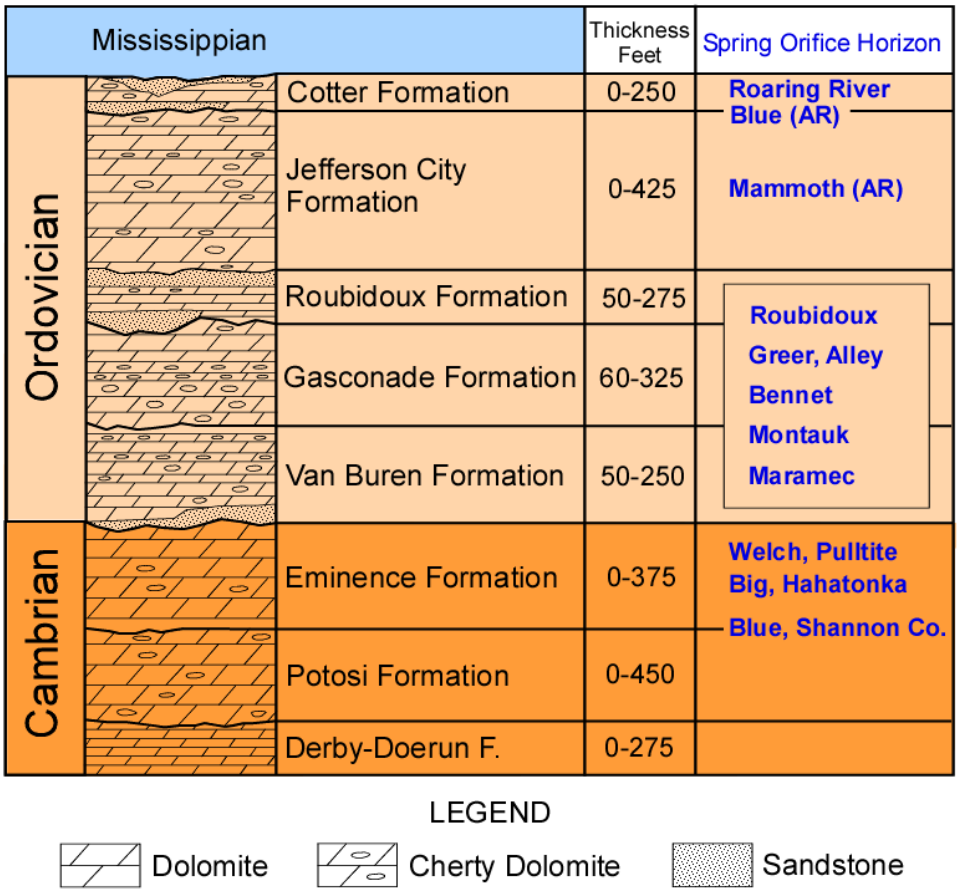


Figure 2.1.2 Principal geologic units of the Ozark Aquifer, with the horizons feeding large karst springs described in Chapter 3.1. Modified from Beckman and Hinchey, 1944.

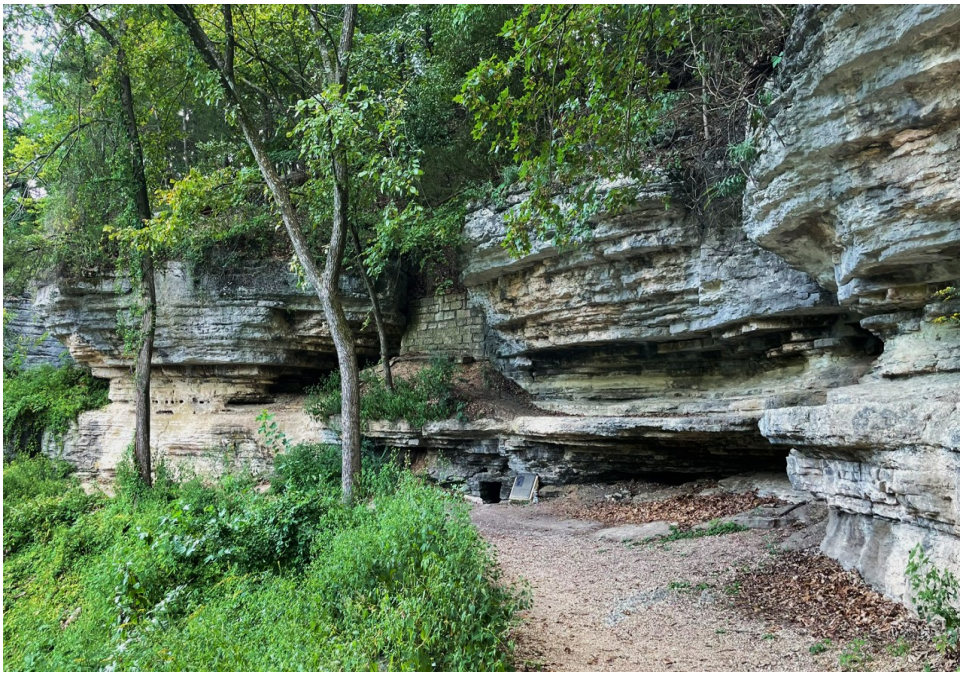


Figure 2.1.3 Outcrop of the Boone limestone (St. Joe limestone member) at Basin Spring, the first spring captured for water supply of Eureka Springs, Arkansas. Native Americans believed that the spring had healing powers.

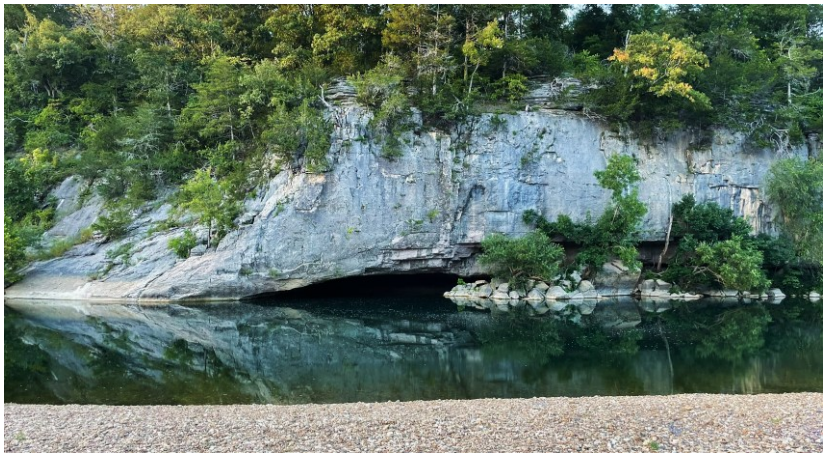


Figure 2.1.4 Bluff on Buffalo River, Arkansas at Grinders Ferry (near highway 65 bridge) made of Mississippian Boone Formation limestone. The Buffalo River National Park Region was established in 1972 to protect one of the few remaining undammed rivers in the nation. Nestled in the northern part of Arkansas, the Buffalo River was recognized by National Geographic as one of America's most underappreciated National Parks. It has more than 500 caves within its boundaries, making it one of the most cave-rich units in the National Parks.



Figure 2.1.5 *Left*: Lost Valley Cave on Clark Creek, a tributary to the Buffalo National River near Ponca, northwestern Arkansas.



Figure 2.1.6 One of the most photogenic bluffs on the entire Buffalo National River called Roark Bluff. Middle Ordovician Everton Formation sandstone and dolostone are the major rock types along this section of the river.



Figure 2.1.7 Intensely karstified Eminence Dolomite on a trail around Alley Spring.

2.2 Edwards-Trinity and Edwards Carbonate Aquifers

The Edwards-Trinity Aquifer consists of rocks of Cretaceous age and occupies the west-central part of Texas, extending from the Hill Country near Austin and San Antonio up to the mountains of West Texas and into the High Plains. The aquifer is referred to in much of the literature as the "Edwards-Trinity (Plateau) aquifer." This area is called the Edwards Plateau, but it also extends into the Trans-Pecos area or Stockton Plateau (Figs. 2.2.1 through 2.2.3). Because of low rainfall, the frequency of drought, and few major rivers, groundwater is an important source of water to the people and environmental resources of the Edwards Plateau area. The aquifer is located in the Great Plains Physiographic Province except for a small part that is in the Basin and Range Physiographic Province.

The topography of much of the area is characterized by flat to rolling, largely rocky plains that are dissected in places to form steep-walled canyons. The area is bounded on the west by mountain ranges. The altitude of the land surface ranges from about 1,000 feet at the Rio Grande in Val Verde County to more than 4,500 feet in the Davis Mountains in Jeff Davis County. Average annual precipitation ranges from about 12 inches in the west to about 30 inches in the east. Average annual runoff ranges from about 0.2 inches in the west to about 5 inches in the east. The Edwards-Trinity aquifer supplies large amounts of water for irrigation, particularly in the northwestern area; it also rovides water to many small towns and cities (Ryder, 1996).

The Edwards aquifer overlies the Trinity aquifer, and the two aquifers are hydraulically connected where no confining unit separates them (Fig. 2.2.3). The groundwater flow systems and permeability of the two aquifers are sufficiently different, however, to allow them to be separately mapped and described.

The rocks that compose the Edwards-Trinity aquifer are relatively flat-lying and are generally exposed at the land surface in the Trans-Pecos and the Edwards Plateau areas (Fig. 2.2.4). The water-bearing units are composed predominantly of limestone and dolomite of the Edwards Group and sands of the Trinity Group. The saturated

thickness of this aquifer system increases from less than 100 feet in the north to greater than 800 feet down-dip to the south. Freshwater saturated thickness averages 433 feet (George et al., 2011).

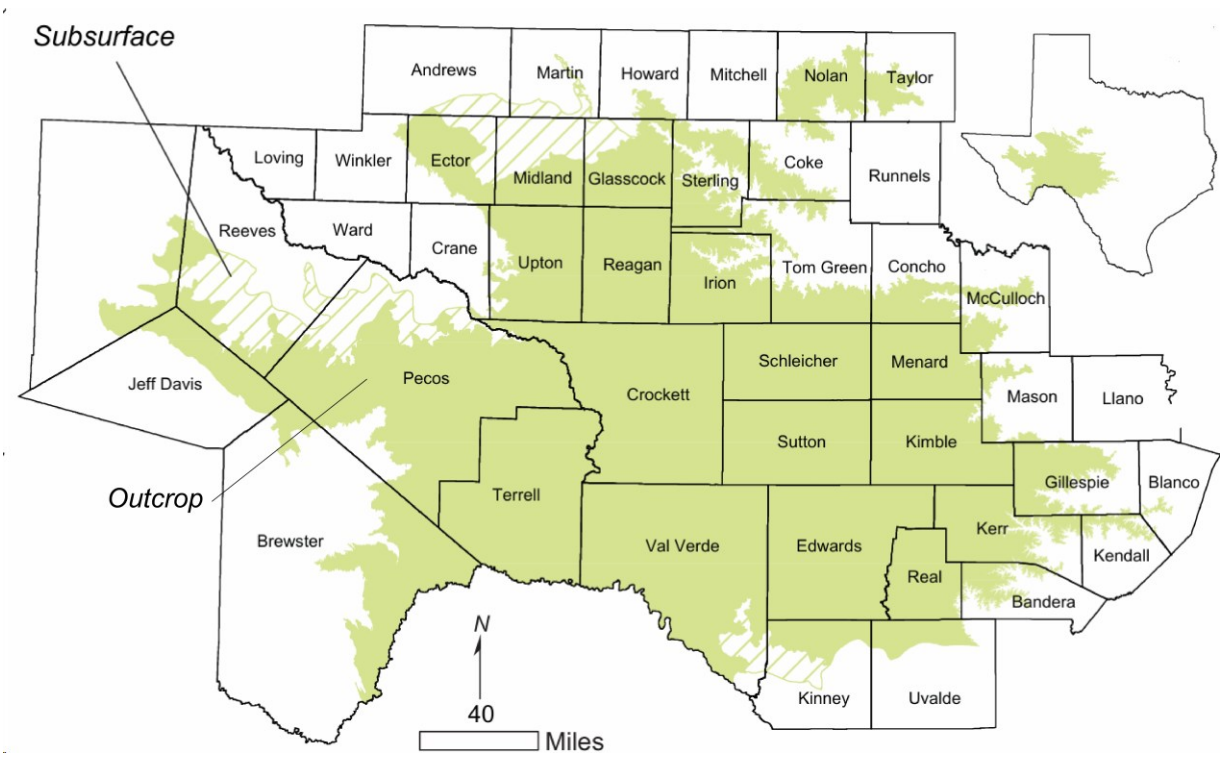


Figure 2.2.1 Edwards-Trinity Carbonate Aquifer in Texas. From George et al., 2011.



Figure 2.2.2 Physiographic map of west-central Texas showing various components of the greater Edwards-Trinity Aquifer System including notable springs. Modified from Land and Veni, 2018, Figure 5; and Barker and Ardis, 1996, Figure 3.

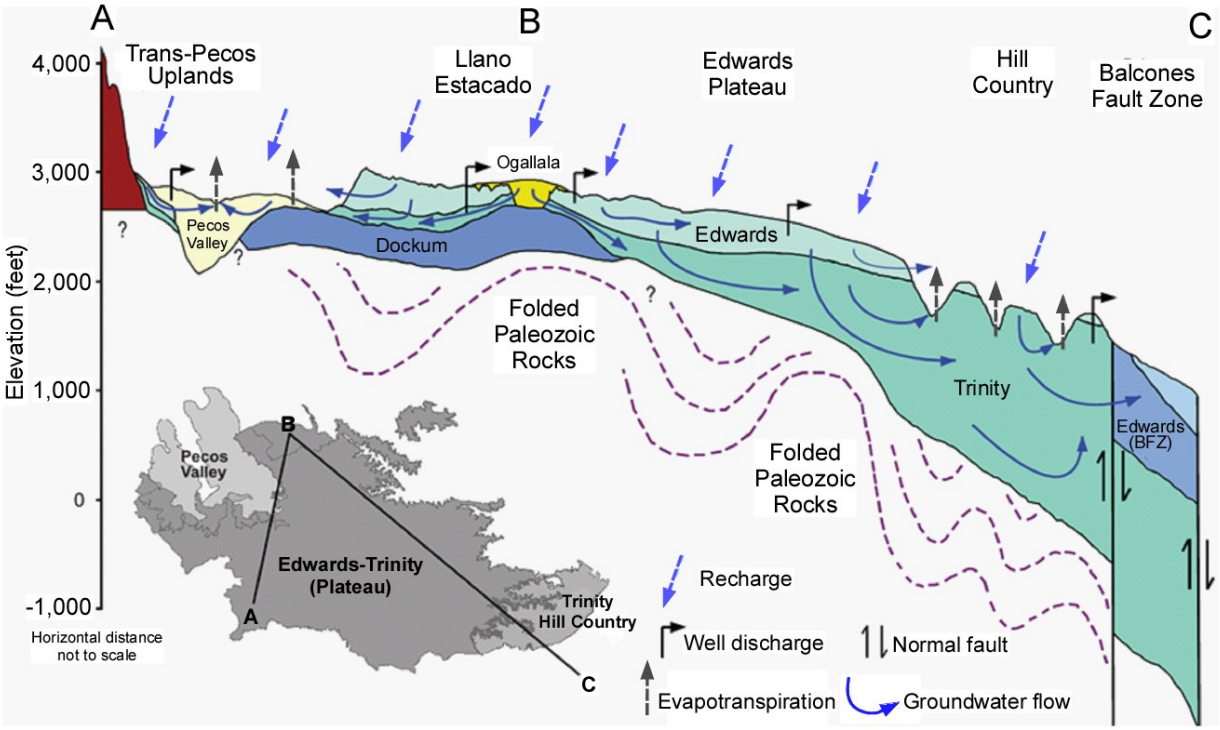


Figure 2.2.3 Conceptual model of the Edwards-Trinity (Plateau) and Pecos Valley aquifers and Hill Country part of the Trinity Aquifer (modified from Anaya and Jones, 2004, 2009). BFZ = Balcones Fault Zone.



Figure 2.2.4 Early Cretaceous limestone and dolomite of the Fredericksburg Group (equivalent to the Fort Terrett Formation) along IH-35 between Bakersfield and Fort Stockton, Texas. From George et al., 2011.

Water quality ranges from fresh to slightly saline, with total dissolved solids ranging from 100 to 3,000 milligrams per liter, and water is characterized as hard within the Edwards Group. Water typically increases in salinity to the west within the Trinity Group. Elevated levels of fluoride in excess of primary drinking water standards occur within Glasscock and Irion counties. Springs occur along the northern, eastern, and southern margins of the aquifer primarily near the bases of the Edwards and Trinity groups where exposed at the surface. San Felipe Springs is the largest spring along the southern margin, and the Balmorhea springs (San Solomon Springs) are the largest in the southwest. Of groundwater pumped from this aquifer, more than two-thirds is used for irrigation, with the remainder used for municipal and livestock supplies (George et al., 2011).



Figure 2.2.5 Looking southwest toward Santa Elena Canyon in the Big Bend National Park. Cretaceous rocks from the Glen Rose Formation to the Santa Elena Formation are well exposed along cliff walls.



Figure 2.2.6 The Cow Creek Limestone of the Edwards-Trinity Plateau Aquifer at Hamilton Pool, Travis County. 2008 Photo Copyright Dave Wilson.'

The Edwards (Balcones Fault Zone) Aquifer, which is geographically also referred to as the Balcones Escarpment, is the most productive karst aquifer within the Edwards-Trinity Aquifer System, with most major springs in Texas (Fig. 2.2.7). It consists primarily of highly permeable, Lower Cretaceous karstified limestone. Area of the aquifer outcrop i.e., the recharge zone, is 1,560 square miles, and area in the subsurface is approximately 2,314 square miles. Aquifer thickness ranges from 200 to 600 feet, and freshwater saturated thickness averages 560 feet in the southern part of the aquifer.

Surface water from springs and streams originating in the catchment area (i.e., the contributing zone) reaches the recharge zone where much of the flow sinks into the Edwards Group Limestone (Fig. 2.2.8). The confined zone (i.e., artesian zone) or that part of the aquifer contained between Del Rio Clay and other less permeable sediments (above) and the Upper Glen Rose Limestone and other less permeable sediments of the Trinity Group (below) has an area of 2,314 square miles (George et al., 2011). Some water also enters the Edwards Aquifer from formations adjacent to the Edwards Group Limestone and from direct precipitation on the recharge zone.

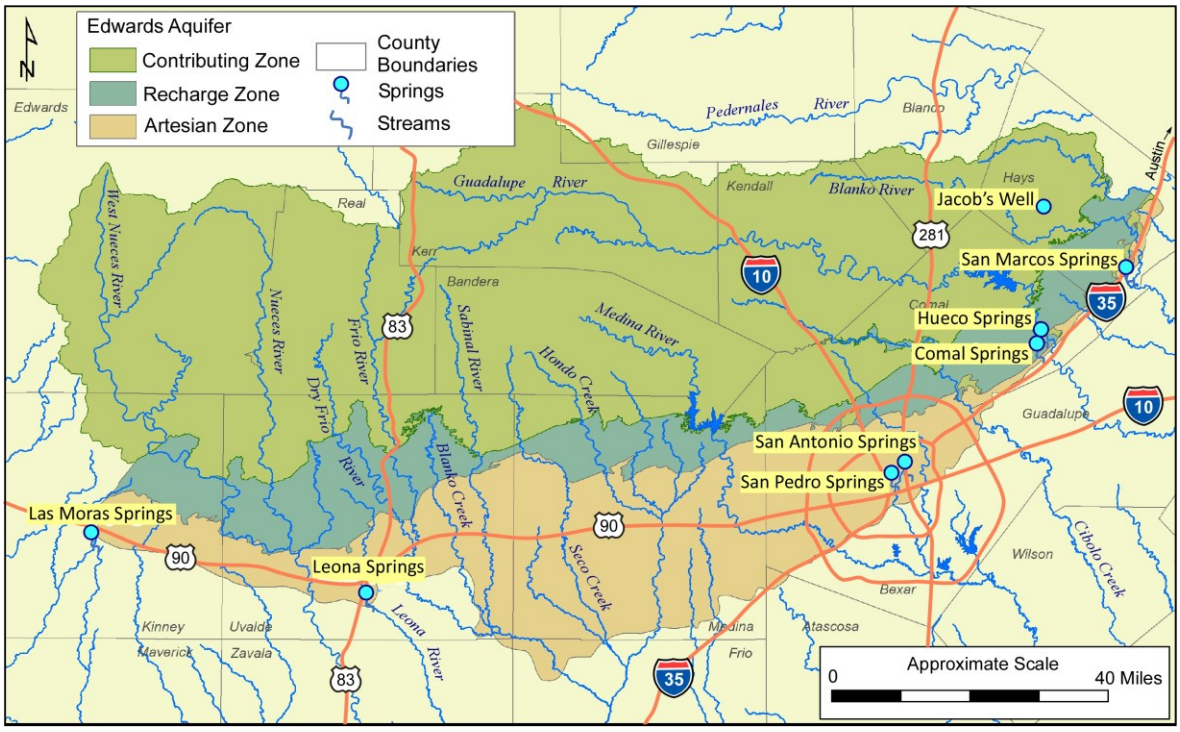


Figure 2.2.7 Major Springs in the San Antonio segment of the Edwards-Trinity (Balcones Fault Zone) Aquifer. Modified from Edwards Aquifer Authority, 2021.

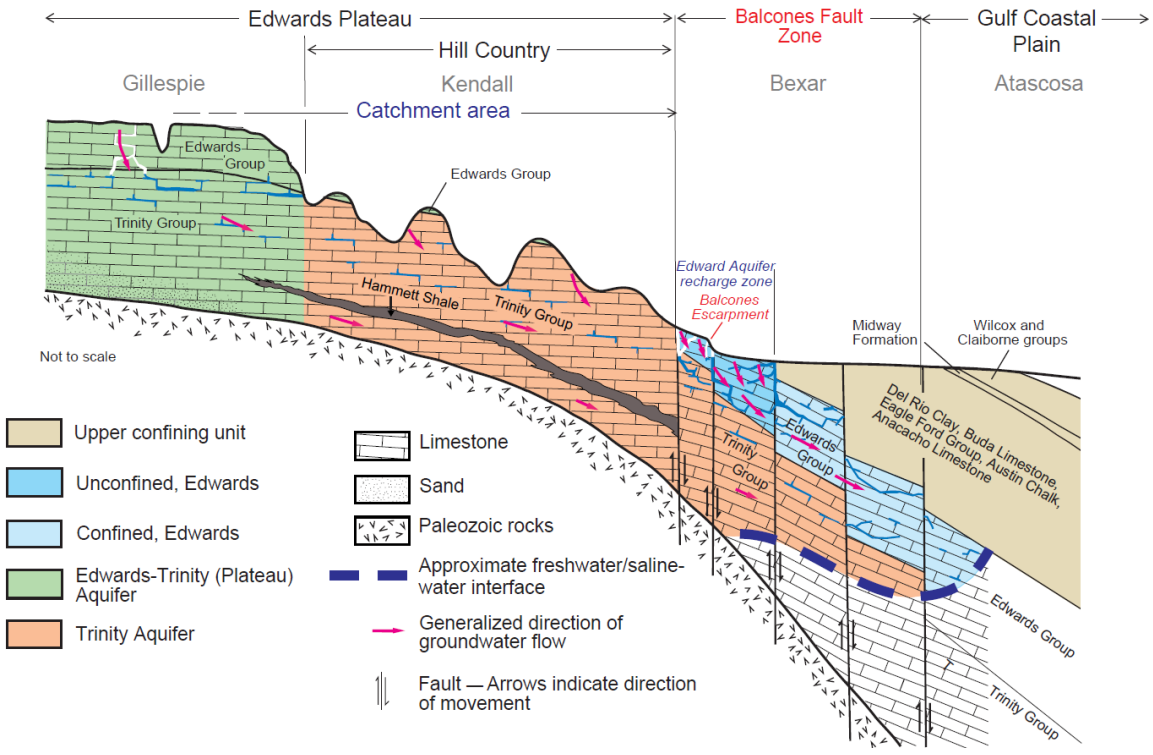


Figure 2.2.8 Diagrammatic cross section showing hydrogeologic framework and generalized groundwater flow through the Edwards (Balcones Fault Zone) Aquifer, San Antonio region, Texas. From George et al, 2011; modified from Barker and Ardis, 1996; Lindgren et al., 2004).

The groundwater, although hard, is generally fresh and contains less than 500 milligrams per liter of total dissolved solids. Water from the aquifer is primarily used for municipal, irrigation, and recreational purposes. San Antonio obtains almost all of its water supply from the Edwards Aquifer (George et al., 2011).

Portions of the artesian (confined) zone are as much as 3,400 feet below the surface where it still contains fresh water. The southern boundary of the artesian zone marks the Aquifer's transition from freshwater to saline water (water with a total dissolved solids concentration greater than 1,000 mg/L).

The residence time of water in the Aquifer ranges from a few hours or days to much longer, depending on depth of circulation, location, presence of karst conduits, and other aquifer parameters. For example, the results of groundwater tracing in the Barton Springs segment demonstrate that a significant component of groundwater flow is rapid, discrete, and occurs in an integrated network of conduits, caves, and smaller dissolution features. Groundwater generally flows from northwest/west to southeast/east within the recharge zone in secondary conduit systems that converge with northeast trending primary conduits defined by troughs in the potentiometric surface parallel to faulting and fracturing (Zappittelo et al., 2019). Groundwater flow is very rapid from recharge features to wells and springs, with velocities ranging from 1 to 7 miles/day depending on hydrologic conditions. Multiple tracing studies since 1996 have further revealed the complexity of groundwater subbasins and helped define the divide between the San Antonio and the Barton Springs segments of the Edwards Aquifer (Figure 2.2.9).

The Edwards Balcones Fault Zone Aquifer is typically divided into two segments mainly because of the groundwater management aspects: the San Antonio segment managed by the Edwards Aquifer Authority (Fig. 2.2.7) and the Barton Springs segment managed by the Barton Springs Edwards Aquifer Conservation District (BSEACD; Fig. 2.2.9).

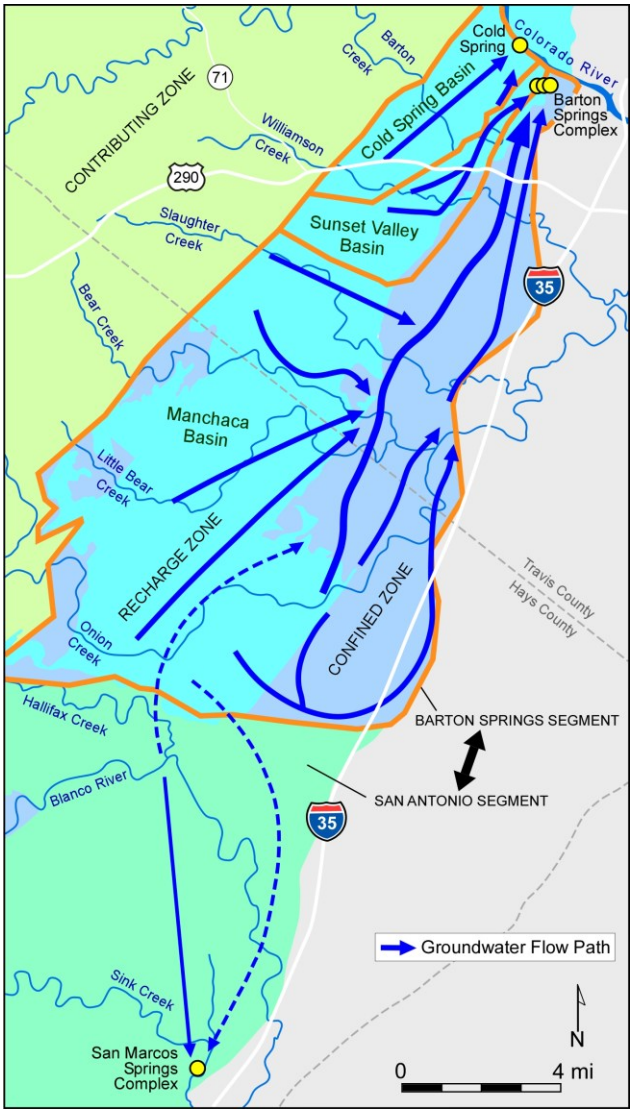


Figure 2.2.9 Barton Springs segment of the Edwards Aquifer. Groundwater basins and flowpaths adapted from Hauwert et al 2004, Hunt et al 2005, Smith et al., 2006, Hauwert 2009, Smith et al., 2012, Hauwert 2012, Hunt et al., 2013, Zappitello and Johns 2018. The boundary between the Barton Springs segment and the San Antonio segment of the Edwards Aquifer varies under different aquifer conditions. The dashed flow paths represent groundwater tracing results that have crossed the groundwater divide. Modified from Zappitello et al., 2019.

The Edwards Aquifer is one of the most productive aquifers in the United States. The competition for groundwater from it has created many controversial water issues in central Texas. It is designated a sole-source aquifer in the San Antonio and Austin areas by the U.S. Environmental Protection Agency. The Aquifer is critical for farming and ranching economies west of San Antonio and recreational economies northeast of the city.

As mentioned in Introduction, unsustainable use of groundwater resources in Texas is rampant, and the ongoing drying of springs is the most obvious consequence. It is evident that Texas regulators, various agencies, developers, and special interests have not learned anything from the past.

On The Edwards Aquifer Website (https://www.edwardsaquifer.net/) one can read the following early descriptions of the main San Antonio Spring, The Blue Hole, courtesy of Gregg Eckhardt:

Before Edwards wells were drilled, early explorers and travelers described The Blue Hole as a fountain spring, with water gushing many feet in the air. In 1857, Frederick Olmsted perceived the entire discharge of the Springs to come from The Blue Hole and he described the natural beauty of the site:

...*The San Antonio Spring may be classed as the first water among the gems of the natural world. The whole river gushes up in one sparkling burst from the earth. It has all the beautiful accompaniments of a smaller spring, moss, pebbles, seclusion, sparkling sunbeams, and dense overhanging luxuriant foliage. The effect is overpowering. It is beyond your possible conceptions of a spring. You cannot believe your eyes, and almost shrink from sudden metamorphosis by invaded nymphdom.*

Richard Everett in 1859 gave this description of San Antonio Springs, as well as of San Pedro springs, the other major natural discharge from the Edwards Aquifer in San Antonio:

Two rivers wind through the city [San Antonio], flowing from the living springs only a short distance beyond the suburbs. One, the San Antonio, boils in a vast volume from a rocky basin, which, environed by mossy stones and overhanging foliage, seems devised for the especial dwelling-place of nymphs and naiads. The other, the San Pedro, runs from a little pond, formed by the outgushing of five sparkling springs, which bear the same name. This miniature lake, embowered in a grove of stately elm and pecan trees, is one of the most beautiful natural sheets of pure water in the Union - so clear, that even the delicate roots of the water-lilies and the smallest pebbles may be distinctly seen.

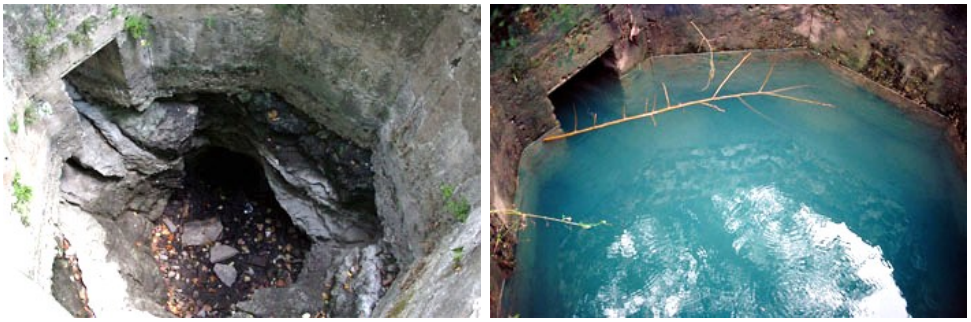


Figure 2.2.10 This is The Blue Hole, described by early explorers and travelers as a "fountain spring", gushing water many feet in the air. John Russell Bartlett described this chasm in 1850: "*The water rises in a cavity some six or eight feet in diameter and twelve or fifteen feet deep, and rushes out in an immense volume. The water of these springs unite with Olmos Creek, forming a river.*" Right photo: The Blue Hole on June 14, 1992, the day the Aquifer stood at its recorded high level of 703.3 feet, as measured by the J17 index well. On that record-setting day, The Blue Hole was a long way from being a "fountain spring". Discharge was still only a tiny fraction of what it must have been when Everett and Olmsted wrote their 19th century descriptions. Notice that you can see several feet below the surface to where the concrete wall ends. Photographs and caption courtesy of Gregg Eckhardt, The Edwards Aquifer Website (https://www.edwardsaquifer.net/saspring.html).

Today, none of the San Antonio Springs flows except during periods of extreme rainfall. The San Antonio River is kept alive mostly by discharges of recycled water in Brackenridge Park, and by one artesian Edwards well inside the San Antonio Zoo.

The Blue Hole is still revered as sacred. Orchard (1983) reported that native peoples conducted ceremonies involving the placement of precious objects near the Springs or in the Springs themselves, and that tradition continues to this day.



Figure 2.2.11 On the Edwards Aquifer Website, Gregg Eckhardt provides the following description of the photograph:
San Antonio began to rely on artesian wells for its water supply several years after the first large wells were drilled in 1891. These are two of San Antonio's first municipal water supply wells, drilled at Market Street for George Brackenridge, who owned the water system and had a contract to supply the city. The photo shows the tremendous amount of pressure that Aquifer water was under at that time. If we estimate the two men in the photo to be around 5 ½ feet tall, then the column of water shooting up from the well is around 25 feet high! The effect of releasing all this pressure through wells was that springflows began to decline immediately and significantly. By 1896 there were approximately 40 wells in the San Antonio area. By around 1900 San Antonio Springs had been reduced to just a trickle in most years. This photograph appeared in R. T. Hill & T. W. Vaughan's 1896 report on the geology and underground waters of the Edwards Plateau. Hill and Vaughan were the first geologists to recognize that wells such as these had impacted springflows. They were the first people to accurately describe the Edwards and how it works. Although they never used the word 'aquifer', they referred to the Edwards as an artesian groundwater system, accurately described the catchment and transmission of water in the Aquifer, and recognized its large extent from Brackettville to Austin.

San Marcos and Comal Springs, the only remaining first-magnitude springs in Texas, were spared the same destiny because of the 1991 lawsuit brought up by The Sierra Club against U.S. Fish and Wildlife Service (USFWS) and the state of Texas. The Sierra Club sued under the Endangered Species Act (ESA), arguing that the state water policy was driving endangered Texas species native only to the San Marcos and Comal Springs toward extinction. The Sierra Club won the lawsuit, and the groundwater use in the San Antonio segment of Edwards Aquifer changed drastically, saving the two springs, for now. Unfortunately, many other large and precious springs in Texas were not that lucky, as explained in Chapter 3.2.

Currently, the habitat protection measures in place for Comal and San Marcos Springs apply to six federally endangered species: the Texas Blind Salamander, Fountain Darter, Comal Springs Dryopid Beetle, Texas Wild-rice, Comal Springs Riffle Beetle, and Peck's Cave Amphipod; and one threatened species, San Marcos Salamander (see Figures 2.2.12 through 2.2.14). San Marcos Gambusia (*Gambusia georgei*) is now officially extinct and was delisted in October 2023. A federally listed species is classified as **endangered** if it is at immediate risk of extinction and **threatened** if it is at risk of extinction in the foreseeable future.

The primary threat to these aquifer dependent species is the intermittent loss of habitat from reduced springflows. Springflow loss is the combined result of naturally fluctuating rainfall patterns, natural discharges at other springs, and regional pumping and drawdown of the Edwards Aquifer.

The Sierra Club lawsuit under the ESA resulted ultimately in the creation of the Edwards Aquifer Authority (EAA). The Texas Legislature directed the EAA to regulate, among other things, pumping from the Aquifer, to implement critical period management restrictions, and to pursue a program "to ensure that the continuous minimum springflows of the Comal Springs and the San Marcos Springs are maintained to protect endangered and threatened species to the extent required by federal law" (EAA Act § 1.14(h)).



Figure 2.2.12 *Left*: Comal Springs Dryopid Beetle (*Stygoparnus comalensis*). *Middle*: Peck's Cave Amphipod (*Stygobromus pecki*). Courtesy of U.S. Fish & Wildlife Service. *Right*: Texas Blind Salamander (*Eurycea* [formerly *Typhlomolge*] *rathbuni*).



Figure 2.2.13 *Left*: Texas Wild rice (*Zizania texana*). The upper stems of Texas wild rice can emerge above the water surface. The male and female flowers occur separately on side branches of a flowering stalk. The male flowers dangle lower down the flowering stalk than the rigid female flowers. Credit: Jackie Poole - Texas Parks & Wildlife Department. *Middle*: Texas wild-rice roots underwater, and the stems bend, flowing parallel to the current. The leaves have a prominent main vein running down the middle of the blade. Credit: John Thomaides - Texas Parks & Wildlife Department. *Right*: San Marcos Salamander (*Eurycea nana*). Courtesy of U.S. Fish & Wildlife Service.



Figure 2.2.14 *Left*: Fountain darter (*Etheostoma fonticola*). Before being inundated by a small dam and lake, the main spring at San Marcos was described as emerging like a fountain from a limestone basin. *Right*: Comal Springs Riffle Beetle (*Heterelmis comalensis*). Courtesy of U.S. Fish & Wildlife Service.

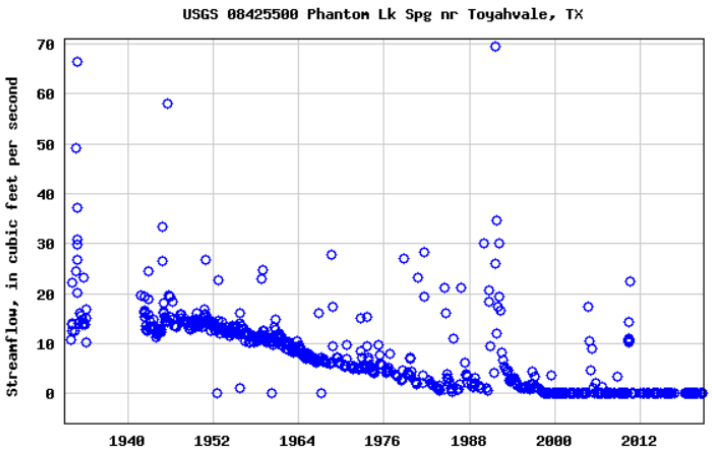


Figure 2.2.15 Hydrograph of discharge rate of Phantom (Lake) Spring near Toyahvale, Reeves County for the period of record. Spring dried up in 1999 because of excessive groundwater pumping. It was the official home to five endangered species: Comanche Springs pupfish (*Cyprinodon elegans*), Pecos gambusia (*Gambusia nobilis*), Phantom tryonia (*Tryonia cheatumi*), Phantom springsnail (*Pyrgulopsis texana*), and diminutive amphipod (*Gammarus hyalelloides*).



Figure 2.2.16 *Left*: Valdina Sinkhole on the recharge zone of the Edwards Aquifer in northwest Medina County during recharge when Seco Creek water is clear. Photo courtesy of USGS. *Middle*: The drop in Valdina Sinkhole when dry. Photo courtesy of Dave Bunnell; all rights reserved; printed with permission. *Right*: Cave passage at the bottom of Valdina Sinkhole when dry. Photo courtesy of Geary Schindel.



Figure 2.2.17 When drilled, this was the highest capacity well in the world. "*There is tremendous artesian pressure at this location. When this well came in, it blew out rocks the size of basketballs 20 feet into the air. The well was massive 30" in diameter and capable of producing 40 thousand gallons per minute."* (2.5 m^3/sec). It was located on the Ron Pucek's Catfish Farm near San Antonio, Texas. The fascinating story about the well is available at Edwards Aquifer Website: Ron Pucek's Living Waters Catfish Farm. Photo copyright by Gregg Eckhardt; printed with permission.



Figure 2.2.18 On January 24, 2012, a 4.5-inch rainfall filled a Storm Water Retention Pond (SWRP) located in the recharge zone of the Barton Springs segment of the Edwards Aquifer with about 10 feet of stormwater. Subsequently, a cover-collapse sinkhole developed within the floor of the SWRP, measuring about 30 ft in diameter and 12 ft deep. About 7 million gallons of stormwater drained into the aquifer through this opening. To determine the path, velocity, and destination of stormwater entering the sinkhole, a dye trace was conducted. Phloxine B was injected into the sinkhole on February 3, 2012. A mass of 16.27 lbs (7,382g) was mixed with water and then gravity injected via a hose using storm water from an adjacent pond. The dye was detected at one well and arrived at Barton Springs in less than 4 days, corresponding to a minimum velocity of 1.3 mi/day. The successful dye trace confirmed conclusions of previously published reports by demonstrating that the sinkhole is well integrated into the aquifer system, and that groundwater in the study area is within the Sunset Valley Groundwater Basin. Phloxine B proved to be a very good, conservative tracer through the collapsed terra rosa material of the sinkhole. Text and photograph from Hunt et al., 2013.





Figure 2.2.19 *Left*: Spelunker examines trash around a flowstone waterfall at the bottom of the entrance to Midnight Cave near Austin, Texas, November 20, 1993. The trash includes household garbage, used oil filters, corroded 55-gallon drums, glass pesticide bottles, partially filled turpentine cans, and automobile parts. Note the trash on the higher ledges of the cave. (From Hauwert and Vickers, 1994.) *Right*: Canyon carved in the Lower Glen Rose Fm., Hays County. Note hydrogeologist in center top for scale. Photo courtesy of Brian A. Smith, available at https://www.austingeosoc.org/geo-photos/

2.3 Madison Limestone Aquifer

The Madison Aquifer, commonly referred to as the Madison Limestone Aquifer, is a massive, deep-seated groundwater system that spans a vast region across Northern Great Plains of the United States and Canada. It underlies approximately five states in the United States: Montana, North Dakota, Wyoming, South Dakota, and Nebraska, and three provinces in Canada: Alberta, Saskatchewan, and Manitoba. It is primarily composed of Mississippian-age carbonate rocks (limestone and dolomite). It outcrops in mountainous areas of the Black Hills in western South Dakota and northeastern Wyoming, the Little Belt Mountains in Central Montana, and the Bighorn Mountains in north-central Wyoming and southern Montana (see also Fig. 2.1).

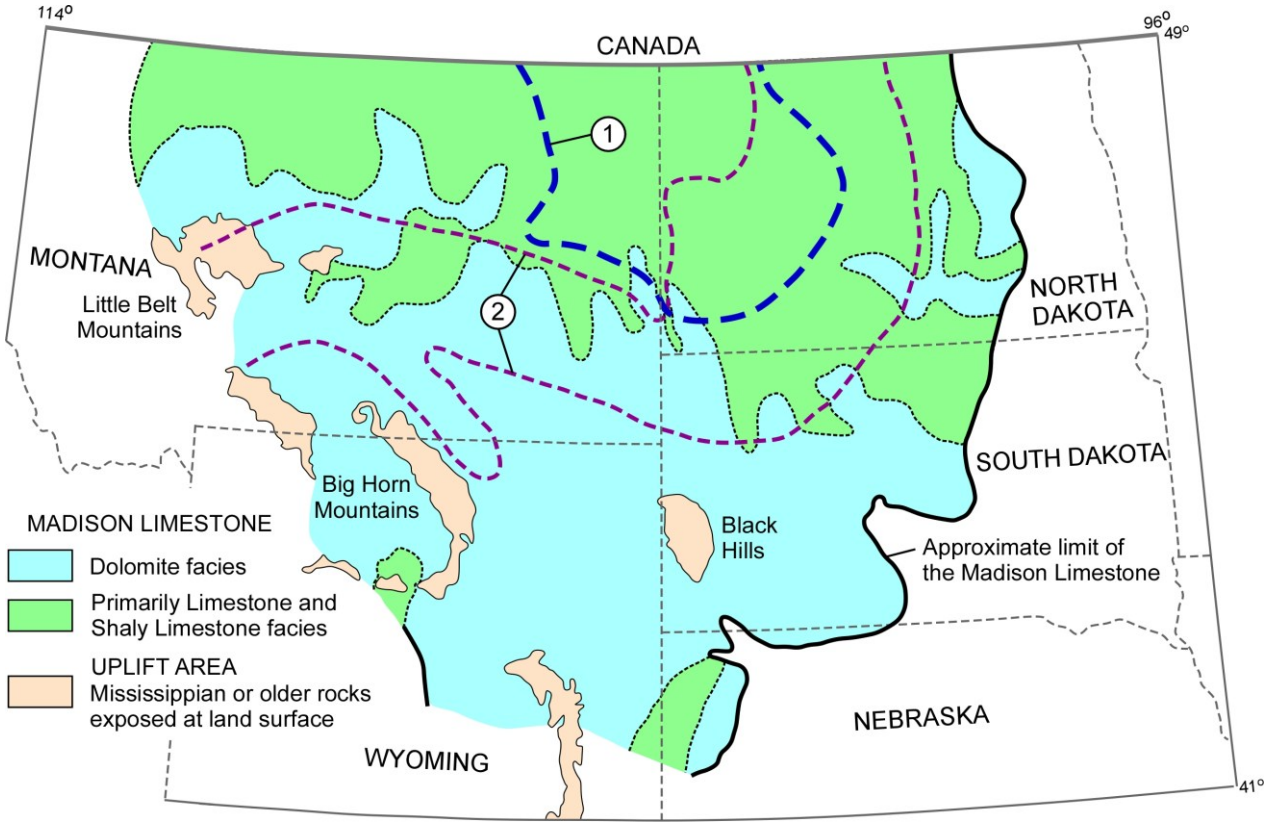


Figure 2.3.1 Lithology of the buried Madison Limestone of Mississippian age. (1): Limit of Charles Salt of informal subsurface usage Sawteeth face salt. (2): Limit of Anhydrite in the middle part of the Madison Limestone Hachures face Anhydrite. Modified from Downey, 1984.

The Madison Limestone consists of a sequence of marine carbonates and evaporites deposited mainly in a warm, shallow-water environment. Depositional environments grade both laterally and vertically from shallow-marine carbonate and evaporite facies to deep-water facies. For the region shown in Fig. 2.3.1, the Madison Group, from oldest to youngest, consists of the Lodgepole Limestone, the Mission Canyon Limestone, and the Charles Formation, or their stratigraphic equivalents in other parts of its vast extent in the western United States. The aquifer is underlain by the Lower Mississippian Bakken Formation which consists of more than 100 ft of black, organic shale and siltstone and acts as an excellent confining bed. The Charles Formation, the uppermost unit of the Madison Group, is a marine evaporite sequence consisting of anhydrite and halite with interbedded dolomite, limestone, and argillaceous units. The Charles Formation ranges from 0 to more than 300 ft in thickness, with an

average thickness of about 250 ft. Pre-Jurassic erosion has removed most of the Charles Formation in the western and southern parts of the area shown in Fig. 2.3.1. Where present, the Charles Formation is a hydrogeologic confining bed for the underlying Madison Formation Aquifer (Downey, 1984).

Because the carbonate rocks of the Madison Limestone are soluble in water, the development of karst (solution) features is common. Sando (1974) describes ancient karst features, including enlarged joints, sinkholes, caves, and solution breccias, that developed in the Madison Limestone in north-central Wyoming. He indicates that most of the open spaces were filled by sand and residual products reworked by a transgressive sea during Late Mississippian time. Large and extensive cave systems in outcrop areas of the Madison Limestone in the Bighorn Mountains and in the Black Hills are further evidence of the importance of the dissolution process in the development of secondary permeability in some of the units of the Madison Limestone.



Figure 2.3.2 Madison Limestone exposed at canyon cliffs along the Missouri River east of Helena, Montana, named the "Gates of the Mountains" by Lewis and Clark. Photo courtesy of https://gatesofthemountains.com/about/

Many of the groundwater reservoirs in the Madison Limestone consist of porous zones in fine- to medium-grained limestone or dolomite. Studies of pressure relationships between these porous zones (Landes, 1970) indicate that in some instances they are interconnected by fractures that cut the less permeable rocks between the zones. Fracture interconnection between zones of greater permeability appears to be the major route of groundwater flow in the Madison aquifers, as well as the deeper Cambrian-Ordovician aquifers (Downey, 1984).

More broadly, the Madison Limestone (or Madison Group) is part of a massive, continent-wide sequence of Mississippian-age carbonate rocks. Because these layers were deposited in a vast, warm, shallow sea that covered much of North America, the same formation is found under different names across the Western United States and beyond. For example, Redwall Limestone is the name used in the Grand Canyon and northern Arizona; Leadville Limestone is the equivalent formation found in central and western Colorado; and Gardison Limestone is used specifically in the East Tintic Mountains of Utah.



Figure 2.3.3 *Left*: Artesian well completed in Madison Limestone in Yellowstone County, Montana. Photo by L.W. Howels, USGS, 1979. *Right*: The Crushing Deep is currently one of the deepest known parts of Jewel Cave in S. Dakota, the second longest cave in the United States (after Mammoth Cave in Kentucky) with 220.33 miles of surveyed passages as of September 7, 2024. The elevation range of Jewel Cave spans 822 feet, which is its official depth. Jewel Cave is formed in the Pahasapa Limestone, which is correlated with the Madison's Mission Canyon and Lodgepole formations found further west and regionally together known as the Madison Limestone. Photo courtesy of NPS.

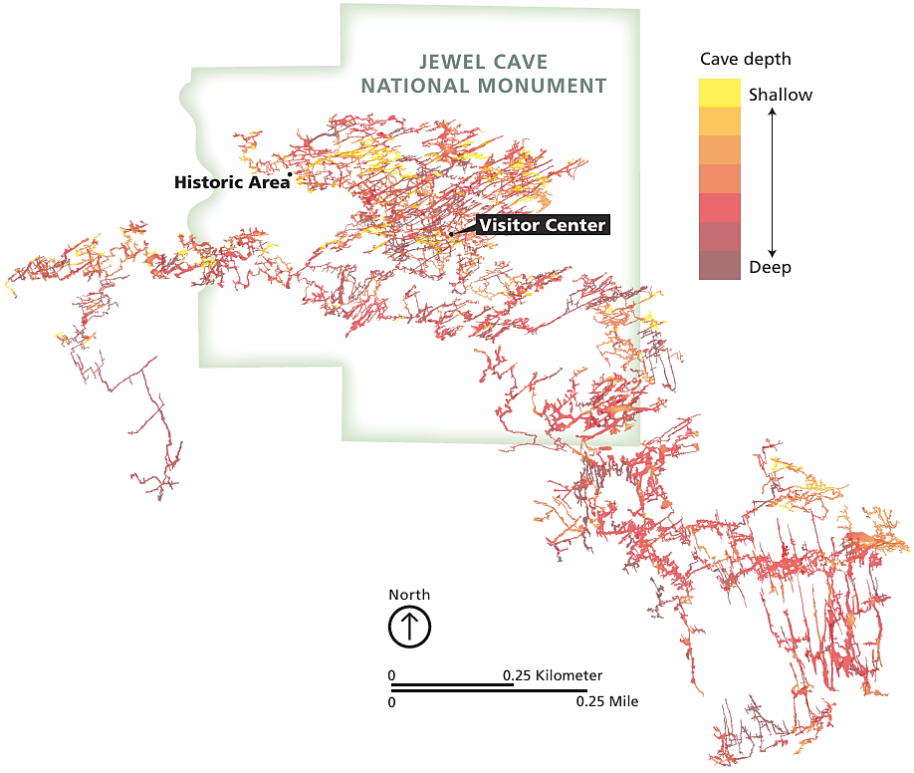


Figure 2.3.4 Map of surveyed passages of Jewel Cave. Courtesy of NPS.
https://npmaps.com/wp-content/uploads/jewel-cave-layout-map.pdf

2.4 Basin and Range, and Bear River Range Carbonate Aquifers

All large springs in Nevada and many in Utah are issuing from the Basin and Range carbonate-rock aquifers. Although the Basin and Range Province is primarily structural, faulting has been accompanied by widespread volcanism which, together with the complicated tectonics and the related deep groundwater circulation, is the main reason for countless thermal springs in the Basin as well.

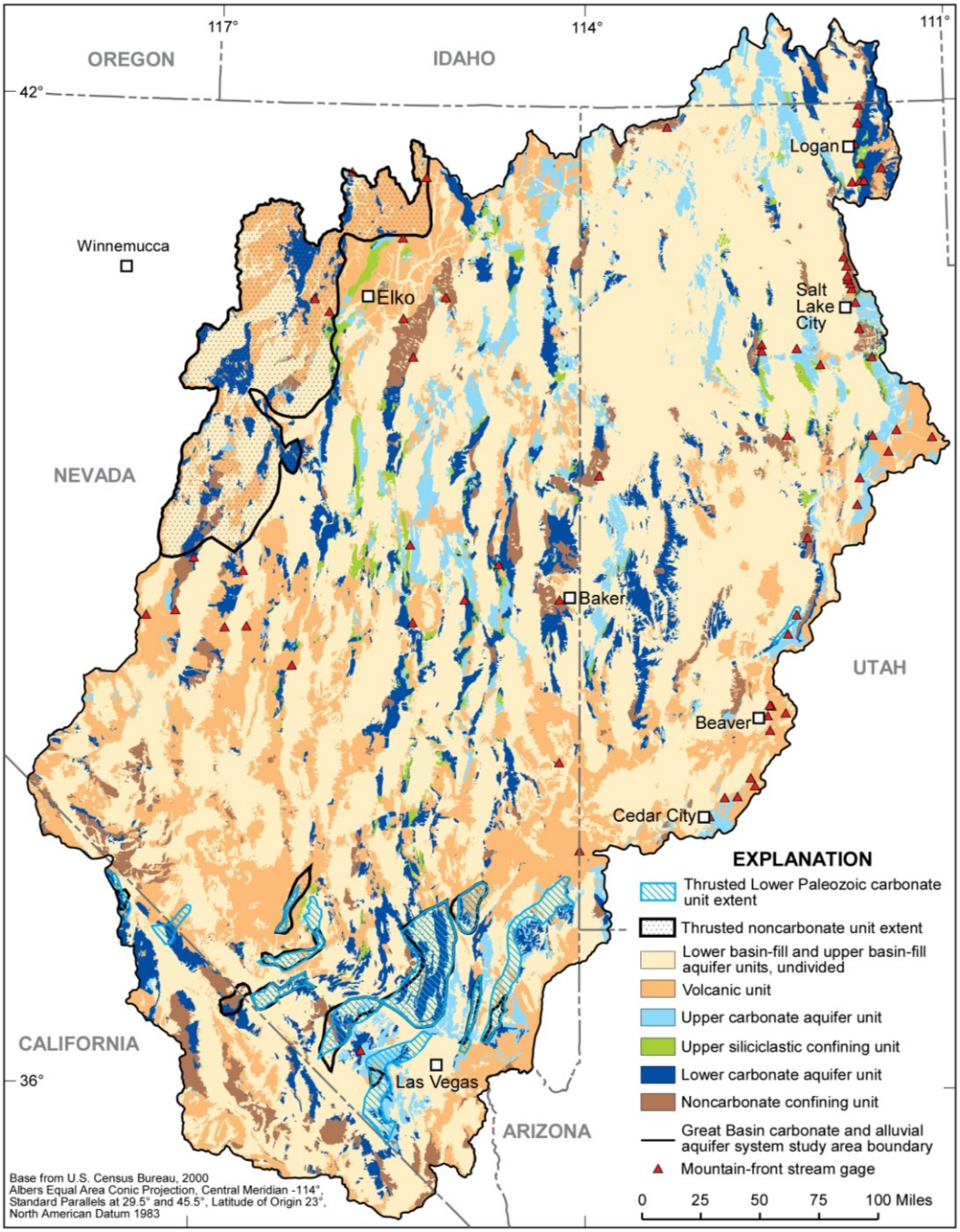


Figure 2.4.1 Surficial hydrogeologic units of the Great Basin carbonate and alluvial aquifer system. Modified from Sweetkind et al. (2011).

The Great Basin comprises the entire northern portion of the Basin and Range physiographic province. It contains parts of southwestern Oregon, southeastern Idaho, western Utah, and nearly the entire state of Nevada. “Basin” in the name reflects the fact that all groundwater is internally drained and does not discharge to the ocean.

Within the Basin and Range province, the Earth's crust has been stretched up to twice of its original width. As it expanded, the crust thinned and cracked, creating large faults. Along the trending faults, mountains were pushed up, and valleys carved below creating a distinctive alternating pattern of ranges and valleys which are typically 10 to 20 miles across. The geology, topography, and dry climate create impressive features and landscapes. These include pediments, alluvial fans, bajadas, bolsons, inselbergs, playas, mud flats, salt flats, lakes, sand dunes, and canyons, many of which are in various national and state parks and national monuments (National Park Service, https://www.nps.gov/articles/basinrange.htm).

The carbonate-rock aquifers store a large amount of water and lie below basin-fill aquifers and in ranges adjacent to them. They are viewed as a potential future source of water in this dry climate environment, including for ever-expanding urban centers such as Las Vegas.

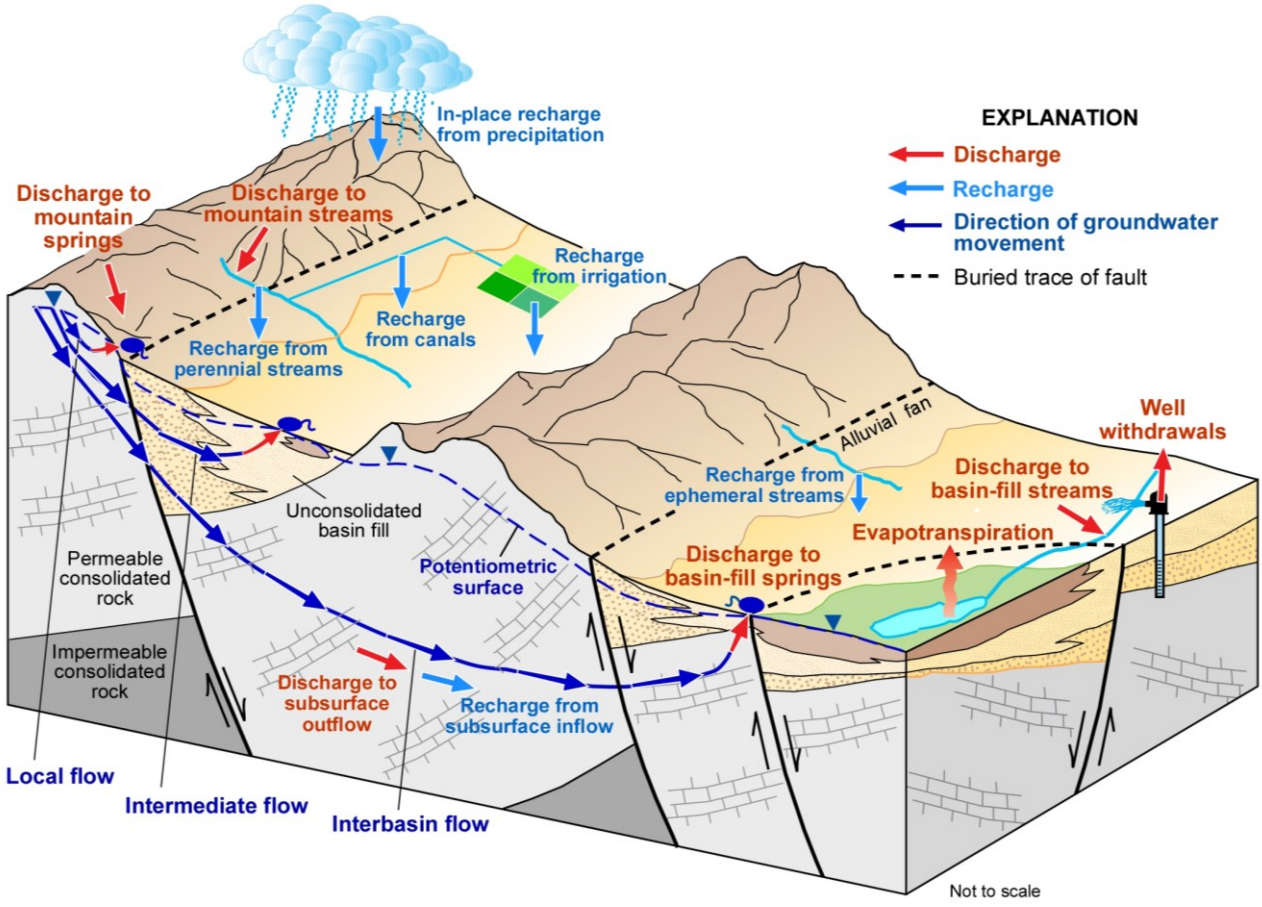


Figure 2.4.2 Schematic diagram showing conceptualized groundwater flow in the Great Basin carbonate and alluvial aquifer system. Modified from Sweetkind et al., 2011.

As explained by USGS (Thiros et al., 2014), the Basin and Range carbonate-rock aquifers cover an area of about 92,000 square miles, mostly in Nevada and Utah (see map in Fig. 2.4.1). The carbonate-rock aquifers consist of thick sequences of Paleozoic limestone and dolomite interbedded with shale, sandstone, and quartzite. A regional

assessment of the availability of groundwater in both the basin-fill deposits and carbonate rocks is presented in Heilweil and Brooks (2011).

In Nevada and Utah, carbonates and other sedimentary rocks were deposited on the continental shelf of the ancestral coast of North America between 570 million and 280 million years ago. The layers of ancient marine sediments have a cumulative thickness of as much as 40 thousand feet. All the sedimentary rocks in this sequence, where deformed and fractured, may transmit some groundwater flow. It is the carbonate-rock layers, because of their brittleness and dissolution by flowing water, that are the principal aquifers (Dettinger, 1989). Due to tectonic forces (faulting and folding), these carbonate aquifers have complex shapes and are connected to aquifers of other rock types. Some of the blocks may be highly productive aquifers, whereas others may transmit only moderate quantities of water or impede flow altogether. Regional groundwater flow, through both the carbonate-rock aquifers and basin-fill aquifers, is from the areas of recharge (rain and snow on the mountain ranges) towards the areas of discharge (springs, playas).

An example of one of the many threats facing springs and groundwater resources in the Great Basin is the fight by locals in Spring and Snake Valleys in Nevada and Utah to protect their water from the relentless thirst of Las Vegas. Various environmental and citizens groups challenged a proposed northeastern expansion of Southern Nevada Water Authority (SNWA) water supply wells and a 300-miles long pipeline.



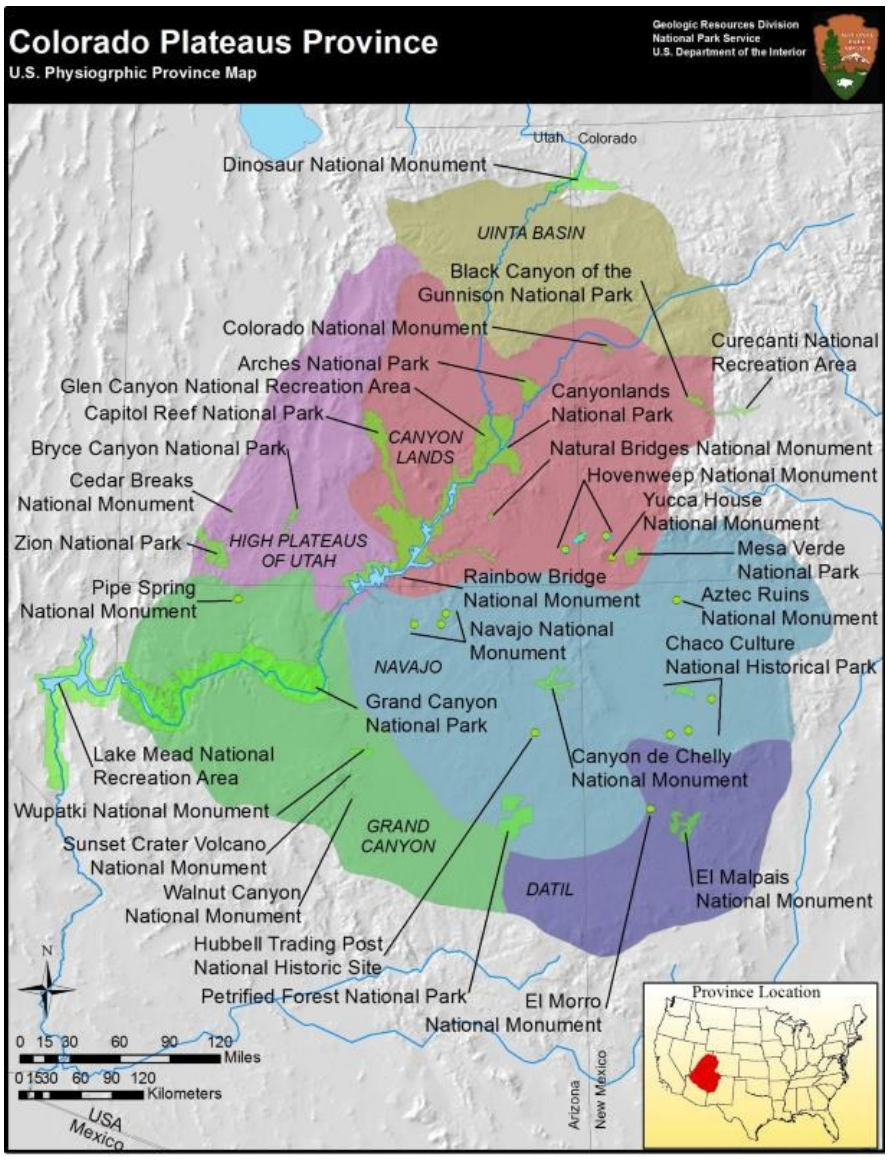
The alpine Bear River Range in NE Utah and SE Idaho consists in large part of a thick sequence of carbonate rocks (more than 3,000 feet of limestone and dolomite) that range in age from Cambrian to Mississippian (Dover, 1987). The sequence is composed of eight principal geologic units, all of which are capable of transmitting water along dissolution-enhanced fractures, faults, and bedding planes. The formations make up the upper part of the Logan Peak syncline, a large regional structure that influences the movement of groundwater in much of the region.

Figure 2.4.3 A spring in Spring Valley (which is adjacent to Snake Valley) that may be affected if the proposed expansion of Southern Nevada Water Authority is revived. Photo courtesy of Gretchen Baker.

In the Bear River Range, karst features include large springs that discharge along major rivers, losing streams in tributary drainages, caves and pits, blind valleys, sinkholes, dolomite pavement, and surficial karst (karren). Glaciation occurred above 8,000 ft during the Pleistocene, resulting in destruction of karst landforms that developed during interglacial periods. Speleothem age-dating, fluvioglacial deposits in caves, and deranged topography indicate that existing karst features, particularly caves, are largely remnants of former karst landscapes. Karst systems in alpine terrains are substantially different from those in relatively flat-lying strata in more temperate regions. Characteristics of alpine karst systems include a large component of vertical solution development and a thick unsaturated (vadose) zone, steep hydraulic gradients, spring discharge that responds primarily to snowmelt runoff, pit development in high-altitude meadows, and cold-temperature dissolution of carbonate rocks (USGS, 2021).

2.5 Colorado Plateaus Karst Aquifers

As explained by the National Park Service, the Colorado Plateaus physiographic province is the world's premier natural showcases for Earth history, featuring numerous national and state parks, monuments, and historic sites (Fig. 2.5.1). Encompassing 130,000 square miles, the Colorado Plateaus straddle the region known as The Four Corners, where the states of Arizona, Utah, Colorado, and New Mexico meet. Ancient volcanic mountains, plateaus and buttes, deeply carved canyons, and amazing ranges in color are the region's defining characteristics. Elevation for the Colorado Plateaus starts at about 610 m (2,000 ft) above sea level, with plateau tops ranging from 1,534 to 2,134 m (5,000 to 7,000 ft) and mountaintops reaching nearly 3,960 m (13,000 ft).



Ancient Precambrian rocks, exposed only in the deepest canyons, make up the basement of the Colorado Plateaus. Igneous rocks injected millions of years later form a marbled network through parts of the Colorado Plateaus' darker metamorphic basement. These rocks formed deep beneath the surface of the earth and were uplifted, eroded, and exposed for eons. By 600 million years ago North America had been eroded to a remarkably smooth surface. It is on this crystalline rock surface that the younger, more familiar layered rocks of the Colorado Plateaus were deposited.

Figure 2.5.1 National parks, monuments, and historic sites in the Colorado Plateau physiographic province. Courtesy of NPS.

The extent of Colorado Plateaus aquifers is approximately coincident with the Colorado Plateaus Physiographic Province. The distribution of aquifers in the Colorado Plateaus is controlled in part by the structural deformation and erosion that has occurred since deposition of the sediments that compose the aquifers. The principal aquifers in younger rocks are present only in basins such as the Uinta, Piceance, and San Juan Basins. In uplifted areas, such

as the Monument and Defiance Uplifts and the Coconino Plateau, younger rocks have been eroded away, and aquifers are present in older rocks that underlie more extensive parts of the Colorado Plateaus area. Although the quantity and chemical quality of water in the Colorado Plateaus aquifers are extremely variable, much of the land in this sparsely populated region is underlain by rocks that contain aquifers capable of yielding usable quantities of water of a quality suitable for most agricultural or domestic use (Robson and Banta, 1995).

Carbonate formations on the Colorado Plateau, notably the Redwall-Muav, and Kaibab Limestones, are important karst aquifer systems. The Redwall-Muav Aquifer, best known from the Grand Canyon region, is a prolific karst aquifer formed in limestone and dolomite rock. It is the source of some of Arizona's most valuable springs and streams, including the Little Colorado River, Havasu Falls, and Roaring Springs. However, this aquifer lies over a mile beneath the surface in many areas, making it extremely difficult and expensive to access for municipal use. Nonetheless, it is essential to maintain flows of perennial streams and springs that support life in the Grand Canyon region (https://www.grandcanyontrust.org/blog/).

The Cambrian Muav Limestone is composed of fossiliferous limestone, silty limestone, dolomite, and mudstone with interbeds of sandstone, siltstone, mudstone, and shale. It ranges from about 200 ft thick in the eastern part of Grand Canyon to more than 600 ft thick in the western part of the canyon (Billingsley, 2000; Pool et al., 2011).

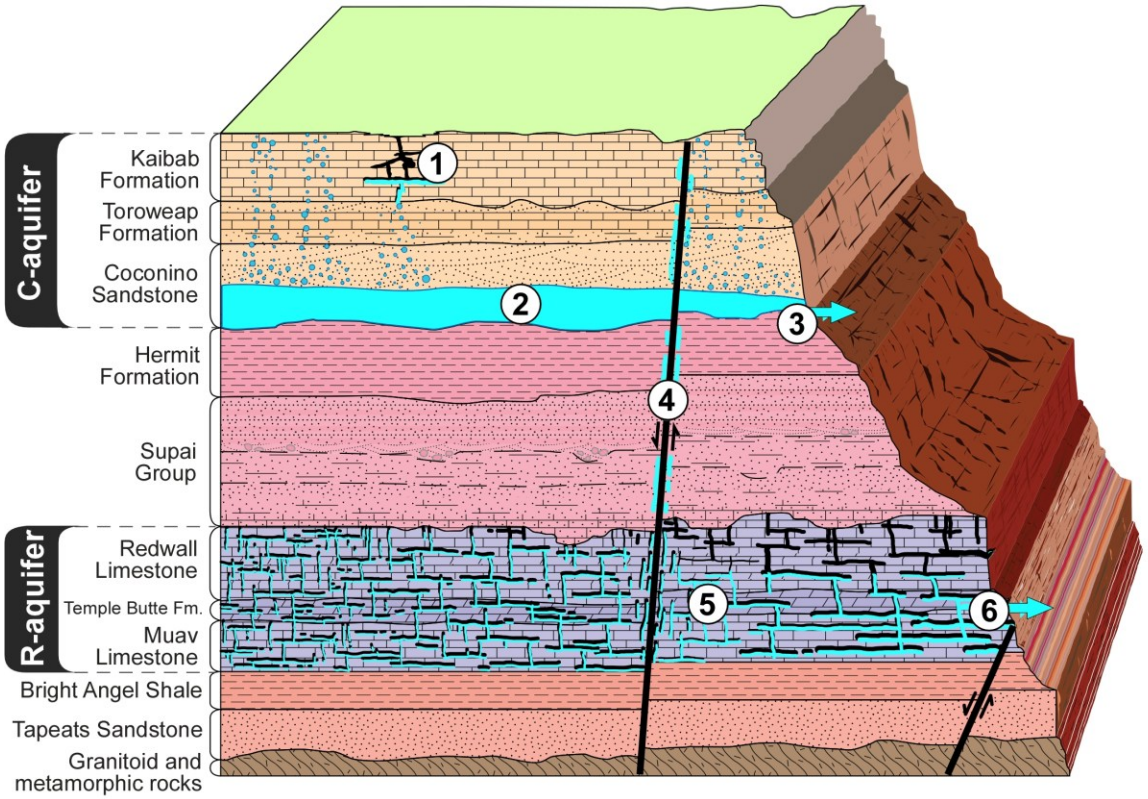


Figure 2.5.2 Conceptual block diagram of hydrogeologic units in Grand Canyon area. (1) Occurrence of recharge to C-aquifer can be widely distributed in space but is likely focused around areas of highly developed epikarst. (2) The majority of groundwater storage in the C-aquifer occurs in the Coconino Sandstone overlying the Hermit Formation confining unit. (3) Discharge from the C-aquifer occurs at springs exposed in canyon walls, or (4) as downward flow to the R-aquifer where substantial fracturing breaks the confining interval (aquitard). (5) Groundwater flow in the R-aquifer occurs via interconnected voids that have dissolved and widened over geologic time. (6) Discharge from the R-aquifer occurs at springs exposed in canyon walls near/at bottom of streams. Modified from Knight and Huntoon, 2022.



Figure 2.5.3 Redwall Limestone in Grand Canyon area. *Left:* Aerial view northwest toward Redeye cave in north wall of Diamond Creek Canyon. Red sediment that fills the cave is Surprise Canyon Formation. The cave is in Mooney Falls Member of Redwall Limestone. From Grand Canyon collection of Billingsle et al., 2019; USGS. *Right* photo courtesy of David Kreamer. Although its name implies that the Redwall Limestone is a red rock, it is in fact a gray stone. It gains its reddish appearance from the rocks above. Both the Supai Group and the Hermit Formation are red in color due to their high iron oxide content. Over time, rain has continually washed these eroding red sediments over the Redwall, staining it red. The Redwall Limestone contains thousands of caves.

The Early Mississippian Redwall Limestone (equivalent of Madison Limestone) is a light-gray to gray, fine to coarse crystalline bedded limestone with dolomite and thin beds and lenses of chert (Beus, 1989). Some beds in the formation are fractured, and the limestone typically contains solution cavities and caverns, some of which have collapsed or filled with sediment. The Redwall Limestone ranges from about 250 to 750 ft in thickness and yields water to wells (Pool et al., 2011). Together, the Muav Limestone and the Redwall Limestone are called R-aquifer in the greater Grand Canyon area. It is a confined aquifer for most of its occurrence with hydraulic heads of several hundred to more than 500 feet above the top of the aquifer in the western part of the Coconino Plateau and more than 2,000 feet above the top of the aquifer in the area near Flagstaff south of the Grand Canyon.

Other Mississippian Madison Limestone equivalents on the Colorado Plateau are the Leadville Limestone (eastern Utah, western Colorado); Deseret Limestone/Gardison Limestone (Central Utah); Modoc Limestone (parts of southeastern Arizona), and Escabrosa Limestone (southeastern Arizona/southern edge of the plateau).



As described by Lance Weaver on his exceptional webpage https://Utahgeology.com, the Early Permian Kaibab Formation forms the rim of the Grand Canyon and the surface of the Kaibab Plateau. Prime exposures occur at Zion National Park (the base of the cliffs), the Hurricane Cliffs in Utah, and San Rafael Swell, a geologic uplift in south-central Utah.

Figure 2.5.4 Kaibab Limestone forms the topmost or caprock of the Grand Canyon. Photo courtesy of Lance Weaver. Available at Kaibab Limestone (Geology of The Grand Staircase) - UtahGeology

Lateral equivalents of the Kaibab Formation are Park City Formation (Northern UT, WY); Phosphoria Formation (ID, MT, WY); San Andres Limestone (NM, TX); and Toroweap Formation (underlying/intertongued).

The Kaibab Limestone is a complex sedimentary package of interbedded and interfingering gypsum, limestone, dolomite, chert, siltstone, and sandstone that is 300–400 ft thick. Erosion-resistant layers of limestone and dolomite form steep cliffs and the rims of the Grand Canyon and its tributary canyons. They also underlie most of the expansive surface of the Kaibab Plateau surrounding the Grand Canyon. Less erosion-resistant sandstones, siltstones, and cherts form distinct recesses along cliff faces (Kaibab Limestone (Geology of The Grand Staircase) - UtahGeology).

Although the Kaibab Formation typically is unsaturated, except for perched zones and north of the Little Colorado River, the formation provides a conduit for infiltration and percolation of precipitation and surface water, including recharge to the deep-seated R-Aquifer (Bills et al., 2000).



Figure 2.5.5 A sinkhole in the Kaibab Limestone east of Flagstaff was once referred to as the 'Bottomless Pit'. Early in the 20th century a plug must have existed in the sinkhole allowing a lake to form. One day the plug failed and the lake waters disappeared into the sinkhole. The sinkhole is now fenced off to prevent access by the public. This photo appeared in Darton er al., 1916, Guidebook of the Western United States: Part C - The Santa Fe Route, with a side trip to Grand Canyon of the Colorado (USGS Bulletin 613. See also Harris, 2002). Available at https://pubs.usgs.gov/bul/0613/report.pdf



Figure 2.5.6. Oblique aerial photograph looking north at McCauley Sinks and Chevelon Canyon, approximately 17 miles southeast of Winslow, Arizona. Photograph by Jon Mason, U.S. Geological Survey, June 27, 2020.

2.6 Roswell Aquifer

The Roswell Artesian Basin consists of an eastward-dipping carbonate aquifer overlain by a leaky evaporitic confining unit of the Artesia Group, overlain in turn by an unconfined alluvial aquifer. The aquifer is developed in the San Andres Formation of Permian age which consists of about 1,000 ft of gray limestone, dolomite, and thin to massively bedded carbonate mudstones, wackestones, and grainstones (Rawling, 2009; Darr et al., 2014). The carbonate aquifer is artesian to the east but under water table conditions in the western outcrop area where there are also numerous springs.

Water-producing zones in the carbonate aquifer rise stratigraphically from north to south and from west to east. Some wells may penetrate as many as five water-producing zones. Secondary porosity is developed in vuggy and cavernous limestone, solution-collapse breccia, and solution-enlarged fractures. Recharge occurs by direct infiltration of precipitation and by runoff from intermittent losing streams flowing eastward across a broad area east of the Sacramento Mountains.

During the initial development of the artesian aquifer, many wells flowed to the surface and high discharge springs fed the Pecos River. Decades of intensive pumping have caused substantial declines in hydraulic head in the aquifer, and by the mid-20th century it was estimated that withdrawals exceeded recharge. Most downgradient flow is intercepted by irrigation wells in the Artesian Basin. Mineral content of the water rapidly increases in an eastward direction. Chloride concentrations range from about 15 mg/kg in the unconfined area to as much as 7,000 mg/kg in an artesian well east of Roswell, N.M. Chloride concentration also increases with depth. The freshwater-saltwater interface migrates westward during periods of low rainfall.



Figure 2.6.1 View from top of the reddish cliffs on the south side of Mirror Lake in Bottomless Lakes State Park. Photo courtesy of the American Southwest. https://www.americansouthwest.net/new_mexico/bottomless-lakes/state-park.html.

The carbonate aquifer provides habitat for several federally listed endangered invertebrate species: the Roswell springsnail (*Pyrgulopsis roswellensis*), Koster's springsnail (*Juturnia kosteri*), and Noel's amphipod (*Gammarus desperatus*) are listed as endangered and the Pecos assiminea (*Assiminea pecos*) is listed as endangered with critical habitat under the Endangered Species Act of 1973, as amended (Act). These four invertebrates occur at sinkholes, springs, and associated spring runs and wetland habitats. They are found at one site in Chaves County, New Mexico,

and *Pecos assiminea* is also found at one site in Pecos County, Texas, and one site in Reeves County, Texas (USGS, 2021).

The largest springs supporting the fragile ecosystems issue at the contacts between the karstified San Andress limestones and the underlying red sandstone, siltstone, and clays of the Yeso Formation. Most of these karst springs have individual discharge rates less than 100 gpm and are typically located on canyon slopes; a few springs issue from collapse breccias of the Yeso Formation. As noted by Davis et al. (1979), nearly vertical joints are an important control upon the location of springs in limestone.

The best-known natural attraction of the Roswell Basin is Bottomless Lakes State Park located 14 mi southeast of Roswell, on the east edge of the Pecos River valley. Two major geologic units are exposed at Bottomless Lakes State Park: Quaternary alluvial deposits, and the cliffs and sinks developed in the Seven Rivers Formation of the Artesia Group, composed of gypsum, dolomite, and fine-grained, red to white sandstone. The lakes are cenotes, small perennial lakes within deep, steep-walled sinkholes that extend below the water table. Groundwater flowing through the underlying San Andres Formation, predominantly from the Sacramento Mountains to the west, slowly dissolves the carbonates forming solution cavities that result in collapse of the overlying rocks of the Artesia Group from their own weight. A spectacular Santa Rosa Blue Hole spring in the nearby Santa Rosa is formed in a similar way (see Chapter 3.6)

2.7 Arbuckle-Simpson Aquifer

The Arbuckle-Simpson Aquifer encompasses portions of Murray, Pontotoc, Johnston, Carter, Coal, Garvin, and Marshall Counties in Central Oklahoma. The area where the aquifer is exposed at the surface (outcrop) encompasses about 520 square miles, but in some areas, where fresh groundwater flows beneath shallower geologic units, the aquifer extends beyond the outcrop. Freshwater from the Arbuckle-Simpson aquifer is produced from wells in an area west of Sulphur and the Chickasaw National Recreation Area (CNRA). The groundwater chemistry of the aquifer changes quickly west of the freshwater zone, and becomes saline (Osborn, 2009). The area where the freshwater zone of the aquifer is known to extend beneath younger geologic units is termed “Arbuckle-Simpson aquifer subcrop” as shown on Fig. 2.7.1. The total area of the Arbuckle-Simpson aquifer defined this way is about 600 square miles.

The Arbuckle-Simpson aquifer lies in the Arbuckle Mountains Uplift geologic province, which is commonly referred to as the Arbuckle Mountains. The Arbuckle Mountains consist of folded and faulted Proterozoic and Cambrian igneous rocks and Paleozoic sedimentary rocks ranging in age from Cambrian through Late Pennsylvanian (Table. 2.7.1). The Arbuckle Mountains consist of a series of northwest-southeast trending structures that are separated from each other by major Paleozoic fault zones. Structural deformation is greatest in the western Arbuckle-Simpson aquifer, where vertical and overturned beds occur. Structural deformation is much less pronounced in the central and eastern Arbuckle-Simpson aquifers, where the rocks are more flat lying (dips are less than 20 degrees), and are deformed mainly by block faulting. The Arbuckle-Simpson aquifer is contained within three major rock units of Upper Cambrian to Middle Ordovician age: the Timbered Hills, Arbuckle, and Simpson Groups. These are described below, from youngest to oldest (Osborn, 2009).

The Cambrian-age Timbered Hills Group is the lowermost geologic unit of the aquifer. The Timbered Hills Group is exposed only in small areas in the Arbuckle-Simpson aquifer and consists of up to 700 feet of limestone, dolomite, and sandstone. Although rocks in the Timbered Hills Group comprise part of the Arbuckle-Timbered

Hills aquifer in western Oklahoma, little is known about the water-bearing properties of the Timbered Hills Group in the Arbuckle Mountains. There is no identifiable confining layer that separates the Timbered Hills Group from the Arbuckle Group, and thus the Timbered Hills is assumed to be part of the same groundwater flow system.

The Arbuckle Group of Late Cambrian to Early Ordovician age comprises the major portion of the aquifer and consists of a thick sequence of carbonate rocks (limestone and dolostone). The Arbuckle Group ranges in thickness from 6,700 feet in the western Arbuckle-Simpson aquifer to less than 4,000 feet in the eastern Arbuckle-Simpson aquifer. Water occurs in cavities, solution channels, fractures, and intercrystalline porosity present in the limestone and dolostone.

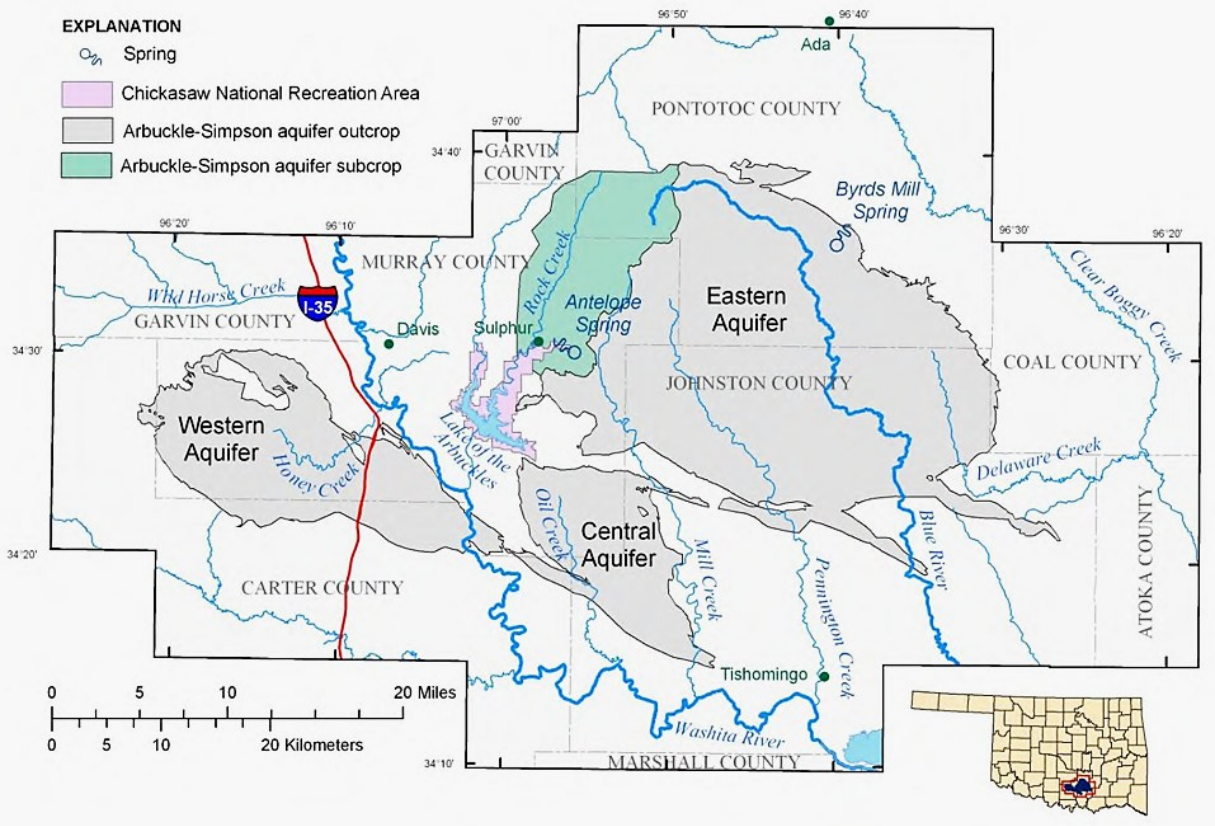


Figure 2.7.1 Extent of the Arbuckle-Simpson Aquifer in Central Oklahoma. Eastern Aquifer corresponds to Hunton Anticline, Central Aquifer to Tishomingo Anticline, and Western Aquifer to Arbuckle Anticline. From Osborn, 2009.

The Simpson Group of Early to Late Ordovician age is the uppermost geologic unit of the aquifer. The Simpson Group is characterized by thick beds of cleanly washed quartzose sandstone interbedded with limestones, dolostones, and shales. The Simpson Group is up to 2,300 feet thick in the western Arbuckle-Simpson aquifer, but generally is less than 1,000 feet in the eastern Arbuckle-Simpson aquifer. Water in the Simpson Group occurs in pore spaces between the sand grains in the sandstones.

The Arbuckle-Simpson aquifer is confined below by basement igneous rocks, consisting of Cambrian rhyolites and Proterozoic granites. In areas where the top of the aquifer dips below the surface, it is confined above by younger rocks of various ages. The time-stratigraphic, rock-stratigraphic, and hydrogeologic characteristics associated with the aquifer are listed in Table 2.7.1.

The Arbuckle-Simpson aquifer receives water primarily from infiltration of precipitation on the outcrop area. Generally, groundwater flows from topographically high areas to low areas, where it discharges to springs and streams. Where the Arbuckle-Simpson aquifer dips beneath rocks of lower permeability, the aquifer is confined, and wells that penetrate below the confining layer may be artesian. Several artesian wells flow in the valley of Rock Creek, near Sulphur. The most well-known artesian well is Vendome Well in the CNRA.

Table 2.7.1 Comparison of time-stratigraphic, rock-stratigraphic, and hydrogeologic units in the Arbuckle-Simpson aquifer. From Osborn, 2009.

Time-Stratigraphic	Rock-Stratigraphic Unit		Hydrogeologic Unit
Pennsylvanian to Late Ordovician	Post-Simpson Geologic Units, Undifferentiated		Upper Confining Unit
Middle Ordovician	Simpson Group	Bromide Formation	Arbuckle-Simpson Aquifer
		Tulip Creek Formation	
		McLish Formation	
		Oil Creek Formation	
		Joins Formation	
Early Ordovician	Arbuckle Group	West Spring Creek Formation	
		Kindblade Formation	
		Cool Creek Formation	
		McKenzie Hill Formation	
Late Cambrian		Signal Mountain Formation Butterly Dolomite	
		Fort Sill Limestone Royer Dolomite	
	Timbered Hills Group	Honey Creek Limestone	
		Reagan Sandstone	
Middle Cambrian	Colbert Rhyolite		Basement Confining Layer
Proterozoic	Tishomingo Granite, Troy Granite, granodiorite, and granitic gneiss		



Figure 2.7.2 Cool Creek limestone of Arbuckle Group exposed at Turner Falls on Honey Creek, in Turner Falls Park near Davis. From Puckette et al., 2009.

The occurrence and movement of groundwater in the Arbuckle Simpson aquifer is strongly controlled by lithology and structure. Associated with the major fault zones are numerous minor faults and joints that occur in the denser beds such as the Arbuckle carbonate rocks. Geologic structure is of significance because fractures caused by folding and faulting provide channels for groundwater movement. Acid water enters the fractures, joints, and bedding planes and enlarges them by solution. The result is an irregular network of openings of all sizes and shapes extending both vertically and laterally. The association of springs with faults and other fractures indicates the significance of structural control on groundwater flow (Fairchild et al., 1982).

Many springs discharge from the Arbuckle-Simpson aquifer. The USGS National Water Information System (NWIS) database lists 140 springs in the study area. Springs are unique ecosystems that provide habitat for a wide diversity of aquatic and terrestrial life. Spring habitats differ from other flowing-water systems in that they have constant flow and temperature, exist as small and isolated habitat areas, and have a general lack of large predators. Arbuckle-Simpson springs provide water for drinking water, livestock, fisheries, and recreation. Some springs, such as those at CNRA, have cultural and historical significance. Byrds Mill Spring, which flows about 8,000 gallons per minute (gpm), is the largest spring in Oklahoma and serves as the primary water supply for the City of Ada (Osborn, 2009).

Larger springs occur in the eastern part of the area, which is underlain primarily by dolomite, than in the western part of the area which is primarily limestone. However, the difference in the amount of discharge from the springs could be a result of the difference in size of the catchment areas rather than a difference in lithology. The catchment area of the eastern part is much larger than the catchment area of the western part of the study area (Fairchild et al., 1982).

The depth to which the network of solution channels can reach is limited by the depth at which groundwater circulation took place. Caliper logs, which show the diameter of a hole with depth, indicate openings at various depths in wells as deep as 2,000 ft (Fairchild et al., 1982).

Puckette et al. (2009) provide the following examples of deep karstification and presence of fresh water in deep exploratory wells: two wells report freshwater in the Arbuckle aquifer on the western end of the Tishomingo Anticline at depths of 1614 feet and 760-780 feet; one well located on the Belton Anticline reports freshwater at a depth of 1460 feet, which is -401 msl; and several petroleum-related wells provide important water data on the Hunton Anticline including presence of fresh water at depths of 1421 ft, 1834 ft, and -321 msl.



Figure 2.7.3 Oxidized or “terra rosa” zones in the West Spring Creek Formation along I-35 in the Arbuckle Mountains. From Puckette et al., 2009; originally from Musselman, 1994.

2.8 Northern Midwest Paleozoic Carbonate Aquifers

Two major groups of Paleozoic karst aquifers in the northern Midwest are parts of the Cambrian-Ordovician, and the Silurian-Devonian aquifer systems. The regional geology of the Michigan Basin is described in detail by Ells (1963). The regional hydrogeology of the Cambrian-Ordovician aquifer system is described in USGS Professional Papers 1305 and 1405–B (Imes, 1985; Young, 1992b), and in USGS Scientific Investigations Report 2011–5229 (Wilson, 2012). The Silurian-Devonian aquifer system is described in the Ground Water Atlas of the United States, Chapter HA-730J (Olcott, 1992). The summary presented here is, in large part, from those sources.

A series of large-scale structural basins and arches control the position and extent of bedrock units in the north-central part of the United States. Among these are the Michigan Basin, Illinois Basin, Appalachian Basin, the Wisconsin Arch, the Kankakee Arch, the Cincinnati Arch, and the Findlay Arch (Fig. 2.8.1). Bedrock in this area ranges from Precambrian to Jurassic in age. Precambrian units consist of crystalline rocks and metamorphosed sedimentary rocks. Paleozoic and Mesozoic rocks are sedimentary. Bedrock units dip away from the crests of the arches and generally thicken into adjacent basins. The surficial aquifer system overlies much of the area covered by Wisconsinan glaciation. The structural basins and arches control the depth to the aquifers. As the depth to the top of the aquifers increases, water quality tends to degrade (as dissolved solids increase), and water use from these aquifers correspondingly declines (Sheets and Simonson, 2006).

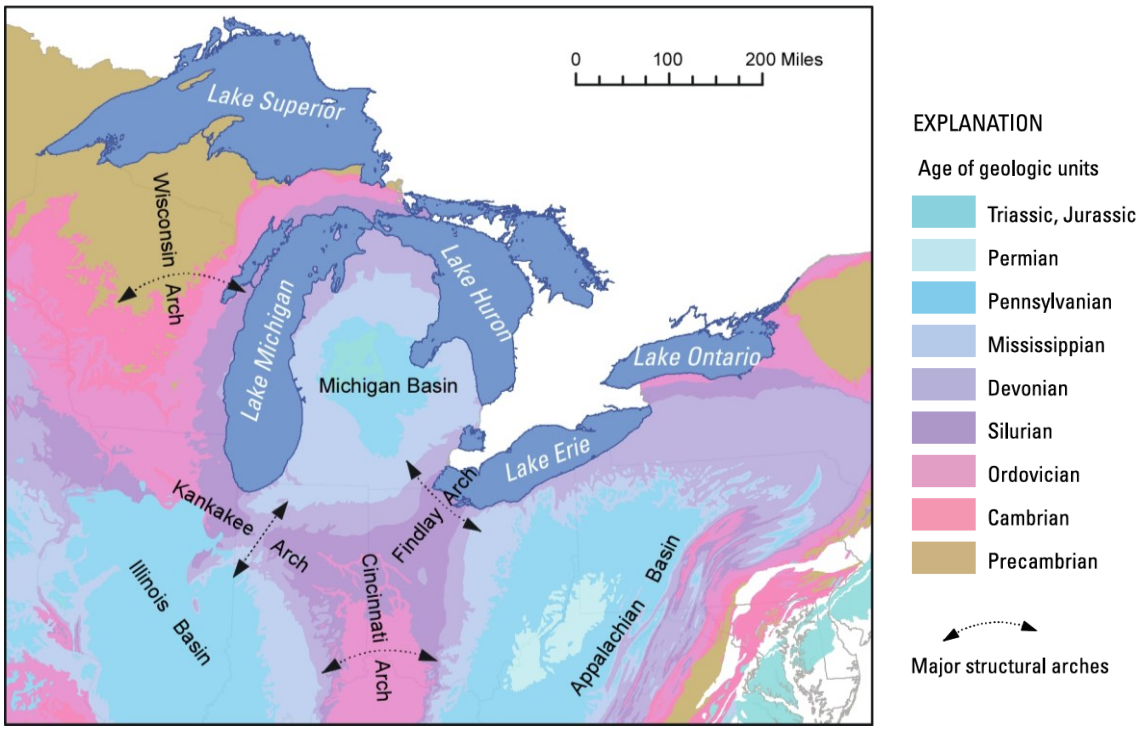


Figure 2.8.1 Bedrock geology of the Great Lakes region in the United States, including major structural features. Modified from Sheets and Simonson, 2006.

Cambrian and Ordovician rocks crop out or subcrop beneath glacial drift or Holocene deposits in northern Illinois, northeastern Iowa, the Upper Peninsula, southeastern Minnesota, and southern Wisconsin. A band of Silurian and Devonian rocks subcrop beneath glacial drift from the Upper Peninsula, through eastern Wisconsin and northeastern Illinois, and on the western side of the Kankakee Arch in northeastern Iowa and western Illinois. Mississippian and younger rocks border and cover the major sedimentary basins (Wilson, 2012).

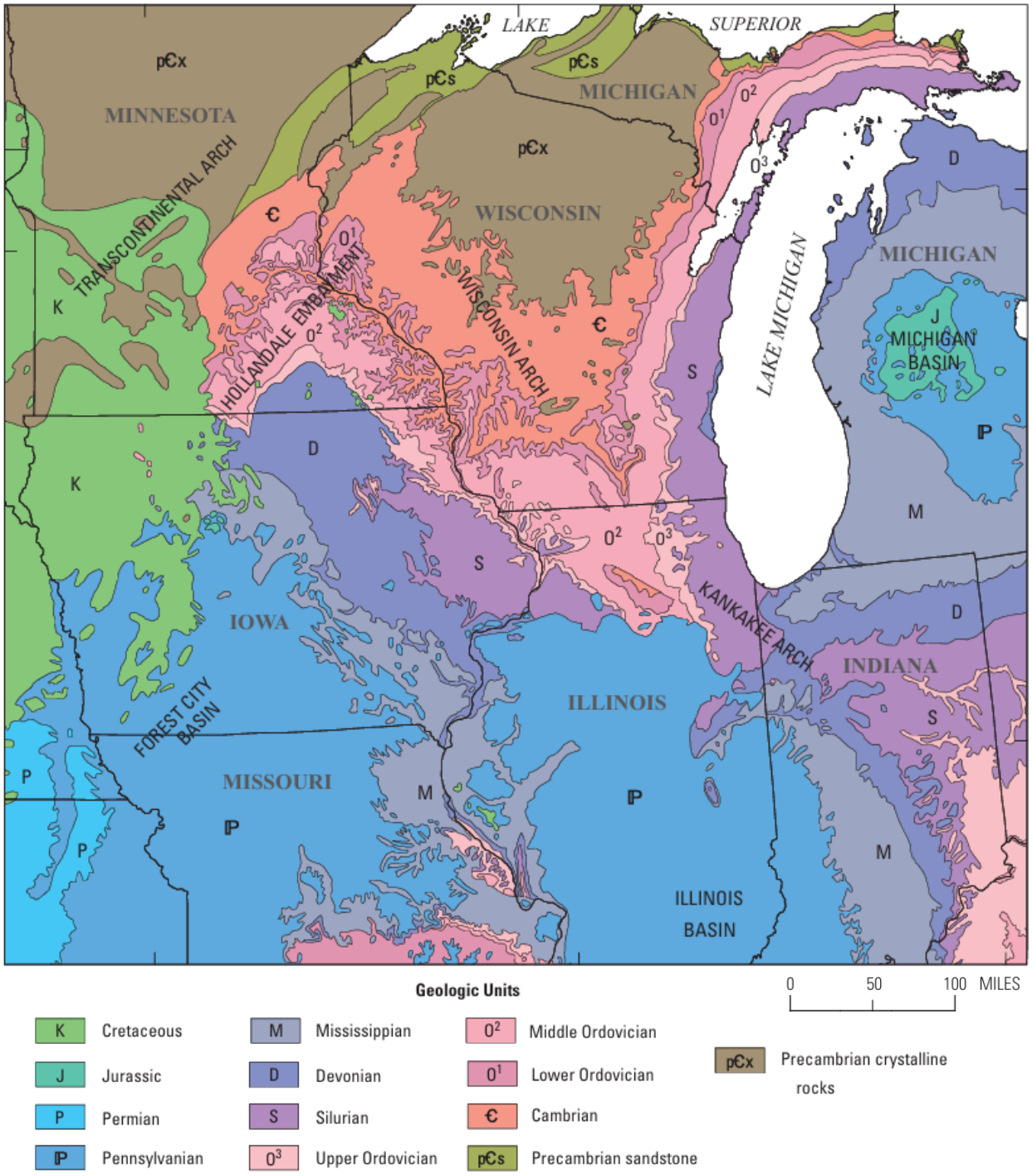
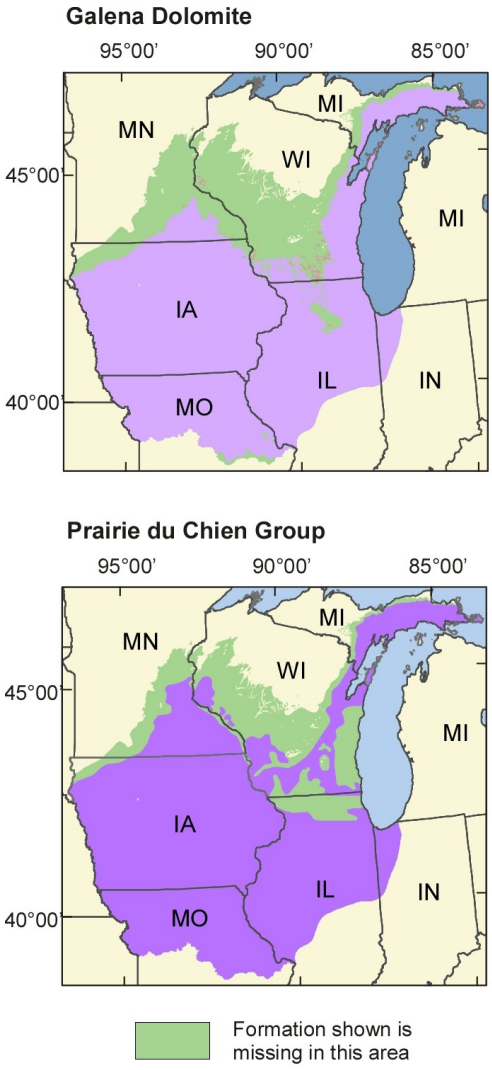


Figure 2.8.2 General bedrock geology of the northern Midwest. From Wilson, 2012.

The Cambrian-Ordovician aquifer system is a complex, multilayered sequence of marine sedimentary rocks. Clastic sedimentary rocks (sandstone and shale) that transition to more carbonate rocks to the south and east in Missouri, Illinois, and Indiana dominate Cambrian strata. Ordovician strata are dominated by carbonates, except for the St. Peter Sandstone. The widespread of soluble carbonates, in particular dolomites of St. Lawrence Formation (Upper Cambrian), Prairie du Chien Group (Lower Ordovician), and dolomites and limestones of Galena Group (Middle to Late Ordovician), has led to numerous caves, sinkholes, and springs in the Upper Midwest.



The Lower Ordovician Prairie du Chien Group underlies the St. Peter Sandstone except in Missouri where the Everton Formation is present. It consists of the Shakopee Formation and the underlying Oneota Dolomite. In extreme northern Illinois and southern Wisconsin, the group was almost totally removed by weathering and erosion. Praire du Chien Group is a significant karst aquifer and landscape in specific areas of the Upper Midwest. It is recognized as one of the primary carbonate units displaying karst development, including sinkholes, caves, and solution-widened joints, particularly where it lies near the surface. Karst features in the Prairie du Chien Group are most pronounced in the so-called "Driftless Area" in southeastern Minnesota, northeastern Iowa, and southwestern Wisconsin, where the unit is exposed or covered by thin glacial drift (see also Chapter 3.7).

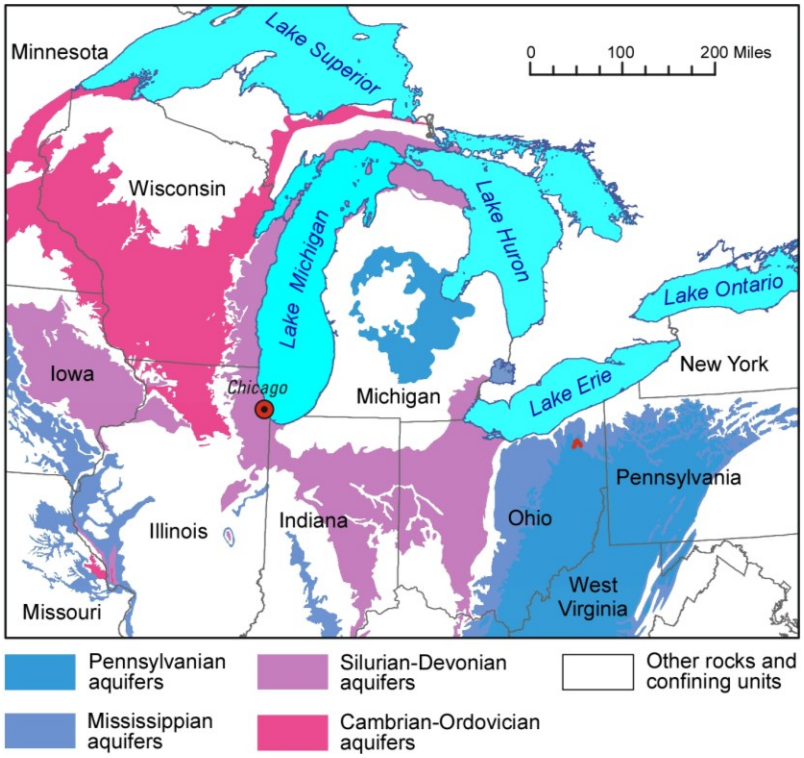
The Middle Ordovician Sinnipee Group underlies the Maquoketa Shale and occurs in some areas where the Maquoketa Shale is missing. The Sinnipee Group consists of the Galena Dolomite, Decorah Shale, and Platteville Formation. The Galena Dolomite is primarily a carbonate, either dolomitic limestone or interbedded limestone and dolomite, but thin shale layers also are present in some areas.

Figure 2.8.3 Approximate extent of two main karst aquifers, Prairie du Chien Group (Lower Ordovician) and Galena Dolomite (Middle Ordovician) within the Cambrian-Ordovician aquifer systems, including where confined and deeply buried. Modified from Wilson, 2012.

The Decorah Shale is mostly dolomitic limestone with variable amounts of thin shale beds. The lower part of the Decorah is mainly shale in Missouri and Illinois. The Platteville Formation is mainly a dolomite or limestone. The formations of the Sinnipee Group are considered Upper Ordovician in Iowa (Iowa Geological Survey, 2004) and Minnesota (Mossler, 2008). The term "Middle Ordovician" is now considered mostly Upper Ordovician in internationally proposed series definitions (Webby, 1998).

The Silurian-Devonian aquifer has been referred to by a number of different names. It is known as the carbonate aquifer in Ohio, the Silurian-Devonian aquifers in Indiana, and the upper part of the shallow dolomite aquifer in Illinois. The aquifer is present in Iowa, Wisconsin, Michigan, Illinois, Indiana, and Ohio (Fig. 2.8.4).

The Silurian-Devonian aquifer consists mostly of limestone and dolomite in all five States but locally contains interbedded shale and evaporite beds. The Silurian-Devonian aquifer generally is overlain by the surficial aquifer system where the aquifer forms the bedrock surface. It crops out along the shoreline of Lake Michigan in eastern Wisconsin, the eastern Upper Peninsula of Michigan, and along the shoreline of Lake Erie in the extreme southeastern part of the Lower Peninsula of Michigan. The aquifer dips and thickens radially into the Michigan Basin; thicknesses range from 1 to 200 feet at the outcrop where the rocks characteristically form an escarpment.



The aquifer is 400 to 500 feet thick at its downdip extent. In Wisconsin, the base of the aquifer slopes eastward from an altitude of about 800 feet above sea level along parts of its western margin to slightly below sea level along parts of the shore of Lake Michigan. In the Upper Peninsula of Michigan, the base of the aquifer is at an altitude of about 800 feet above sea level and slopes southward into the Michigan Basin.

Figure 2.8.4 Principal Paleozoic aquifers in the Great Lakes region at or near the land surface. Modified from Sheets and Simonson, 2006.

In areas where the aquifer forms the bedrock surface and where it is overlain by the surficial aquifer system, especially in northeastern Iowa, the northern part of the Lower Peninsula of Michigan, and the southeastern part of the Upper Peninsula of Michigan, dissolution has resulted in extensive areas of karst features, such as sinkholes and caves. Dissolution of evaporite beds in the Salina Group of Michigan also has caused collapse of overlying Silurian and Devonian rocks and has produced the Mackinac Breccia which has increased the permeability of the aquifer. The Devonian carbonate rocks have extensive karst development in their outcrop and subcrop areas of the northeastern part of the Lower Peninsula of Michigan and in the subcrop areas of the southeastern part of the Lower Peninsula.

Era	System	Stratigraphic unit		Principal lithology	Hydrogeologic unit
Paleozoic	Devonian	Ellsworth Shale		Shale	Confining unit
		Antrim Shale			
		Traverse Group		Limestone and basal shale	Silurian-Devonian aquifer
		Roger City Limestone		Limestone	
		Dundee Limestone			
		Detroit River Group	Mackinac Breccia	Dolomite	
		Sylvania Sandstone		Sandstone	
		Bois Blank Formation		Dolomite	
		Garden Island Formation			
	Silurian	Bass Island Group			
		Salina Group		Limestone, shale, and evaporite beds	
		Niagara Group		Limestone and dolomite	
		Cataract Group		Shale and dolomite	Confining unit
	Ordovician	Richmond Group		Shale and limestone	

Figure 2.8.5 The Silurian-Devonian aquifer in Michigan consists primarily of dolomite and limestone with interbedded sandstone, shale, and evaporite beds. The Mackinac Breccia results from collapse of Devonian rocks after dissolution of some of the underlying Silurian evaporite beds. From Olcott, 1992. Modified from Western Michigan University, 1981.

In areas where the aquifer is overlain by thick confining units, such as Mississippian and Devonian shales which impede recharge to the aquifer, neither the formation of karst features nor the dissolution along joints, fractures, and bedding planes have occurred. In these areas, the aquifer tends to have minimal permeability (Olcott, 1992).

The Niagara Escarpment

The following description of the Niagara Escarpment, one of the most spectacular karst features on a continental scale in North America, is based primarily on the material presented in an excellent paper by John A. Luczaj entitled *Geology of the Niagara Escarpment in Wisconsin* (Luczaj, 2013).

The Niagara Escarpment is the topographic expression of a bedrock unit that extends at least 650 mi from eastern Wisconsin through the Upper Peninsula of Michigan, through Manitoulin Island and the Bruce Peninsula into southern Ontario and into the Niagara area of New York. This internationally recognized geologic feature was named for the region around Niagara Falls in New York and Ontario where the Niagara River plunges over the Silurian Lockport Dolostone into a gorge cut through the underlying lower Silurian and uppermost Ordovician mudstones and sandstones. Notably, the dolomites (dolostones) that form the Niagara Escarpment have different names in different states and are not necessarily all at the exact same stratigraphic horizon because of the complicated and extensive reef and carbonate bank-type deposits associated with the Michigan Basin margin. However, they are all formed by a differential erosion of the resistant Silurian dolomites that overly much softer Silurian and Ordovician shales.



Figure 2.8.6 In New York, the Niagara Escarpment originates near Watertown, passes through Niagara County, and terminates east of Rochester; in Michigan, the ridge forms the Garden Peninsula and stretches across the Northern Peninsula along the northern shore of Lake Michigan; in Wisconsin**, t**he escarpment is a dominant feature, spanning from the Door Peninsula down through the eastern side of the state, known as "The Ledge" in Dodge County. The Escarpment portion occurring in Ontario, Canada has been designated as a UNESCO World Biosphere Reserve. Courtesy of https://www.wiledge.org/learning.

The Niagara Escarpment is the most prominent topographic and geomorphic feature in eastern Wisconsin. It includes outcrops that form a dolostone ridge or series of ridges at the Silurian bedrock surface along the western edge of the ancestral Michigan basin. The mainly west and northwest-facing escarpments have developed on resistant eastward-dipping Silurian dolostones that overlie the much softer Ordovician Maquoketa Shale. The exposed, but discontinuous escarpment runs from Rock Island on the tip of Door County all the way south to Ashippun in Dodge County and ranges in height from a few feet to over 200 ft.

The Mayville Formation Dolostone is the principal cliff-forming unit over much of the Niagara Escarpment in Wisconsin. The thickness of the Mayville formation varies from 66 ft in Fond du Lac County to as much as 230 to 270 ft in Door County. The Burnt Bluff Group, which consists of Byron Dolostone and the overlying Hendricks Dolostone, is also a prominent cliff-forming unit, especially along the western shore of the Door Peninsula north

of Little Sturgeon Bay where extensive cliffs are 100 to 200 ft high. The Burnt Bluff Group is often heavily fractured and exhibits some of the best-developed karst features in northeastern Wisconsin (Luczaj, 2013).

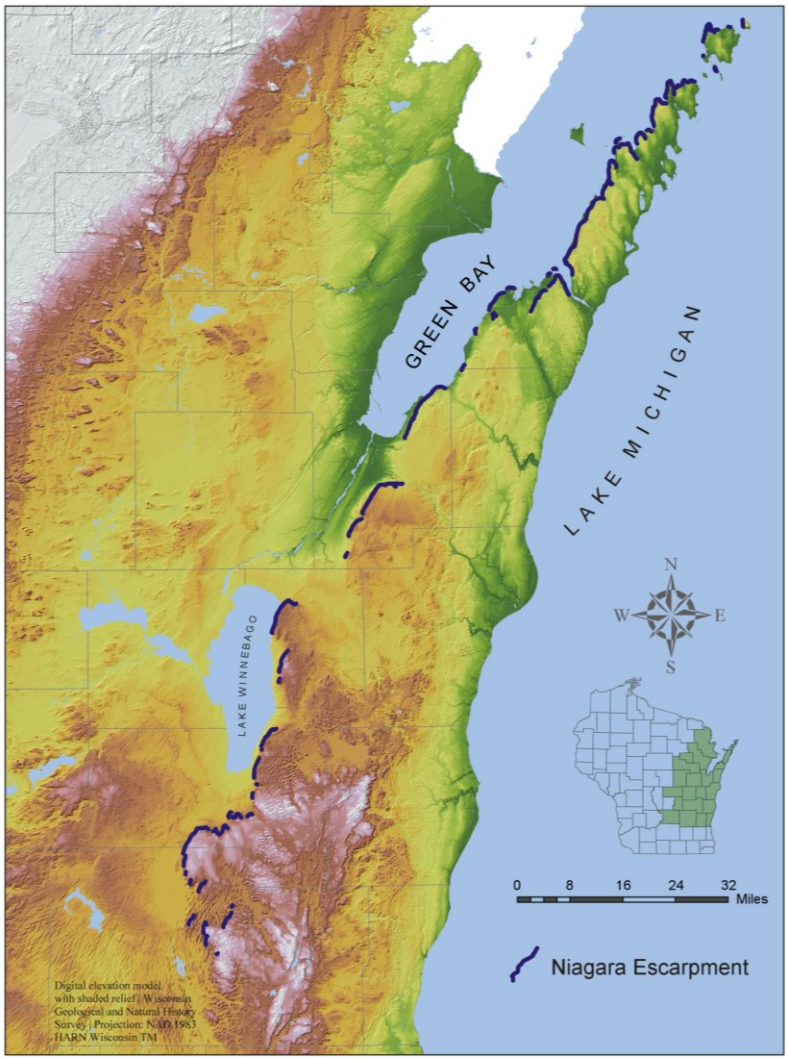


Figure 2.8.7 Digital elevation model with hillshade showing the location of the Niagara Escarpment in Wisconsin. Location data sources: Niagara Escarpment Resource Network (NERN), outcrops, well construction reports, USDA soil maps, and other sources. From Luczaj, 2013.



Figure 2.8.8 Cliff face of the Wisconsin Niagara Escarpment. Photo by Eric Fowle, Courtesy of Wisconsin Public Radio Exploring The Wisconsin Niagara Escarpment - WPR

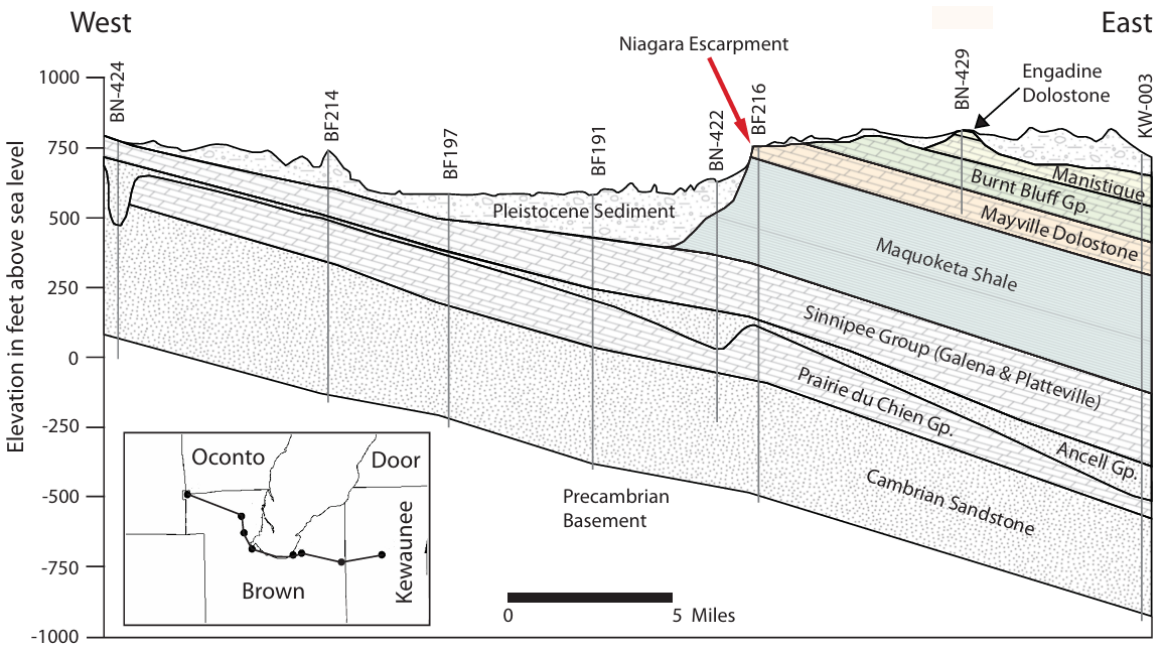


Figure 2.8.9 Cross section through the midpoint of the Niagara Escarpment in Wisconsin. Rock strata dip eastward toward the ancestral Michigan basin. Note the relief on the bedrock surface west of the escarpment that is partially concealed by Pleistocene glacial sediments. Vertical exaggeration is 45 x. Wells (vertical lines) are BN-424 = McKeefry Borehole, Pulaski; BF214 = Village of Howard Well 2; BF197 = City of Green Bay Well 10; BF191 = City of Green Bay Well 4; BN-422 = Shorewood Golf Course Well at UW-Green Bay; BF216 = Town of Scott Well 1; BN-429 = Green Bay Water Treatment Plant Core; KW-003 = former Green Bay and Western Railroad Well, Luxemburg. From Luczaj, 2013; Modified after Maas, 2009.

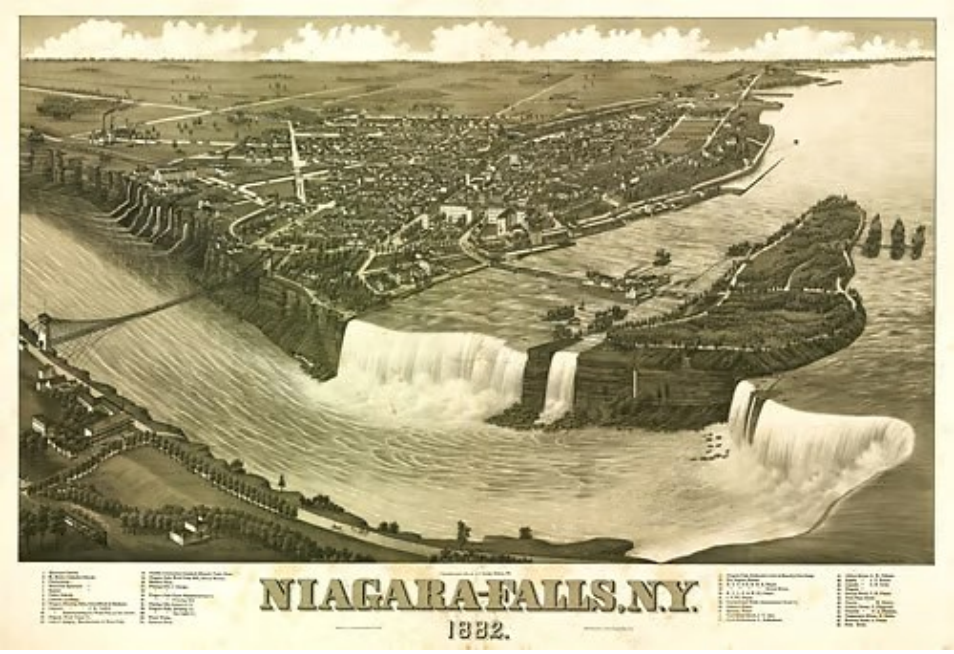


Figure 2.8.10 Niagara Falls State Park is the oldest state park in the United States, established in 1885 as the Niagara Reservation. Geologically speaking, Niagara Falls is quite young: 10,000 to 12,500 years old. It formed at the end of the last Ice Age as melting glaciers released massive amounts of water, causing the Niagara River to carve the Niagara Gorge through the Niagara Escarpment. The falls have retreated about 7 miles from their original location. Since the escarpment is most famous as the cliff over which the Niagara River drops off at Niagara Falls, it was named the Niagara Escarpment in other states as well. Photo of this historic poster courtesy of Niagara Falls State Park.



Figure 2.8.11 Historic photograph of Niagara Falls power plants and mills lining the cliff formed by the erosion resistant Middle Silurian Lockport Dolomite. The potential power of Niagara Falls attracted industrialists who worked to harness its force using water wheels to drive their mills and factories. In 1882, the Niagara Falls Hydraulic Power & Manufacturing Company used a small, 1,800-horsepower direct current (DC) plant to provide electricity for streetlights in the town of Niagara Falls, New York. The world's first large-scale hydroelectric generating station opened in Niagara Falls in 1895. The famous electrical engineer and inventor Nikola Tesla designed the first alternating current (AC) hydroelectric power plant at Niagara Falls. His successful application of AC technology at the falls proved it could generate and transmit electricity over long distances, cementing its victory over Thomas Edison's DC and enabling the electrification of the globe. The Edward Dean Adams Power Plant began operations on August 25, 1895, and on November 16, 1896, power was successfully transmitted to Buffalo, New York, 22 miles away. To ensure the project's success during financial strain on Westinghouse, Tesla famously tore up his royalty contract, giving up potential millions to keep his dream alive. The Niagara Parks Power Station offers a "Tesla: A Spark of Genius" exhibit, highlighting the nine patents of his used at the plant. Photo courtesy of Niagara Falls State Park.

2.8.1 Upper Carbonate Aquifer (Paleozoic)

The Upper Carbonate Aquifer is an important aquifer only in southeastern Minnesota and for a short distance into northeastern Iowa. It is part of a sequence of sedimentary bedrock units deposited in Paleozoic seas that occupied a depression known as the Hollandale embayment. The aquifer is comprised of four formations, which, in ascending order, are the Ordovician Galena Dolomite, Dubuque Formation, Ordovician Maquoketa Shale, and the Devonian Cedar Valley Limestone. The aquifer is underlain by shale, dolomitic limestone, and limestone of the Decorah Shale, the Platteville Formation, and the Glenwood Shale that form an effective confining unit. Fracturing and subsequent dissolution of carbonate facies of the rock has produced the very productive karst aquifer.

The Upper Carbonate Aquifer ranges in thickness from a featheredge along its periphery to about 650 feet in a small area in the center of the embayment along the Minnesota-Iowa border. The rocks are extensively fractured and jointed, and numerous solution-enlarged rock openings, including sinkholes, solution cavities, and caves, have made the rocks extremely porous, especially in areas where glacial deposits are thin or missing. The aquifer extends a short distance into Iowa. In Iowa, however, dissolution has been inhibited and, therefore, little secondary permeability has developed in the rocks. This is because (1) the rocks are deeply buried, (2) there is a facies change

from dolomite to shale in the Maquoketa Shale, and (3) the upper part of the Maquoketa contains shale units in Iowa that are not present in Minnesota. These rocks are not considered to be an aquifer in Iowa, and the southern limit of the aquifer has not been defined in the literature.



Regional groundwater flow in the upper carbonate aquifer generally is outward toward the periphery of the aquifer. The aquifer is recharged through the overlying surficial aquifer system that also acts as a leaky confining unit where it contains large quantities of clay and silt. Water movement is along short flow paths toward the rivers that drain the area eastward to the Mississippi River, northwestward toward the Minnesota River, and southward into streams flowing into Iowa.

Figure 2.8.12 The Upper Carbonate Aquifer is present in southeastern Minnesota and extends a short distance into Iowa. From Olcott, 1992.

The karstic nature of the upper carbonate aquifer and the thinness or absence of overlying glacial deposits along the eastern margin of the aquifer make this part of the aquifer susceptible to contamination from the land surface. The potential for contamination from the land surface in the central and the western parts of the aquifer is much less where thick till is present (Olcott, 1992; https://pubs.usgs.gov/publication/ha730.)

2.9 Marble Karst in California

Marble areas in California are limited to isolated portions of the northern Klamath Mountains, and the Sierra Nevada foothills, with prominent sites including the Marble Mountain Wilderness (Siskiyou County), and Crystal Cave in Sequoia National Park. While not considered prolific aquifers or utilized for water supply by large public utilities, they nevertheless feature nice examples of well-developed karst including impressive caves.

The Marble Mountain Wilderness in Klamath Mountains is one of California's oldest formally designated wilderness areas. Originally established in 1931 as the Marble Mountain Primitive Area, it later became one of the original wilderness areas established by the Wilderness Act of 1964. The 225,000+ acre area, situated within the Klamath National Forest and Six Rivers National Forest in Siskiyou County, southwest of Yreka, is managed by the Forest Service.

The Wilderness contains rugged mountainous terrain, diverse vegetation and picturesque high elevation lakes and streams. It was named for Marble Mountain, a monolith of white marble standing 6,880 feet above sea level. Boulder Peak at 8,229 feet is the highest point in the wilderness and the mouth of Wooley Creek at the wilderness boundary is the lowest point at 640 feet.



Figure 2.9.1 Marble Gap in Marble Mountain. Photograph by Curt Mekemson, from his blog *Wandering through Time and Place* at https://wandering-through-time-and-place.com/2025/07/14/the-marble-mountain-wilderness-magnificent-views-up-close-and-far-away-part-2/

The marble rocks in Sequoia & Kings Canyon National Parks contain more than 270 marble caves, including the longest cave in California, Lilburn Cave, with nearly 17 miles of surveyed passages. Lilburn is a very complex maze cave with beautiful blue- and white-banded marble. Nearby mines attest to the unusual geology in the Lilburn area, and the cave has displays of rare and colorful minerals including green malachite and blue azurite. Beautiful Crystal Cave features a trail and lights for park visitors. This commercialized cave has seen millions of visitors since it first opened to the public in 1941. It has beautifully banded marble, many cave formations, large rooms, and the creative Spider Web Gate. Soldier's Cave has been a favorite with California cave explorers since its discovery in 1949.



Figure 2.9.2 Stills from an exceptional video of Crystal Cave in Sequoia National Park, produced by Steven M. Bumgardner for National Park Service, featuring Rebecca Jones of Sequoia Parks Conservancy and geologist Greg Stock of National Park Service. The video is available at https://www.nps.gov/seki/planyourvisit/crystal-cave.htm

The most recognized marble karst in California is in the Mother Lode region, along the western edge of Sierra Nevada. While the area is famous for gold, it also contains significant deposits of high-quality marble that have been mined for over a century. The marble is part of the Calaveras Complex of metamorphic rocks, with particularly well exposed marble areas in the Tuolumne, El Dorado, and Calaveras Counties. The Town of Columbia in Tuolumne County, known as the "Gem of the Southern Mines", was home to several major marble quarries that supplied marble for numerous historic buildings in San Francisco and throughout the region. Other notable marble mines and quarries include Indian Diggins in El Dorado County, and those in Vallecito, Murphys, and Cave City in Calaveras County.

The following excerpts describing the importance of marble for the Calaveras County are from an excellent publication by Fuller (1979):



Figure 2.9.3 Outcrops of the Calaveras Formation showing solution features. Photograph by Bruce Rogers (USGS, retired). Courtesy of Mercer Caverns, https://mercercaverns.net/geology.html

"*The coral islands of the late-Paleozoic seas, although a minor feature of that landscape, were later on to have great significance. These islands, buried by succeeding sedimentation, were metamorphosed during late-Jurassic time into beds of marble. Subsequently, some of these were exposed by the deep pre-Tertiary erosion so that today we see them at the surface.*

Probably commencing early in the Tertiary Period, ground water dissolved out a number of caves in the beds of marble. The process forming these caves, below the ground-water level, is referred to as the "phreatic" process. Carbonic acid derived from the rainwater and other acids from decaying vegetation, reinforced by acids of mineral origin, built up ground water locally acid enough to slowly but surely dissolve out great caverns in the marble: These were formed just a short distance below the early Tertiary land surface, although some of the "roots" of these caves extended as much as several hundred feet below the old surface as almost discontinuous and very sinuous narrow openings.

The end of the Tertiary and into Pleistocene time, as the great canyons such as the Stanislaus and Mokelumne rivers flow down today were cut, the water table was lowered, and many of these caves were drained. Those that were situated near the canyon slopes were enlarged by the "vadose" process, that is, by the action of water percolating downward from the surface to the water table. For the most part, however, the old openings were partly filled in by beautiful "speleothems" such as stalactites and stalagmites. Others were destroyed during late-Tertiary when erosion by Mehrten stream channels cut into them, and filled them with debris. Again, in more recent time, erosion has destroyed more caves in the same fashion.

The best development of marble caves is along the slopes adjacent to the Stanislaus Canyon and certain of its tributary ravines, in. the marble terrain from Cataract Gulch downstream to the Abbey's Ferry crossing. Another cave district is found where this same marble belt crosses the Indian Creek and McKinney Creek drainages. The best-known caves include those at Cave City, Mercer's (north of Murphys), and Moaning Cave near Vallecito on the Parrot's Ferry road, all of which are open to the public. Others, particularly along the Stanislaus River, are smaller, less accessible, and on private land. A remarkable cave, almost completely filled with mud and other debris, was mined out by the Calaveras Cement Company at their Calaveritas quarry. Here, part of the roof of the cave was eroded out in Holocene or late-Tertiary time, and the remaining roof and upper walls collapsed into the cavern, nearly filling it. A similar but much larger cave, near Wool Hollow, on the north side of the Stanislaus Canyon, was cut into by a Mehrten channel, and filled tightly with Mehrten sand. The modern Flintkote limestone slurry plant was built on top of this sand-filled cave, and the quarry has excavated to the edge of the old cave. Thus, the little coral islands of the Paleozoic Sea have supplied us with the raw materials of limestone and marble for the limestone caves for which Calaveras is noted, entered definitely into Indian life, and culture, although the manner in which they did constitutes a controversy among anthropologists. Did they use the caves as inhabitation or refuge, or as places of burial? Or did victorious tribes condemn their victims to a terrible fate in these dangerous subterranean openings? Or were they carefully avoided, and merely existed as traps for the unwary and unsuspecting aborigines? Many skulls and other skeletal remains have been found in these caves, often encased in speleothem coatings suggesting a great antiquity. This feature led to a spectacular "hoax" and controversy in the 1860's when one of these skulls was planted in Valley Springs sediments in a mine shaft near Altaville. This was "discovered" and eventually delivered to Harvard scientist J. D. Whitney who spent the rest of his lifetime trying to prove that modern man existed back into the dim Miocene ages."



Figure 2.9.4 Moaning Caverns holds California's deepest cave chamber, where one can descend deep underground on a guided walk tour. Photo by Dave Bunnell. https://www.gocalaveras.com/business/caves/moaning-cavern-adventure-park/

Calaveras County is home to several spectacular marble cave systems including California Cavern, the oldest show cave in California which opened in 1854, Moaning Caverns, and Mercer Caverns. California Cavern in Mountain Ranch (https://cavetouring.com) is a State Historic Landmark and was once visited by John Muir and Mark Twain; it offers walking tours of extensive marble passageways, as well as "wild" cave tours that go into undeveloped sections.

Moaning Caverns in Valecito (https://moaningcaverns.com/) has the largest single cave chamber in California, “tall enough to hold the Statue of Liberty.” Mercer Caverns in Murphys (https://mercercaverns.net/) is a National Natural Landmark known for rare "flos feri" aragonite formations. Equally spectacular is Black Chasm Cavern, a National Natural Landmark in Volcano, Amador County (https://cavetouring.com/).

Calaveras County secures its potable drinking water through both groundwater and surface water sources. While the majority of the large public water systems within the County use surface water as a source of supply, all small public water systems use groundwater, primarily from wells drilled into bedrock aquifers including limestone and dolomite dominated lenses within the Calaveras Formation near Mountain Ranch.



Figure 2.9.5 Speleothems in Mercer Caverns. *In a final frosting of the Caverns, solutions rich in calcium carbonate deposited walls of delicate aragonite crystals in the far recesses of the cave (right photo). As the minerals concentrated in what little solution remained, uncommon, blue-white, cream cheese-like hydromagnesite moonmilk formed on cave passage walls and on the spiky aragonite crystals. Resembling miniature ice-sheathed trees, this showy display of aragonite is so outstanding that a special prize was awarded to the Mercers at the 1900 Paris World's Fair*. https://mercercaverns.net/geology.html
Photography by Dave Bunnell, https://mercercaverns.net/photos.html.

Less known karst features in Calaveras County are the Natural Bridges that span Coyote Creek between Columbia State Historical Park and Vallecitos on Parrot's Ferry Road. According to Garry Hayes, teacher of geology at Modesto Junior College (https://geotripper.blogspot.com/), at Coyote Creek, springs of carbonate-laden water emerged from the hillside above the river, and calcium carbonate, the mineral that makes up marble, was precipitated into large masses of flowstone. They blocked the creek, but eventually water worked through the base of the spring deposits. Eventually the creek carved a course under this natural dam. The cavern has since been decorated with stalactites, stalagmites, columns, curtains, and other speleothems. As can be seen when visiting the area of the bridges, the springs are still active.

There are two bridges on Coyote Creek. Most hikers visit the upper cave, which is at the end of a relatively gentle one-mile trail. The bridges are on public lands administered by the Bureau of Reclamation (New Melones Reservoir lies just downstream), The link for directions and maps to the caves is at http://www.usbr.gov/mp/ccao/newmelones/planning_visit.html.

As pointed out by Siera Nystrom, a California naturalist and avid birder, the idea that these caverns were formed by mineral deposits from natural springs has apparently been around for a while (https://natural-history-journal.blogspot.com/2016/09/geology-of-natural-bridges-calaveras.html). In *Scenes of Wonder and Curiosity in California* published in 1862, James M. Hutchings, an early promoter of Yosemite who came to California during the Gold Rush, provided this account of Natural Bridges:

"The entire rock formation of the vicinity is limestone, and various are the conjectures relative to the first formation of these natural bridges or tunnels. Some believing them to have been formed by the rocky deposit contained in, and precipitated by, the water of countless springs, issuing from the banks of the creek, that, gradually accumulating and projecting, at length united the two sides, forming these great arched passages."

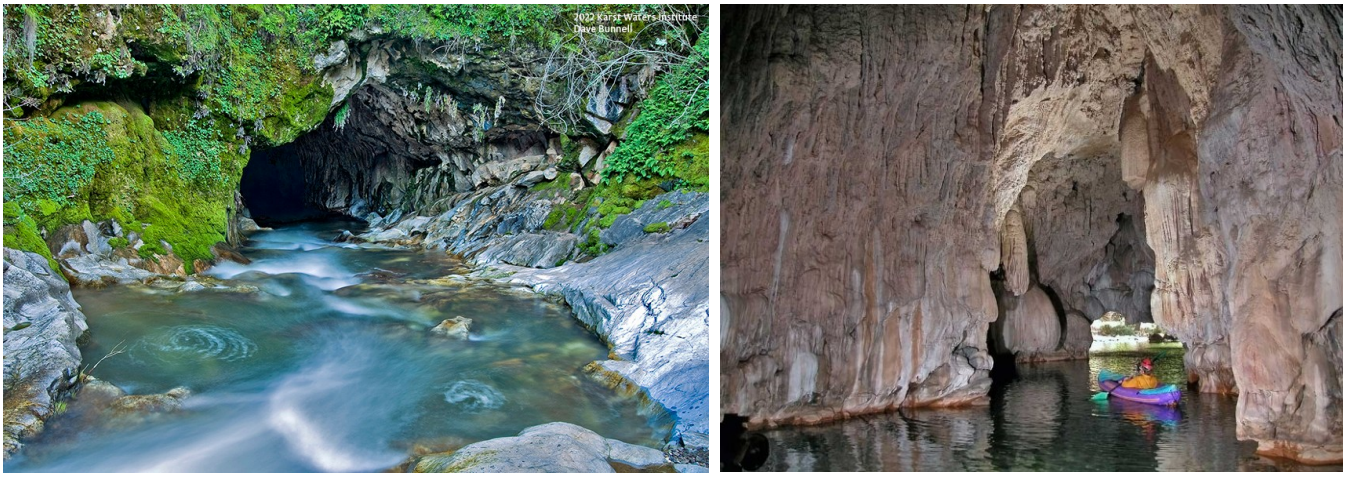


Figure 2.9.6 Left photo of the Calaveras Natural Bridge by Dave Bunnell was the 2022 1st Place Winner in Karst Waters Institute's photo contest. Right photo, also taken by Dave Bunnel, is from inside the cave through which Coyote Creek flows under the bridge (https://www.gocalaveras.com/business/outdoor-recreation/natural-bridges/).

2.10 Pseudokarst Aquifers

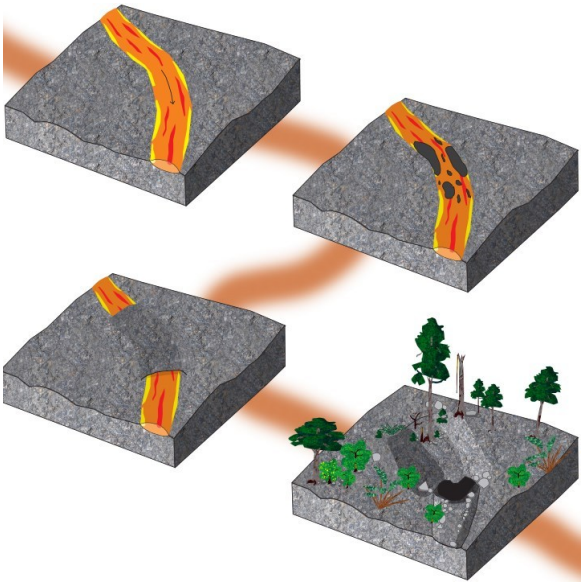
Except for the four springs in Montana which drain the karstic Madison aquifer (see Chapter 3.3), all other first-magnitude springs in the West issue from the vast Pacific Northwest basaltic and volcanic rock aquifers in Oregon, the lava fields of northeastern California, and the most prolific of them all, the basaltic Snake River Plain aquifer in south central Idaho. What these aquifers have in common is their pseudokarstic nature. Pseudokarst can occur in non-soluble rocks including metamorphic and volcanic and has features similar to those developed in classic karst – hummocky topography, sinking streams, sinkhole-like closed depressions, preferential flow paths, open fissures, and even caves. For example, ninety-nine (99) percent of Craters of The Moon National Monument & Preserve (CMNMP) in Idaho was identified as containing pseudokarst (KellerLynn, 2018). These landforms are created through mechanical processes rather than the chemical dissolution typical of carbonate bedrock.

While limestone caves with stalactites and stalagmites might be what most people think of when they think of caves and karst, there are several other types of caves. The second most common are lava tubes which abound at

CMNMP and contribute significantly to the groundwater flow characteristics in the underlying basaltic aquifer. It is virtually certain that paleo lava tubes, i.e., lava tubes formed during major previous phases of volcanic activity in the Snake River Plain, are present throughout the >5,000 thick Quaternary basalt flows of the Snake River Group and not only within the limits of the CMNMP. Although paleo lava tubes must have been altered multiple times due to the pressure of overlying rocks and magmatic/tectonic activity, they still play a role in groundwater movement at greater aquifer depths.

The following description of lava tubes formation and presence in CMNMP, other geologic features related to pseudokarst, and photographs in Figs 2.10.1 through 2.10.4 are courtesy of National Park Service (NPS) and U.S. Forest Service. This and other information on pseudokarst, illustrative diagrams, and color photographs are available in an excellent NPS publication entitled "*Craters of the Moon National Monument and Preserve. Geologic Resources Inventory Report*" by Katie Keller Lynn (2018).

Lava is thousands of degrees hot and can flow like a river. The top of this white-hot river will cool with exposure to the much colder temperatures in the air above. As it cools the lava solidifies into black stone while down below the molten river keeps flowing. The rock above keeps cooling. It thickens and widens and will finally form a roof across the molten rock below creating the ceiling of the cave. Eventually as the eruption ends or the lava flow moves to someplace else, the lava tube drains of molten rock leaving a cave tunnel behind.



Most often lava tubes are close to surface. This makes collapses and entrances more common than in other types of caves. With more entrances there can be a closer connection between these caves and the surface. While most lava caves are shallow, some are buried by newer lava flows. Occasionally new lava may fill an existing older tube destroying it.

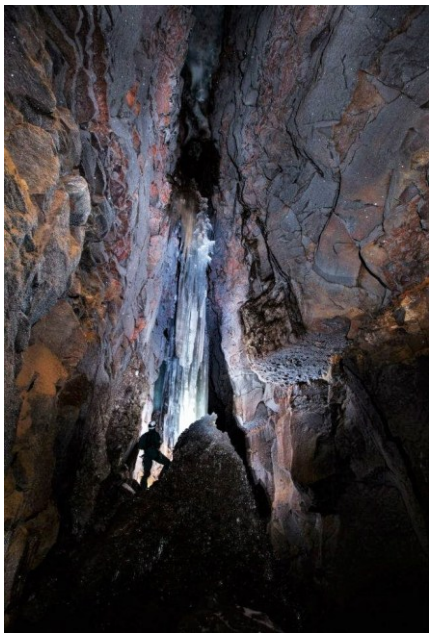
Figure 2.10.1 When fluid, molten lava flows out of the ground, it works its way downhill. Soon the surface of this lava stream cools and hardens into a crust. Although the outer crust is hard, the lava inside is still molten, and continues to flow downhill. Once the molten lava has passed through, it leaves an empty tunnel called a lava cave, or more commonly, a lava tube. NPS image by Joel Despain.



Figure 2.10.2 Example of a lava tube in Craters of The Moon National Monument & Preserve, Idaho: Buffalo Cave, housed in the Broken Top flow, which has multiple levels and six entrances. The height of the opening in the upper level shown in the photograph is about 4 ft. NPS photographs from Niles et al., 2011.

Many lava caves are single tubes that can extend for miles. But occasionally lava tubes can be quite complex with passage splits and junctions, multiple levels, and even sudden drop offs, all of which reflect how the lava flowed and formed. The number of documented caves within the park currently exceeds 700 and more are found each year.

Lava caves are born of fire, but at high elevations or in cold regions they can contain ice (Fig. 2.10.3). Caves are very well insulated with constant temperatures reflecting the average temperature of the region where they form. Summer high temperatures and winter lows have almost no effect underground. Lava tubes in cold places eventually become cold once the volcanic eruption ends. Often this means cold enough for ice.



In some instances, the shape of cave passages may make a trap for cold air that sinks into the tube in the winter and remains cold through the summer. These caves lack stalactites and stalagmites, but they can have large, spectacular ice formations. Their wet surfaces and transparent nature make for a delightful light show in the ice caves. Ice caves are well represented in CMNMP.

Like limestone caves, lava caves have lots of interesting features inside. Some have stalactite-like “lavacicles” hanging from the ceiling where molten rock had dripped. The drips can also hit the floor making stalagmite-like features that rise like stacks of tiny pancakes.

Figure 2.10.3 Ice Formations in King's Bowl. Some caves in Craters of the Moon hold ice year-round like this fissure cave at King's Bowl. This cave is located inside the Great Rift and has been measured to be more than 600 feet deep. Photo courtesy of NPS.

Most caves in the park are lava tubes but other types of caves are also present including caves developed along fissures (fissure caves), and caves formed by differential weathering of lava flows.

Fissure caves are found within the deep cracks that make up the Great Rift. Some of these caves are remarkably deep, including one particular fissure that may be passable to a depth of 600 feet from the surface. King's Bowl is an excellent example of one of these caves, but entry is restricted due to extreme safety hazards in and around the caves. All caves at Craters of the Moon, except for Indian Tunnel, the most famous lava tube, are closed for safety reasons and to protect sensitive bat habitat.



Figure 2.10.4 Aerial view of King's Bowl, CMNMP. Photo courtesy of NPS.



Figure 2.10.5 Lava Tube in Lava River Cave, Deschutes National Forest, 12 miles south of Bend, Oregon. It is Oregon's longest continuous uncollapsed lava tube, with a total surveyed length of approximately 6,711 feet (about 1.27 miles). Lava River Cave was one of the first lava tubes to be discovered by settlers in Central Oregon. It was first known as Dillman Cave, named after its discoverer, Leander Dillman, who found it while hunting in 1889. Dillman was a stockman and trapper who lived nearby and used the cave's 40°F year-round temperature as a natural refrigerator to cool his venison. For 32 years the cave was known as Dillman's Cave, but in 1921 the name was changed to Lava River Cave. Photo courtesy U.S. Forest Service – Pacific Region. Taken on July 19, 2022.

"Sinkholes" in lava generally lack the symmetry of those developed in solution terrain but there are nice examples of perfect “karst-like” sinkholes (Fig. 2.10.6). The lava sinks are commonly less than 100 ft (30 m) wide, but a few large sinks are as much as 1 mile or wider. Most of the lava sinks are irregular in shape and generally are shallow features (less than 30 ft deep), although some are 150 ft or deeper. Many of the sinks have near-vertical sides or overhangs.

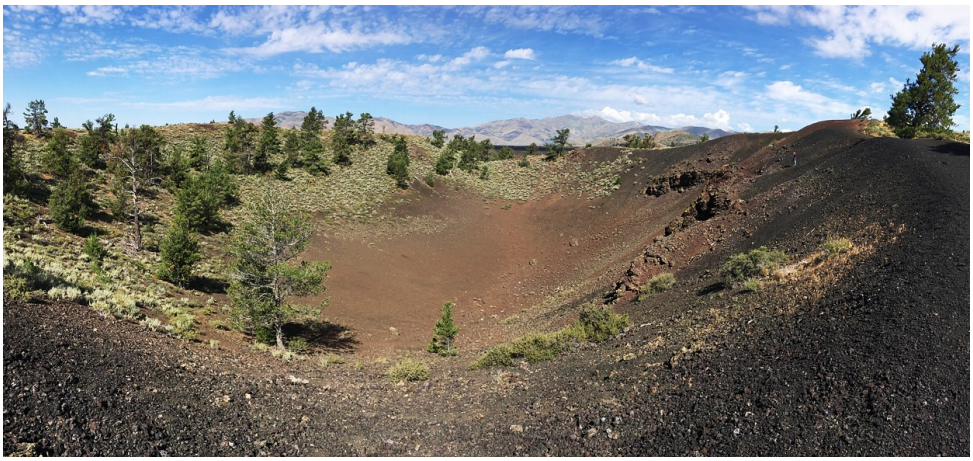


Figure 2.10.6 One of many large, closed depressions (“sinkholes”) in the basalts of Craters of The Moon National Monument & Preserve, Snake River Plain, Idaho.

The most important for groundwater flow characteristics in the aquifer and the emergence of large springs are preferential flowpaths in the basalt where transmissivity and groundwater velocity can be extraordinarily high, like in true karst, as illustrated in Fig. 2.10.7.

Most large springs in the Eastern Snake Plain Aquifer (ESPA) issue from vesicular and brecciated tops of lava flows and pillow lavas which represent the most obvious preferential flow paths in the aquifer (Fig. 2.10.8). However, because of multiple episodes of Quaternary lava flows of various magnitudes and sizes, and in many different directions, these preferential flow zones are quite convoluted and unpredictable, similar to conduit networks in karst aquifers.

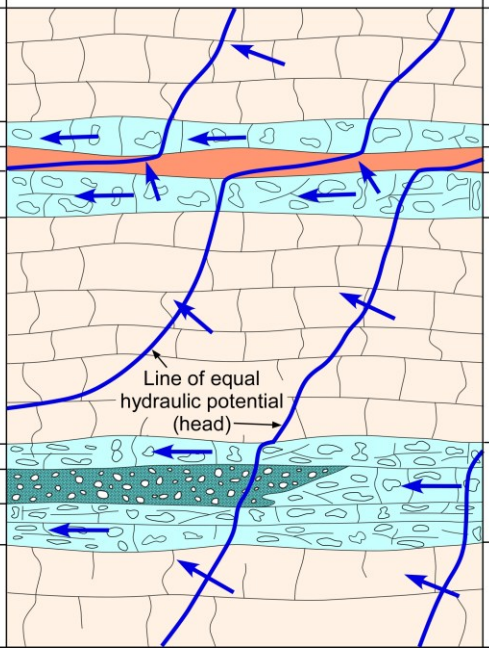
POSITION IN BASALT FLOW	LITHOLOGY	GEOHYDROLOGY	HYDRAULIC CONDUCTIVITY
Center	See below		See below
Base	See below		See below
Interbed	Clay, silt		Very low; confining unit
Top	Basalt, vesicular, brecciated		High; may be extremely high
Center	Basalt, less vesicular than above, vesicularity decreases with depth: vertically jointed		Horizontal: low to moderate, depends on vesicularity and degree of fracturing Vertical: depends on degree of fracturing; commonly several orders of magnitude lower than horizontal conductivity
Base	Basalt, vesicular, platy		Base and interbed moderate to high; combined with top zone, may be extremely high (interflow zone)
Interbed	Sand, gravel		
Top	See above		
Center	See above		See above

Figure 2.10.7 Effect of basalt aquifer lithology and sedimentary interbeds on hydrogeologic characteristics and groundwater movement. Blue arrows show general direction of groundwater flow in cross section. Modified from Lindholm, 1996.

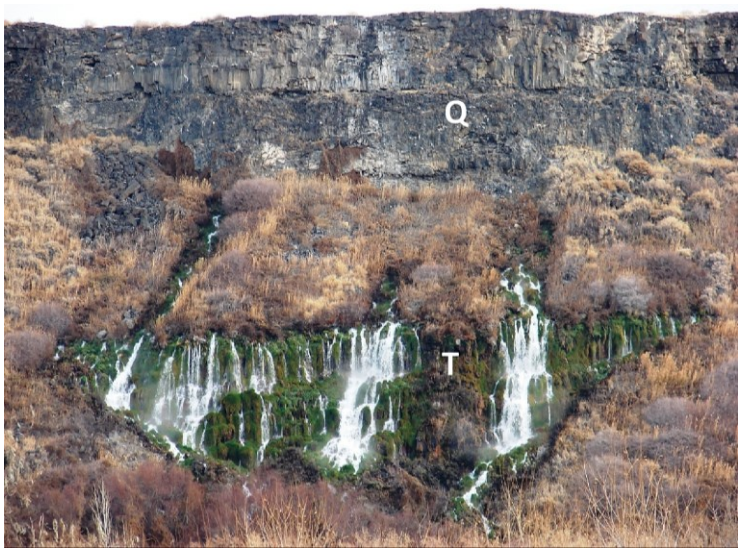


Figure 2.10.8 Minnie Miller Falls Springs, in Thousand Springs State Park south of Hagerman, central Idaho, issue from a rubble/pillow lava zone at the base of Quaternary basalt (Q) sitting over Tertiary basalt (T). A higher level of spring discharge is also visible above this contact, within the gray blocky Quaternary basalt. This level can be quite active depending on the pattern of aquifer recharge and groundwater withdrawal for irrigation. Some of the primary groundwater discharge locations may be covered by talus. Photograph and annotations from Farmer, 2021.

In his seminal work *Large Springs in the United States,* published in 1927, Oscar Meinzer of USGS lists 16 first-magnitude springs in Oregon, all issuing from pseudokarst, making it the state with the most first-magnitude springs in the country; Idaho's Snake River Basin had 15, Florida (Floridan Aquifer together with adjacent parts of Alabama and Georgia) had 11, the Ozark region of Missouri and Arkansas had eight, Sacramento River Basin in California had seven, "Balcones fault belt" in Texas had four; Montana had three, and northern Alabama had one, for a total of 65 first-magnitude springs in the country at that time. In summary, 38 out of 65 first-magnitude springs in the country, or more than half, issued from pseudokarst. All these springs and pseudokarst aquifers they drain

are explained in detail in a book entitled *Notable Springs of the United States* (Kresic, 2024). The book is free for download from various web sites (e.g., Groundwater Academy at www.groundwateracademy.org; and International Groundwater Resources Assessment Center, IGRAC, at https://un-igrac.org/data/resources/notable-springs-of-the-united-states/). Below are several photographs of some of the largest and most spectacular pseudokarst springs in the country from this high-resolution electronic book.



Figure 2.10.9 Historic postcard of the Thousand Springs 8 miles southeast of Hagerman, central Idaho, before development for a power plant. Thousands Springs is the second largest spring group in the country, after Malad Gorge Springs which are about 5 miles northeast of Hagerman. *Northwest Historical Postcards Collection*, PG 9, Postcard Collection, University of Idaho Library Special Collections and Archives http://www.lib.uidaho.edu/special-collections/.



Figure 2.10.10 Box Canyon Springs are in Earl M. Hardy Box Canyon Springs Nature Preserve, one of six units of Thousand Springs State Park south of Hagerman, central Idaho. The headspring, with the long-term average discharge rate of 365 cfs, is the third largest individual spring in the country after Big Spring in Missouri and Wakulla Spring in Florida, based on publicly available, long-term USGS discharge measurements. The extraordinarily large discharge rate of the majestic Box Canyon headspring can be best appreciated at this waterfall some 2,300 feet downstream of the spring. The USGS gaging station is about 20 yards downstream of the waterfall. Photograph taken in July 2019.



Figure 2.10.11 Springs in Ahjumawi Lava Springs State Park, California. Photograph copyright Michael Wier, all rights reserved. Printed with permission.



Figure 2.10.12 One of two Metolius Springs which form headwaters of the Metolius River in Oregon. Water bubbles out of ground near willows in the center of picture. Courtesy of U.S. Forrest Service – Pacific Region.



Figure 2.10.13 Upper Lake of Blue Lakes Springs near Twin Falls, Idaho. Photograph by Kenneth Skinner, U.S. Geological Survey, Idaho, 2005.

Chapter 3 Important Karst Springs of the United States: West and Midwest

3.1 Missouri, and Arkansas

The Ozark Plateaus aquifer system hosts three of the five largest individual karst springs in the country: Big Spring in Carter County, Missouri, Greer Spring in Oregon County, Missouri, and Mammoth Spring in Fulton County, Arkansas. These three springs also have the longest record of continuous daily discharge rate measurements of all springs in the country, through present day, with Big Spring and Greer Spring dating back to 1921. The Ozarks karst aquifer has nine (possibly ten) first magnitude individual karst springs in total, eight (or nine) in Missouri and one in Arkansas, with many more second and third magnitude springs. As of May 2026, there are 4,456 springs in the Missouri Department of Natural Resources GIS database (GeoSTRAT, Fig. 3.1.1), the vast majority being karst springs. In comparison, there are 2,905 springs in the Kentucky Division of Geographic Information database, and 1,122 springs in the Florida Department of Environmental Protection Geospatial Open Data database, all three states widely recognized as being the most "karstic" in the United States.

Table 3.1.1 includes discharge rates for all first magnitude springs (average flow is greater than 100 cfs) and some better-known second magnitude springs (10-100 cfs) in the Missouri Ozarks as listed by Vineyard and Feder in 1974. The first magnitude spring in Arkansas, Mammoth Spring in Fulton County, is included by Vineyard and Feder because virtually its entire drainage area is in Missouri.

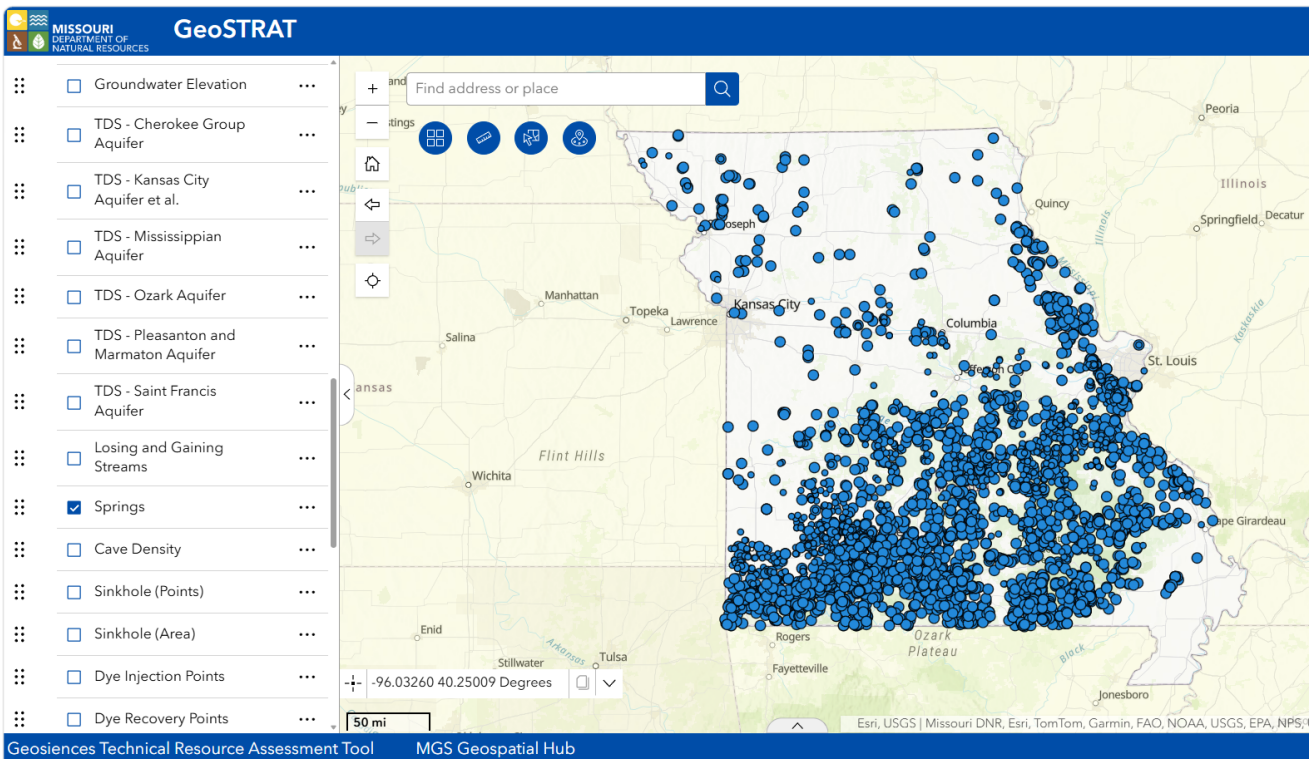


Figure 3.1.1 Screenshot of the landing page of the Missouri Department of Natural Resources GIS that includes various layers such as bedrock geology, springs, sinkholes, dye traces, losing and gaining streams, and many more information. Available at https://experience.arcgis.com/experience/c247577fff5b4a06bae08bb9837616a0

Name	County	Latitude	Longitude	Vyneard and Fedder, 1982; average, million gallons per day; MGD	Missouri DNR GIS		Missouri DNR[1] average; cubic feet per second cfs	Missouri DNR[1] average; million gallons per day MGD
					Qmin (gallons per minute)	Qmax (gallons per minute)		
Big Spring	Carter	36°57'08.31'' N	90°59'38.98'' W	276 MGD	105,917	583,440	470 cfs	304 MGD
Greer Spring	Oregon	36°47'11.47'' N	91°20'56.14'' W	214 MGD	46,675	405,266	360 cfs	233 MGD
Mammoth Spring	Fulton (AR)	36°29'52.23'' N	91°32'09.52'' W	155-278 MGD**	107,712	193,433	347 cfs	224 MGD
Bennet Spring	Dallas	37°42'59.64'' N	92°51'24.24'' W	100 MGD	39,943	77,642	185 cfs	120 MGD
Rainbow Spring***	Ozark	36°43'10.95'' N	92°11'14.40'' W	100 MGD*	21,094	104,122	155 cfs	100 MGD
Maramec Spring	Phelps	37°57'13.48'' N	91°31'58.74'' W	96 MGD	25,133	291,720	149 cfs	96 MGD
Blue Spring	Shannon	37°09'57.73'' N	91°09'44.09'' W	90 MGD*	27,825	105,917	139 cfs	90 MGD
Alley Spring	Shannon	37°09'17.15'' N	91°26'29.66'' W	81 MGD	24,235	475,728	125 cfs	81 MGD
Welch Spring	Shannon	37°23'38.20'' N	92°34'27.44'' W	75 MGD*	31,416	148,553	116 cfs	75 MGD
Boiling Spring	Pulaski	37°53'22.71'' N	92°02'06.70'' W	68 MGD*	25,133	31,506	105 cfs	68 MGD
Montauk Springs	Dent	37°27'36.91'' N	91°40'59.67'' W	53 MGD*	17,144	55,202		
Hahatonka Spring	Camden	37°58'26.67'' N	92°46'01.50'' W	48 MGD	44,027	44,027		
North Fork Springs	Ozark	36°43'28.29'' N	92°11'13.30'' W	43-49 MGD**	29,711	33,795		
Roubidoux Springs	Pulaski	37°49'31.24'' N	92°12'05.55'' W	38 MGD	2,091	86,170		
Pulltite Spring	Shannon	37°20'06.05'' N	91°29'24.54'' W	0.4-92 MGD**	2,648	63,730		
Round Spring	Shannon	37°16'49.61'' N	91°24'40.57'' W	26.5 MGD	4,488	233,376		
Roaring River Spring	Barry	36°35'30.00'' N	93°49'56.95'' W	20 MGD	3,608	79,438		

Table 3.1.1 Flow rates of large karst springs in the Ozarks of Missouri and Arkansas. Notes: [1]Value from: Missouri Geological Survey, July 26, 2021. *Follow the Water. Understanding Springs in the Missouri Ozarks*. (https://storymaps.arcgis.com/stories/b046232085004488a24ec12e2631e793). ** Average flow not given; range of several measurements *** Previously known as Double Spring. From Kresic, 2024.

For comparison, the table also includes average flow rates, in million gallons per day, for all first magnitude springs as provided by the Missouri Department of Natural Resources (DNR) in July 2021. Missouri's GeoSTRAT GIS database provides minimum and maximum flow rates for quite a few springs; however, the average flow rates are not listed, and it is not clear how the reported values were derived (e.g., from a few manual measurements, based on continuous monitoring for a specific period, etc.)

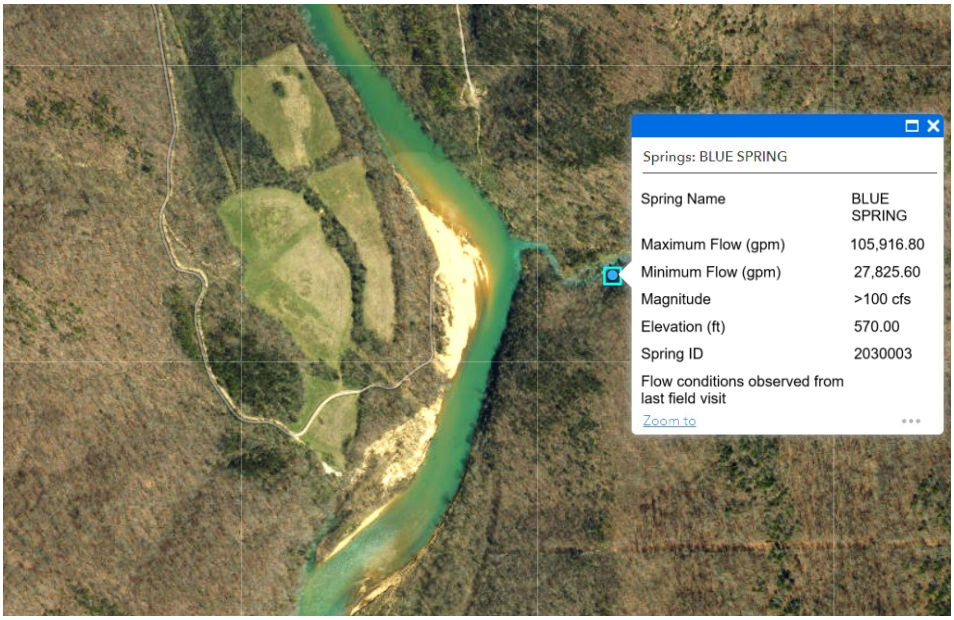


Figure 3.1.2 Screenshot of the Missouri DNR GIS database displaying search results for Blue Spring near Eminence, Shannon County (top image) and bedrock geology of the area (bottom image shows highlighted Ordovician and Cambrian Eminence Dolomite from which Blue Spring, denoted with yellow circle, discharges) Data and images: State of Missouri, Maxar | Esri Community Maps Contributors, Missouri Dept. of Conservation, Missouri DNR, © OpenStreetMap, Microsoft, Esri, HERE, Garmin, SafeGraph, GeoTechnologies, Inc, METI/NASA, USGS, EPA, NPS, US Census Bureau, USDA.

In the opinion of this author, Missouri's GeoSTRAT is the best geologic and hydrogeologic database in the country, unrivaled when it comes to communicating important aspects of the springs, bedrock geology, karst features in general including sinkholes, losing and gaining streams, dye tracing details, mining sites, and other information. https://experience.arcgis.com/experience/c247577fff5b4a06bae08bb9837616a0

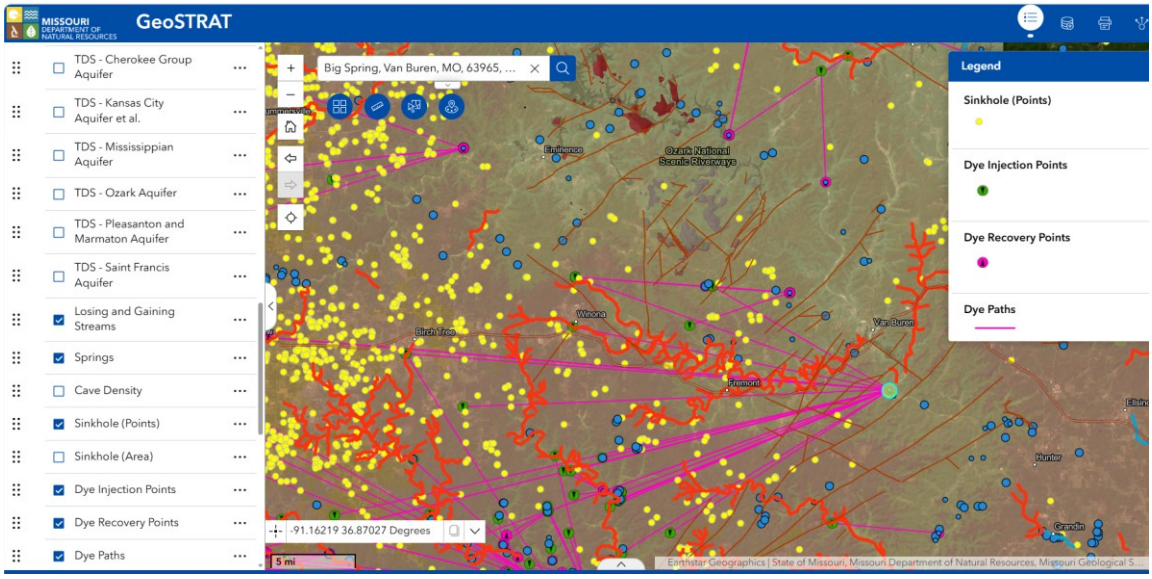


Figure 3.1.3 Screenshot of the Missouri GeoSTRAT focusing on Big Spring in Carter County near Van Buren, denoted by larger light-blue dot where dye tracing paths (pink lines) all converge.

Different sources including various web pages may give differing values for average daily flow of the same spring, often not citing where the original information came from. It is therefore important to understand the methodology of deriving an average flow rate value even when a longer period of record is available.

As discussed by Jerry Vineyard and Gerald Feder in their anthological publication *Springs of Missouri* (first published in 1974, revised in 1982, and available at https://dnr.mo.gov/land-geology/geology/karst-missouri/springs):

Through the years records of the rate of flow of the large springs of the Ozark region of Missouri have been collected. Daily records are available from 1921 to present (1972) for Big Spring in Carter County and Greer Spring in Oregon County. Records for several years are also available for other large springs. Occasional measurements of discharge are available at most springs. From these continuous and occasional measurements it is possible to determine only an approximation of the relative size of the springs. Table 1 shows the relative size of the large springs in the Ozarks. This order of magnitude has been revised from that given by Beckman and Hinchey (1944, p. 16) because of the availability of longer records.

During periods of high water many springs are flooded by nearby streams making discharge measurements difficult or impossible. For this reason the maximum discharges of many springs are not accurately known. Estimates of spring discharge during floods can be made by measuring the discharge of the receiving stream above and below the spring inflow, and using the difference as the estimate.

Spring discharge measurements are given in this report in terms of gallons per day (gpd) or cubic feet per second (cfs). Discharge measurements are routinely made in terms of cfs. then converted to gpd when considered desirable. One cfs is a measure of volume and means one cubic foot of water passing a fixed point in one second.

One cfs is equivalent to 646,000 gallons per day (gpd). Thus, to convert measurements in cfs to gpd, one simply multiplies the cfs figure by 646,000. Occasionally discharge measurements are needed in other terms. Conversion factors to other units of volume are:

1 cfs = 449 gallons per minute = 646,000 gallons per day = 1.98 acre-feet per day = 28.3 liters per second.

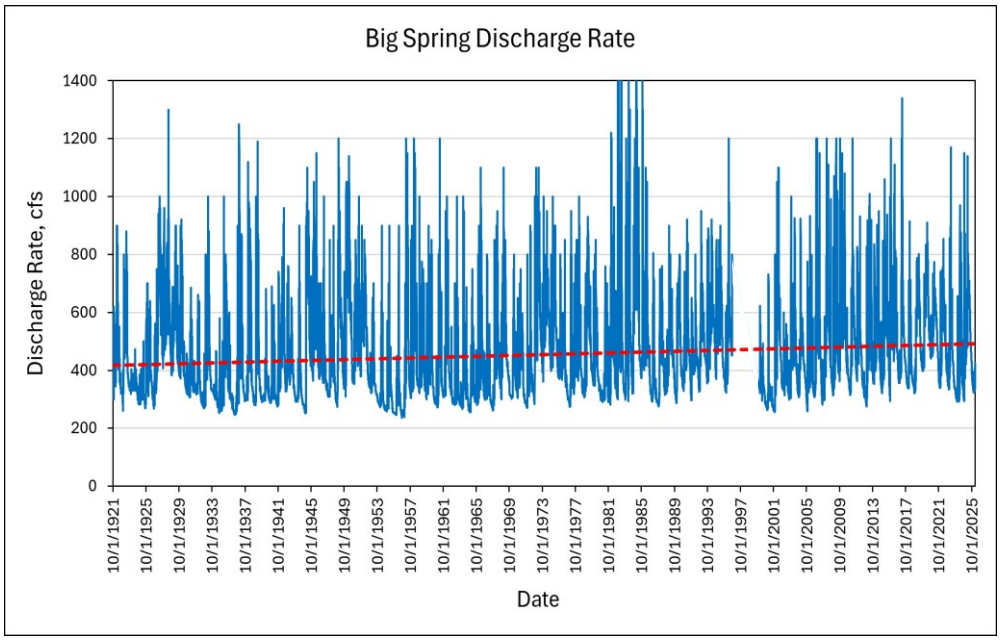


Figure 3.1.4 Hydrograph of Big Spring daily discharge rate, in cubic feet per second (cfs), October 1921 through May 2026. Vertical scale is truncated at 1,400 cfs to better show the increasing trend denoted with a dashed red line. Extreme high rates (e.g., > 1,000 cfs, >1,200 cfs, and >2,000 cfs) often could not be measured due to back flooding from the Current River. The average for the entire period of record is 454 cfs; the minimum is 236 cfs; the maximum is >2,000 cfs. The average for the last 20 years is 489.8 cfs. Data for the USGS gaging station 07067500 Big Spring near Van Buren, MO.

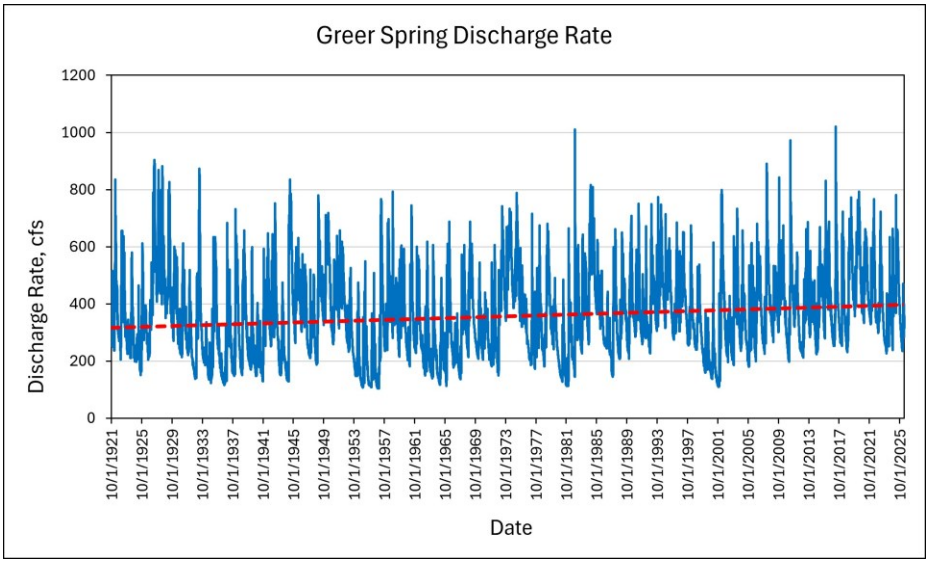


Figure 3.1.5 Hydrograph of Greer Spring daily discharge rate, in cubic feet per second (cfs), October 1921 through May 2026. The average for the entire period of record is 356 cfs; the minimum is 104 cfs; the maximum is 1,020 cfs. The average for the last 20 years is 434 cfs. Data for the USGS gaging station 07071000 Greer Spring at Greer, Missouri.

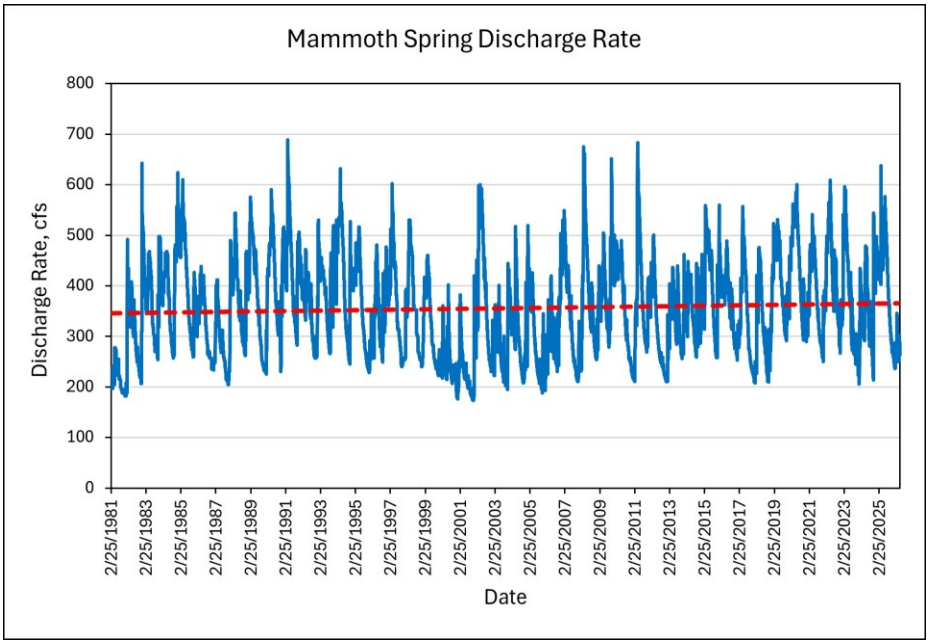


Figure 3.1.6 Hydrograph of Mammoth Spring daily discharge rate, in cubic feet per second (cfs), February 1981 through May 2026. The average for the entire period of record is 355 cfs; the minimum is 173 cfs; the maximum is 689 cfs. The average for the last 20 years is 377 cfs. Data for the USGS gaging station 07069190 Mammoth Spring at Mammoth Spring, Arkansas.

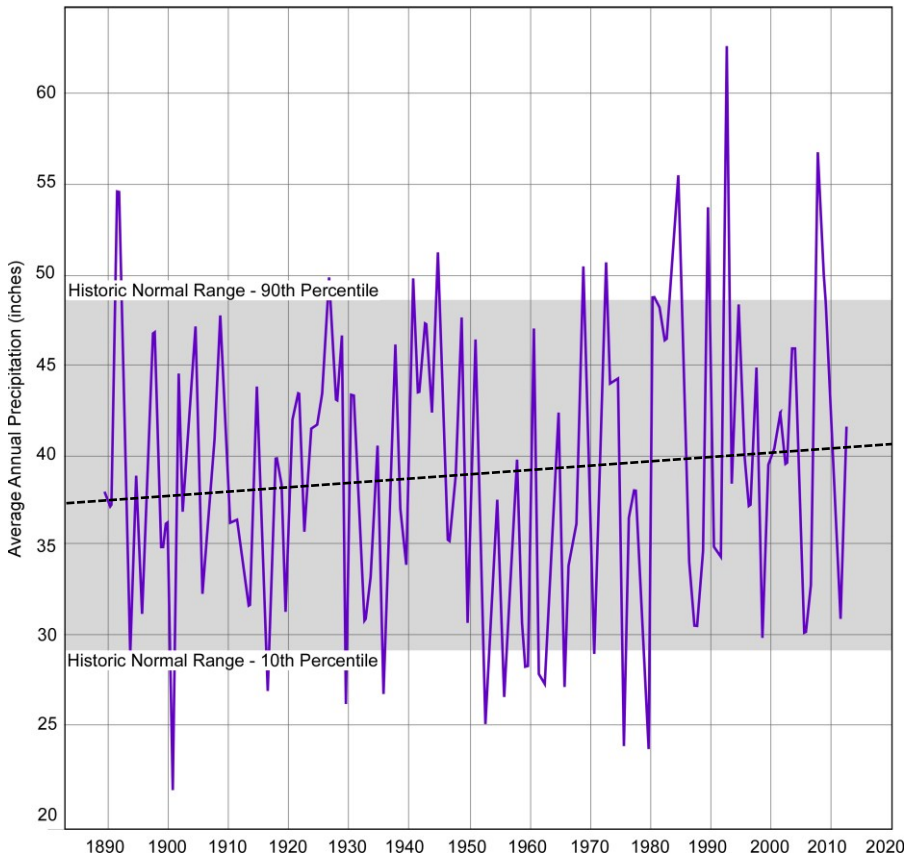
The increasing trend in the discharge rate of the three largest springs in Missouri shown in the above hydrographs could most likely be attributed to the observed increasing precipitation in Missouri as discussed by Leasor et al., 2024. Below are some key points of their analysis.

- Average annual precipitation has risen by 5.2” between 1979 and 2021, with the greatest increases observed during the spring (3.9”) and summer (2.6”).
- A decrease in seasonal precipitation was observed in the fall months (-1.8”).
- Extreme precipitation events (> 2.0”) have become more frequent, with an increase of 2.0 days annually between 1979 and 2021.
- Most of the state experiences > 40% of its annual precipitation total on the 10 wettest days of the year.
- Missouri’s average annual dew point temperature has increased since 1979.

- Models of future climate predict annual precipitation is expected to increase in the future, with the largest seasonal increases likely during spring.
- An increase in the frequency and severity of heavy precipitation events are also projected to increase by mid-century under continued emissions.
- Decreases in total precipitation and greater variability are projected during the summer. These changes are larger under the very high probability scenario and for the late 21st century (2080-2099).

These findings are echoed by Pat Guinan, Missouri's state climatologist and associate professor of climatology at the University of Missouri Extension (https://www.stlpr.org/health-science-environment/2021-08-17/climate-change-driving-unprecedented-warming-and-precipitation-in-missouri):

> *Here in Missouri, we've been in an unprecedented wet period since about the early 1970s. In fact, four out of our top five wettest years have occurred since 1973: 1973, 1993, 2008 and 2015. We're seeing higher extreme events, not only on a daily basis, but when we look at the annual totals, when we look at seasonal totals, it all shows upward trends over the past few decades versus the long-term record. So if you think we've been seeing more extreme precipitation events, more flooding over the past few decades, we have — and the data really back that up.*



One of the contributors is the warming temperatures [that are] allowing more water vapor into the atmosphere. These warming temperatures are occurring in the oceans, and the Gulf of Mexico is the source region for moisture into the middle part of the country. When we get those southerly winds, it advects that moisture — Gulf of Mexico moisture — into the middle part of the country [and] you get atmospheres that become more efficient at creating precipitation. That's one of the reasons we're seeing this increasing trend of precipitation.

Figure 3.1.7 Recorded annual average precipitation in Columbia, MO with trendline. Modified from City of Columbia Climate Action & Adaptation Plan.

The following excerpts are from the City of Columbia, Missouri, Climate Action & Adaptation Plan. They refer to a figure shown here as Fig. 3.1.7 and are applicable to Missouri as a whole: *Figure shows recorded rainfall in Columbia since 1890. In the past 30 years, annual rainfall was nearly 10 percent more than the past century's average. There has been more rainfall in all four seasons, but the most notable increase has been during winter*

and spring. Heavy rain events have also become more frequent. During the past three decades, extreme rainfall events, with 3 or more inches of rain, are happening more than twice as often compared to the 1890-1984 average. The Columbia area more often sees unusually heavy rainfall during spring and fall, rather than winter or summer. (ColumbiaCAAP_Climate-trends-summary_final.pdf)

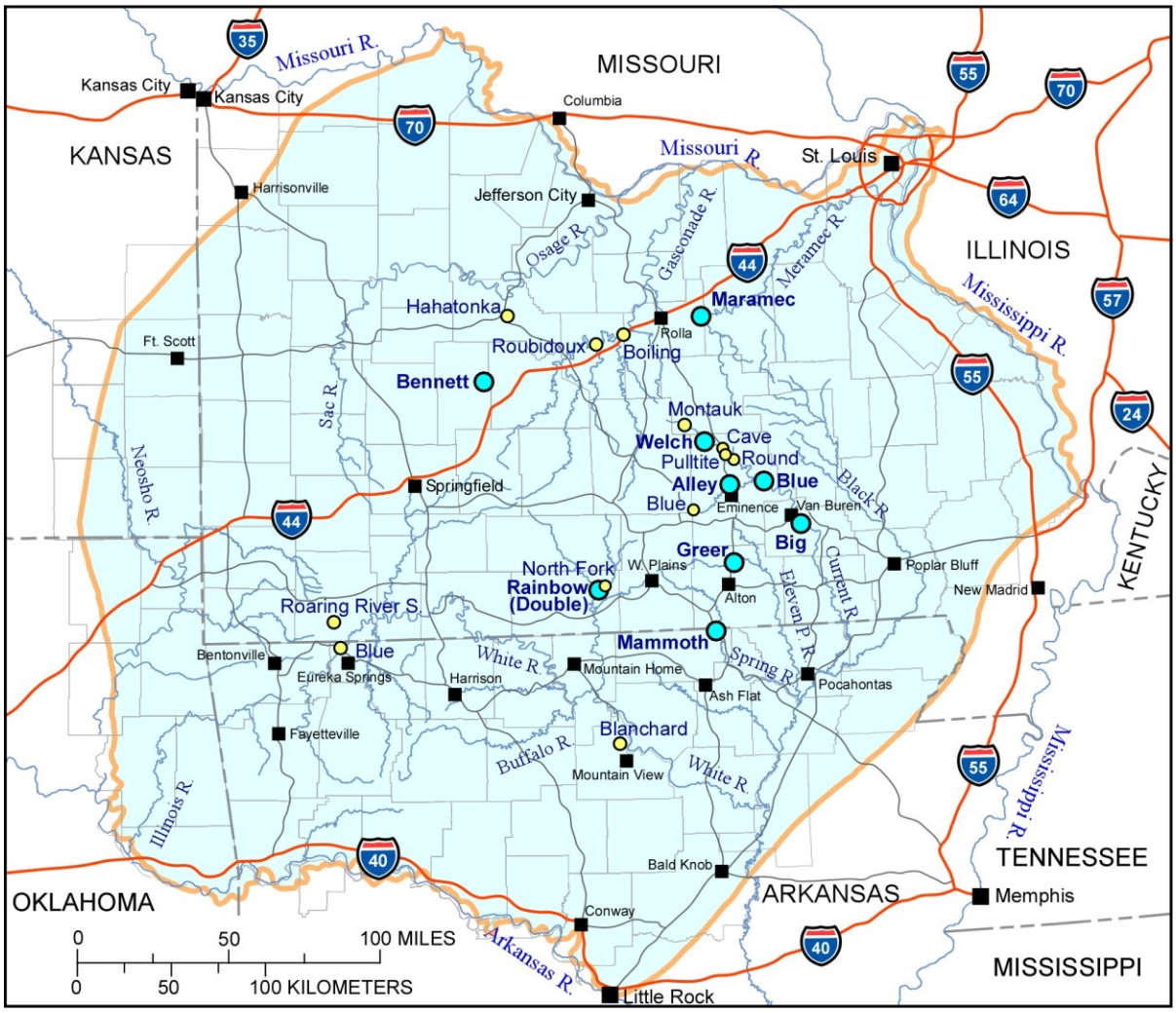


Figure 3.1.8 Large and notable karst springs in Missouri and Arkansas featured in this book. Large blue circles and bold blue names denote first magnitude springs (average discharge rate >100 cfs); smaller yellow circles are some better-known 2nd magnitude springs (average discharge rate 10-100 cfs.). Note: Boiling Spring in Pulaski County, Missouri is possibly a first magnitude spring. Modified from Kresic, 2024.

As seen in Fig. 3.1.8, there are nine first-magnitude karst springs in an area of about 10,000 square miles of the Ozarks karst aquifer. Together with Florida springs, this is the largest concentration of first-magnitude karst springs in the country. As mentioned earlier and discussed in more detail in the next chapter, Florida also has 9 individual, first-magnitude, true karst springs which are spread over a much larger area of the Floridan aquifer (see Chapter 3.1 in Volume 1). Sadly, many beautiful Florida springs and spring-fed rivers (which in Florida are often called "spring groups") are experiencing a catastrophic demise, both in terms of rapidly declining discharge rates and water quality. In contrast, the magnificent Missouri springs are as strong and pristine as ever and enjoy a protected

status as state or national parks, or simply because the communities and citizens of Missouri, as well as the state regulatory agencies and lawmakers, are taking good care of them and preserving them for future generations.

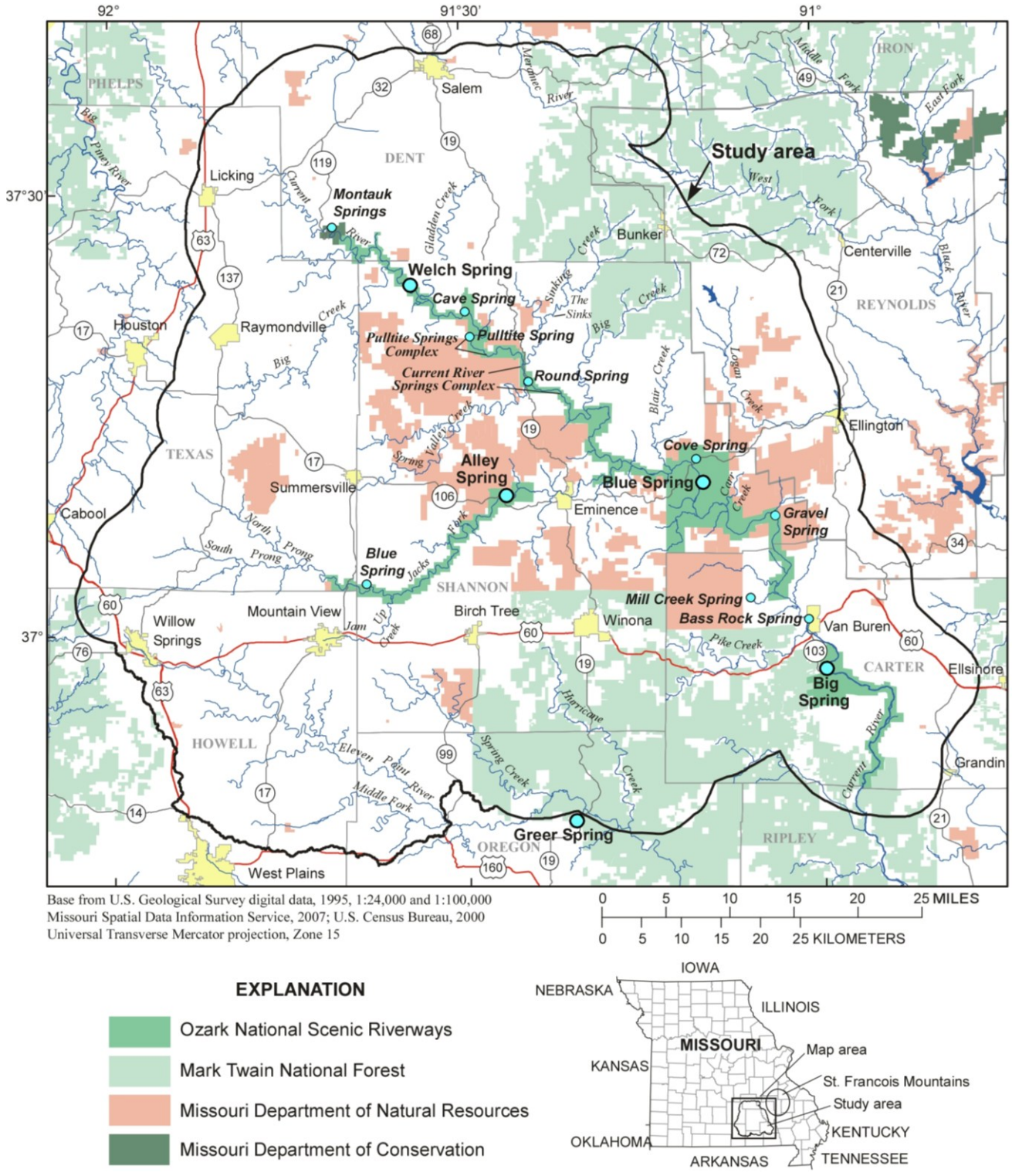


Figure 3.1.9 Ozark National Scenic Riverways (ONSR) area showing locations of streams, selected springs, and public land ownership. Larger blue circles mark first-magnitude springs. Modified from Mugel et al., 2009; USGS.

Ozark National Scenic Riverways (ONSR) located in south central Missouri, is part of the National Park Service, within the Department of the Interior. The 134 miles of scenic and wild riverways are comprised of two rivers, the Current and Jacks Fork (Fig. 3.1.9). ONSR was the first national park to protect a river system in its wild, undammed state. The movement for the park started as a grassroots reaction to plans to build a series of dams on the two rivers. While local residents and conservation groups in nearby cities disagreed on how the rivers should be preserved, most agreed the rivers should not be dammed but should remain free-flowing. In 1964, after much debate and compromise, Congress officially established Ozark National Scenic Riverways. The State of Missouri donated three state parks—Alley, Big Spring, and Round Spring—to the National Park Service as a gift from Missouri to the American people. Land along the rivers between the former state parks was purchased, and the whole area knitted together as America's first national river park. The successful effort to preserve these rivers was the prototype for the Wild and Scenic Rivers Act of 1968, which protects many of America's free-flowing rivers.

Since 1964, millions of people have come to ONSR to enjoy the rivers' beauty, along with many recreational opportunities. Canoeing and water sports are popular, as are hiking, camping, and birdwatching. Spring wildflowers and autumn tree colors here are spectacular. Today the park continues to protect the rivers, the springs, and the watershed to ensure enjoyment of these natural and cultural resources for future generations. The park's springs offer people beauty and respite, and they supply most of the flow for the Current and Jacks Fork rivers. Without the protection of National Park status, these and other rivers would have eventually disappeared because of the trend of river damming at the time.

As noted by the NPS (https://www.nps.gov/ozar/learn/nature/spring-laws.htm)

The first paragraph of the enabling legislation for Ozark Riverways reads: Be it enacted by the Senate and House of Representatives of the United States of America in Congress assembled, That, for the purpose of conserving and interpreting unique scenic and other natural values and objects of historic interest, including preservation of portions of the Current River and the Jacks Fork River in Missouri as free-flowing streams, preservation of springs and caves, management of wildlife, and provisions for use and enjoyment of the outdoor recreation resources thereof by the people of the United States,…



Figure 3.1.10 Blue Spring on Jacks Fork near Mountain View is a popular swimming hole. It is in the Ozark National Scenic Riverways (ONSR), a part of the National Park Service system.

*The second specific purpose listed for the establishment of the park is "*preservation of springs*", out ranked only by "preservation of portions of the Current River and the Jacks Fork River". The word preservation was chosen here and not the weaker term conservation. It is clear that the springs must be protected to the highest standard. Big Spring is certainly by virtue of its magnitude alone one of, if not the most, valuable single resource feature in the park. It is clear therefore that actions and activities that affect Big Spring must be judged using the highest standard.*

Springs are unique ecosystems. Their stable thermal regime and water chemistry characteristics provide a unique environment for a variety of flora and fauna not commonly found in the main channel of their receiving stream. By definition springs are small systems, which makes them more sensitive to disturbances than the main river channel.

Watercress and other aquatic macrophytes play a key role in the spring ecosystem by providing organic input, habitat for macro invertebrates and substrate for periphyton growth (Converse 1994). Watercress grows mainly in shallow areas and is easily dislodged by wading activities. Wading activities by large numbers of visitors in the spring branch would significantly reduce the watercress biomass in the spring ecosystem. This biomass reduction would negatively affect the thermal regime present in these springs. At least one species of caddisfly is known to occur only in Big Spring and Greer Spring. Small changes in small sensitive systems like springs can have large effects. In addition, the export of large amounts of macrophyte material to the receiving river is significant (Converse 1991). No less than 38 animal species are restricted in distribution to Ozark springs and subterranean waters (Vineyard and Feder 1974). Several aquatic invertebrate species are known from as few as two Ozark springs including springs within Ozark National Scenic Riverways.

The scenic values of our large springs are probably the most obvious values to park visitors. The number of publications using photographs of these springs on their cover or in the body every year evidences this. Tens of thousands of people come to Big and Alley Springs every year just to look at these jewels of the Ozarks.



Figure 3.1.11 Big Spring in Carter County, Missouri, the largest individual karst spring in the United States based on 105 years (1921-2026) of verified daily measurements by USGS. The average discharge rate for the entire period of record is 454 cfs; the minimum is 236 cfs; the maximum is >2,000 cfs. The average for the last 20 years is 489.8 cfs. Photo taken in June 2025.

Big Spring, Carter County, Missouri

As stated by the National Park Service (NPS) on its website for the Ozark National Scenic Riverways (ONSR), *The Big Spring is sometimes called America's biggest spring. In reality there are three contenders for that title: Big Spring, Idaho's Snake River Springs Complex and Florida's Silver Springs complex. Since the flow from springs varies with local rainfall, any of these three might be biggest on any given day depending on the weather in Missouri, Idaho, and Florida! The truth is they are all about the same size.*



Figure 3.1.12 Big Spring rises at the base of a cliff formed in the Late Cambrian Eminence Dolomite. Photos taken in October 2023 during below-average flow conditions.

Nevertheless, publicly available official USGS records of the flow rates at these springs confirm that the Big Spring is the largest individual spring in the United States based on its average discharge rate of 454 cfs for the period of record, October 1921–May 2026 (Fig. 1.1.4). The Malad Canyon Springs and The Thousand Springs in Idaho are the two largest groups of closely spaced springs with an average discharge rate of 1,229 and 1,198 cfs respectively (Kresic, 2024), whereas the headspring of the Silver Springs complex, sometimes referred to as Mammoth Spring, has experienced a dramatic decline over the several decades preceding 2010 when USGS stopped reporting its discharge rate which, at the time, was significantly lower than that of Big Spring. On the other hand, Wakulla Spring in Florida has experienced an apparent increase in discharge since 2010, possibly due to saltwater intrusion from the Gulf of Mexico. Notably, USGS never performed continuous daily discharge measurements at Wakulla Spring, only sporadic field measurements which were discontinued altogether in 2010. Based on these measurements, performed between 1907 and 2010, the average discharge rate of Wakulla Spring was 409 cfs. Since 2010, there were several periods when the discharge rate of Wakulla Spring was measured and reported by private contractors and agencies other than USGS, applying a variety of inconsistent methods. If consistent daily flow measurements for a sufficient period in the future are available, Wakulla Spring may become a contender for the title of the largest karst spring in the country.



Figure 3.1.13 Powerful boil of the Big Spring in June 2025, during high flow conditions.

Big Spring was one of Missouri's first state parks. It was a state park from 1924 until 1969 when the people of Missouri donated it, along with Alley and Round Spring State Parks, to the National Park Service to become a part of the ONSR. It is located four miles south of Van Buren at the end of State Route 103 (see Fig. 3.1.9).

Big Spring rises at the base of a cliff formed in the Late Cambrian Eminence Dolomite through a jumble of giant boulders that forces the water to flow through the openings between the rocks, thus causing the dramatic "boil" at the spring orifice. As described by Vineyard and Feder (1982), about one-third of the way up the slope, amidst large breakdown boulders, is the entrance to a small pit cave (closed to visitors) that can be entered with the aid of ropes giving access to a small room through which part of the Big Spring water flows. Here the water is quiet, but it moves swiftly toward the orifice. Divers of the St. Louis Underwater Recovery Team entered Big Spring through this relatively quiet pool to explore the conduit system channeling water to the spring orifice. They also descended into the main boil area of the spring, searching carefully for "dead spots" in the powerful current. Neither effort was successful in getting past the breakdown into the undisturbed conduit system of the spring, though one diver reported a tantalizing glimpse of immense caverns beyond. Since then, research dives showed the spring is at least 80 feet deep (Missouri Department of Conservation, 2023a).

Greer Spring, Oregon County, Missouri

Greer Spring, one of the most significant natural features in Missouri, is located 18 miles south of Winona on the right side of Missouri Highway 19. It is the second largest spring in the Ozarks with a mean daily discharge of 356 cfs (Fig. 3.1.5). The spring and the surrounding area, designated as a protected natural area in 2021, are in Mark Twain National Forest and owned and operated by the USDA Forest Service. Greer Spring was designated a National Natural Landmark in 1980.

Greer Spring flows from two openings – an upper cave outlet and a lower orifice in the bed of the gorge some 250 feet from the cave. The spring cave may be explored for several hundred feet, past a waterfall and plunge pool into a low-ceiling passage in which the air space above water level gradually decreases until the wide, bedding-plane opening becomes completely water-filled, the rise of an underground stream. The lower orifice is a powerful streambed boil that obscures the conduit in the Early Ordovician Gasconade Dolomite that feeds the spring. Divers have been successful in penetrating this opening to a depth of more than 100 feet, despite the powerful current (William Cate, oral comm., 1972; from Vineyard and Feder, 1982).



Figure 3.1.14 Ground level image of Greer Spring cave outlet from a bluff in the Early Ordovician Gasconade Dolomite during low flow. Photo courtesy of John Van Brahana.

Greer Spring has the longest spring branch of all first magnitude springs in Missouri, flowing just over a mile before entering the Eleven Point River where it more than doubles its flow. In addition to hydrogeologic significance, Greer Spring is an important natural aquatic community supporting a diverse assemblage of native fish species and aquatic invertebrates, including some species of conservation concern such as the cold-water crayfish (*Faxonius eupunctus*). The flora of the spring branch and associated Ozark fens is diverse as well. Greer Spring is critically important to the hydrology of the Eleven Point River, a National Wild and Scenic River (Missouri Department of Conservation, 2023b.)

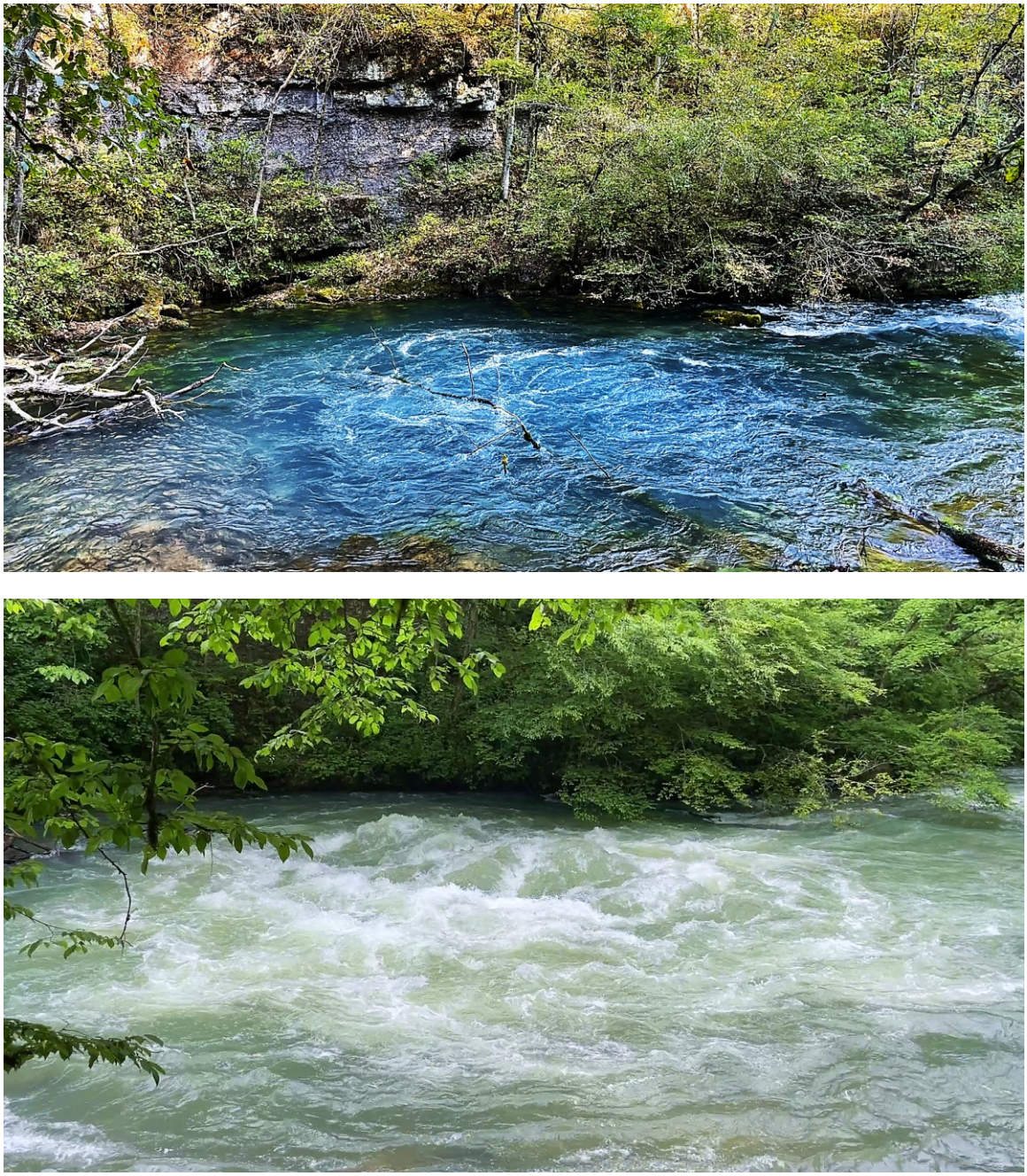


Figure 3.1.15 Boil of the Greer Spring about 250 feet downstream from the spring's cave outlet. *Top*: during below-average flow conditions, October 2023. *Bottom*: during high flow conditions, June 2025.

Vineyard and Feder in their publication on Missouri springs (1974) provide this inspiring story of the lucky destiny of Greer Spring:

"Missouri's second largest spring may be unsurpassed in the beauty of its wilderness setting, so colorfully described by Owen (1898, p. 82-93) and little changed in 75 years. In the days when water power was the wave of the future, engineers dreamed of taming the wild Greer Spring branch, with a flow said to be " *... two hundred and eighty yards per minute, With no appreciable variation,"* by building a dam across the precipitous gorge carved by the spring branch. However, Miss Owen showed foresight far in advance of her time when she said, *"The high walls of Greer Spring gorge will, of course, far more than double the value it would otherwise possess, when it becomes desirable to control and turn to practical account the power now going so cheerily to waste, but the artistic loss will be proportionately severe."* Fortunately for this generation and others to come, those who have owned the spring since it was named for Captain Greer nearly a hundred years ago, appreciated its beauty and no dam was ever built in the gorge. Greer Spring is as wild and beautiful today as it was in the 1880's."

Mammoth Spring, Fulton County, Arkansas

The orifice of Mammoth Spring is a short distance across the Missouri-Arkansas line at the head of the Spring River in the town of Mammoth Spring, Arkansas, east of state highway 63 (Fig. 3.1.16). Mammoth Spring is a permanent spring that ascends upward under pressure along a major northwest-trending basement lineation (fault) and forms a pool which has two natural outlets (Fig. 3.1.17). The spring discharges from the Jefferson City Dolomite, a low-yielding aquifer that is not capable of such large discharge, hence the source of the flow is attributed to deeper dolomitic formations which include Potosi, Eminence, Gasconade, and Roubidoux Formations, ranging in age from Upper Cambrian through Lower Ordovician (Brahana, 2023).



Figure 3.1.16 Google Earth map of the Mammoth Spring area. The Spring pool (shown with dark blue arrow) is in the upper center-left, with two outlets (shown with yellow arrows) discharging into a marshy lake. Dam for the historic Arkansas Missouri Power Company electric power plant is in bottom center. Spring is in the Mammoth Spring State Park. Imagery ©2023 Maxar Technologies, State of Arkansas, USDA /FPAC/GEO, Map data ©2023.

Vineyard and Feder (1974) detail the quest to establish drainage area of the Mammoth Spring including various dye tracing tests. Here is an interesting excerpt: *"Beckman and Hinchey discussed the probable recharge area of Mammoth Spring and recounted local stories of water tracing by cornstalks and straw dumped into Grand Gulf near Thayer, Missouri. In 1966 fluorescein dye was traced from Grand Gulf to Mammoth Spring by Toney Aid in*

a test monitored by Vineyard (Aid, unpublished report, 1966). Subsequently Aley (written comm., 1972) also traced water from a Missouri locality into Mammoth Spring. Grand Gulf is one of the most spectacular surface karst features of the Ozark region; it was recently (1972) acquired by a private foundation for eventual donation to a state or federal agency for the enjoyment of the public.

Owen (1898; 1970, p. 95-101) gave a colorful description of Grand Gulf as it was in the mid-1880's. At that time it was possible to enter and explore a large underground stream by boat – the same stream that emerges some 9 miles away as Mammoth Spring. Always a careful observer, geologist Owen mentioned (p. 100) the blind white fish in the stream: "The small eyeless fish had been noticeable in the water everywhere but now came swimming about the boat in an astonishing multitude, and as unconscious of any possible danger as bees in a flower garden."



Figure 3.1.17 *Left*: Eastern outlet of the Mammoth Spring pool. *Right*: Southern outlet of the Mammoth Spring pool.

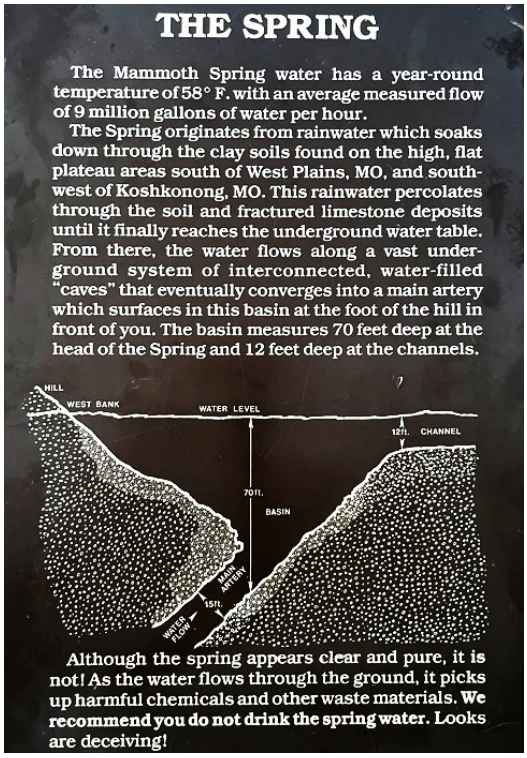


Figure 3.1.18 One of many panels that inform visitors of the Mammoth Spring State Park about various natural and cultural features in the park. This panel states that "...water has a year-round temperature of 58 ^{O}F, with an average measured flow of 9 million gallons of water per hour, MGH (the updated number is 9.5 MGH or 355 cfs – see Figure 3.1.6). And that "the water flows along a vast underground system of interconnected, water filled "caves" that eventually converges into a main artery which surfaces in this basin at the foot of the hill in front of you. The basin measures 70 feet deep at the head of the Spring and 12 feet deep at the channels."



Figure 3.1.19 *Left*: Dam, headwater of the Spring River, built by the Mammoth Spring Milling Company in 1888, and hydroelectric plant built by the Arkansas-Missouri Power Company in 1925-1926. The extraordinary flow of Mammoth Spring is on full display here. *Right*: Original generator in the former power plant building.



Figure 3.1.20 Photographs from the Grand Gulf State Park in Missouri, about 9 miles northwest of Mammoth Spring. Courtesy of Missouri State Parks.

Bennett Spring, Dallas County, Missouri

As explained by James E. Vandike in a comprehensive hydrogeologic report with excellent visuals ("The hydrogeology of the Bennett Spring area, Laclede, Dallas, Webster, and Wright Counties, Missouri", 1992) Bennett Spring, less than a mile west of the Laclede County line in Dallas County, is considered the third largest spring in Missouri and the largest spring in the Niangua River basin. The spring rises from a steeply inclined water-filled cave passage developed in the Gasconade Dolomite on the east side of Spring Hollow, about 1.3 miles upstream from its confluence with the Niangua River. Divers have explored and mapped the inclined spring conduit to a depth of about 80 feet and a horizontal distance of about 130 feet. The passage continues, but gravel chokes most of it. Higher velocity of water, resulting from the decrease in cross-sectional area, has halted exploration (Porter, 1986).

Vandike's analysis shows that, as of 1990, forty-one years of discharge records were available from the U.S. Geological Survey (1916-1919, 1928-1941, 1965-1990), and the average discharge was 170 ft^3/sec (cfs), or about 110 million gallons per day. Fig. 3.1.22 includes additional flow data available as of October 2023 and shows a higher average discharge based on the last two continuous monitoring periods: 196 cfs for 10/1965 to 1/1996, and 200 cfs for 2/2007 to 9/2023.

Bennett Spring is the principal groundwater outlet for an extensive karst area in south-central Missouri. A hydrologic reconnaissance in the Bennett Spring area of Laclede, Dallas, Wright, and Webster counties, which includes the upper Niangua River, Osage Fork of the Gasconade River, and Dry Auglaize Creek, identified nearly 40 streams that lose significant volumes of surface flow into the karst groundwater system. The Bennett Spring recharge area, based on water tracing and existing potentiometric map data, consists of a 265 mi^2 area east, south, and southwest of the spring. Precipitation, discharge, and specific conductivity data show that the discharge of Bennett Spring begins increasing generally within a few hours after precipitation due to pressure-head increase in the recharge area, but the water introduced into the aquifer from a precipitation event does not reach the spring for several days. The magnitude of flow increase depends greatly on soil moisture conditions; flow increases occur after precipitation when soils are wet than during relatively dry conditions (Vandike, 1992).

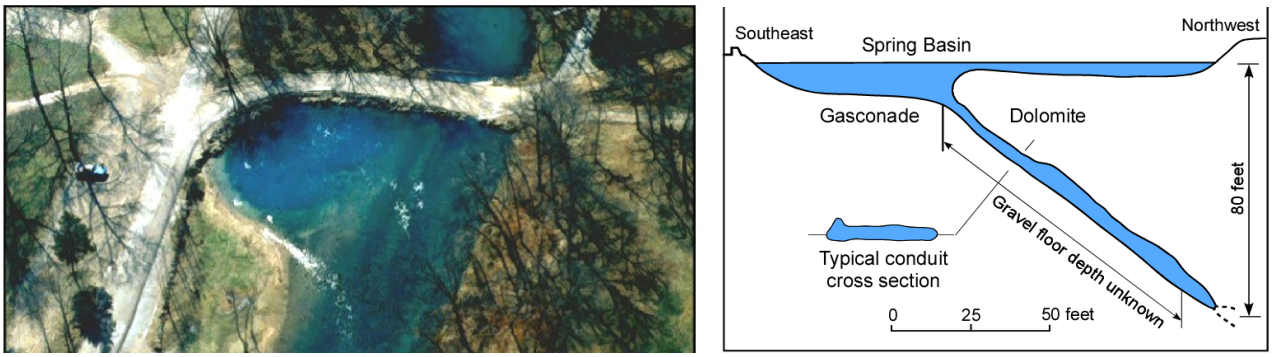


Figure 3.1.21 *Left*: Bennet Spring in Dallas County, Missouri. Photo courtesy of Missouri Department of Natural Resources. *Right*: Longitudinal cross-section through Bennett Spring. Data supplied by D. Rimbach, M. Tatalovich, and M. Grussemeyer. Modified from Vineyard and Feder, 1982.

Bennet Spring is in a Missouri state park of the same name. The park features a nature center, a pool, dining, a store, trails, float trips, lodging, and campsites. Like the other three public trout parks in Missouri—Roaring River, Montauk State Park, and Maramec Spring Park—the Bennet Spring Park has high-quality fishing in a cold-water stream fed by the spring (Fig. 3.1.23) with specific trout fishing seasons. The stream formed by the spring is stocked daily from March 1 to Oct 31.

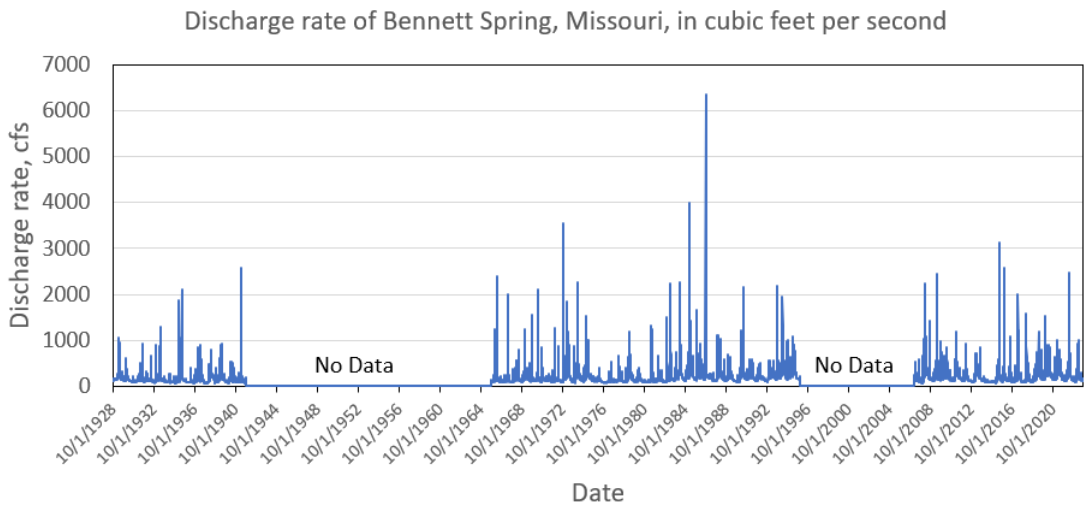


Figure 3.1.22 Hydrographs of the Bennett Springs discharge rate recorded at the USGS gaging station 06923500 Bennett Spring at Bennett Springs, Missouri. Average for the first period 10/1928 to 9/1941 is 152 cfs; average for the last two periods is 196 cfs for 10/1965 to 1/1996; and 200 cfs for 2/2007 to 9/2023. The minimum recorded rate is 55 cfs, and the maximum is 6,350 cfs, which is higher than any other spring in the Ozarks.



Figure 3.1.23 Bennet Spring is part of the Bennet Spring State Park renowned for its excellent trout fishing. Photo courtesy of https://www.springfieldmo.org/listing/bennett-spring-state-park/1143

Maramec Spring, Phelps County, Missouri

Maramec spring is located approximately 8 miles south of St. James in Phelps County, Missouri. The spring is in Maramec Spring Park which is privately owned by The James Foundation (http://maramecspringpark.com/). The spring's average discharge is 153 cfs or 99 million gallons per day.

Waters of the spring rise in a circular basin beneath a high bluff of Gasconade Dolomite which is above the Eminence Dolomite, the main water-bearing unit for the spring. The Eminence is a gray, medium- to coarsely-crystalline gray dolomite and contains very little chert. The unit hosts most of the 10 largest springs in the Ozark

region. To date, the submerged channel of spring has been explored out to 3,800 ft in length and to a depth of 280 ft. The seven-minute video entitled “The Hidden World Beneath Maramec Spring Park,” is on display inside the Maramec Spring Park Museum (OCDA, 2023).



Figure 3.1.24 Maramec Spring, boiling up beneath a high bluff of Gasconade Dolomite. Courtesy of Maramec Spring Park. Photo by Tammie Leigh.

Three Ordovician-age bedrock formations are exposed in the Maramec Spring area. The Gasconade Dolomite, from which the spring discharges, is about 350 ft thick, and is the basal Ordovician bedrock unit. It overlies the Eminence Dolomite and is exposed at Maramec Spring as well as along the parts of the valleys of Dry Fork Norman Creek, and the Meramec River. An artificial lake has been created by the damming of the spring branch downstream from the spring orifice. The dam, which raises the water level 8 to 10 feet, was built in the early 1800's to provide waterpower for a charcoal iron-making industry (Vineyard and Feder, 1982).

As explained by Vineyard and Feder (1982), Maramec Spring played an important role in the early development of Missouri. Near the great spring were deposits of the red iron mineral hematite, brown limonite, and iron pyrites. Thomas James, an ironmonger from Ohio, came to Missouri in 1826 and began to build the Maramec Iron Works at Maramec Spring. The spring was used for waterpower for the Maramec Works, the largest and sometimes the only local supplier of iron products in the Missouri economy. Maramec iron was shipped to markets across Missouri over primitive "iron roads" that later became Missouri's highway system.

In 1843 Thomas James sent his son, William, to manage the Maramec Iron Works with furnaces powered by the rushing spring water. He remained there until operations ceased and the Iron Works closed in 1876. At its peak, the company town had an estimated population of over 500.

Following the death of William James in 1912, his granddaughter, Lucy Wortham James, acquired ownership of Maramec Spring along with the forest and farmlands surrounding it in 1920. Upon her death in 1938, she made her residuary estate a part of a Trust and authorized thencreation of the James Foundation. To her executors, Lucy Wortham James wrote, “As this is considered to be the most beautiful spot in Missouri, it is my great hope that you will arrange that it may ever be in private, considerate control, and ever open to the enjoyment of the people.”

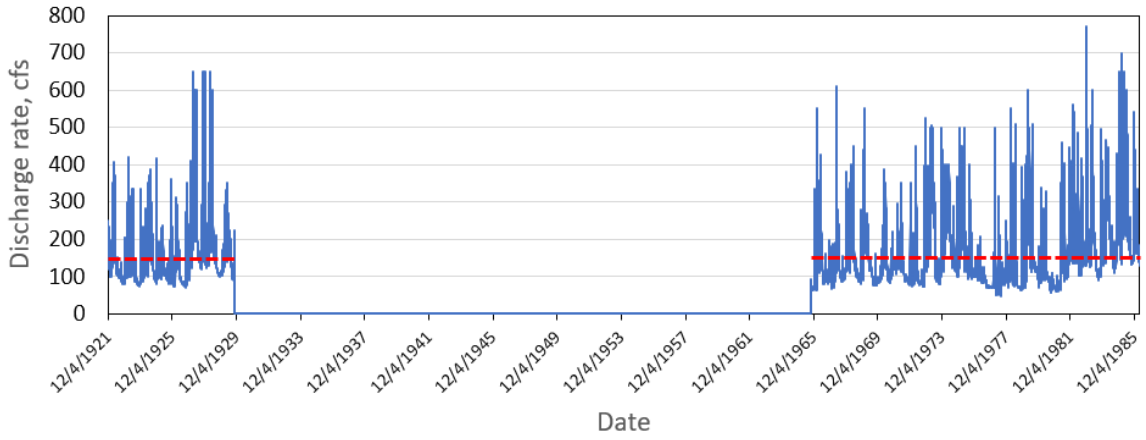


Figure 3.1.25 Hydrographs of the Maramec Spring discharge rate recorded at USGS gaging station 07010500 Maramec Spring near St. James, Missouri. Average for the first period 12/1921 to 10/1929 is 150 cfs; for the second period 10/1965 to 3/1986 it is 156 cfs. For the entire period of record the average discharge is 153 cfs.



Figure 3.1.26 Outlets from the Maramec Spring basin. *Left*: water captured for the state hatchery. *Right*: main natural outlet. Panoramic photo taken in June 2025 during high flow conditions.

Blue Spring, Shannon County, Missouri

Blue Spring is located 14 miles east of Eminence or 14 miles west of Ellington off of State Route 106. A long, steep dirt road leads to the parking area from the highway. It is not recommended for large RVs or buses, although in most weather a normal passenger car should not have trouble making it down. From the parking lot it is about 1/4 mile walk down a pleasant trail.

Blue Spring rises quietly beneath a bluff of Eminence Dolomite from a very deep cave shaft and flows swiftly toward the Current River about 1/4-mile away. The spring's average discharge rate is 108 cfs (70 MGD) based on limited record (see Table 3.1.1). Schumacher (2008) summarized data for 115 discharge measurements made by the USGS from 1923 through 2006, including some measurements that had not previously been compiled. The discharge measurements ranged from 59.3 to 290 ft^3/s with a mean of 127 ft^3/s.

The Blue Spring Natural Area is managed by the Missouri Department of Conservation, while the Current River and surrounding area are administered by the National Park Service, as part of Ozark National Scenic Riverways.



Figure 3.1.27 Blue Spring in Shannon County, the gem of Ozark National Scenic Riverways. The mesmerizing deep blue color is characteristic of periods without significant rain and average- to below-average flow conditions. October 2023.

Many divers, notably members of the St. Louis Underwater Recovery Team and more recently of The Ozark Cave Diving Alliance (OCDA), have explored Blue Spring. The tubular orifice of the spring plunges at a steep angle beneath the bluff. It is this steep angle and extremely deep water that heightens the deep blue color of the spring. The maximum depth reached by divers (currently 278 feet) may not represent the total depth of the spring; divers report that only the technical problems of descending to such depths prevented further exploration of the conduit system that feeds Blue Spring. A map of progressive diving attempts is shown in Fig. 3.1.28 and can be found at the web site of OCDA together with the accompanying narrative (note that dive access is limited to those who obtain a research permit) https://www.ocda.org/exploration/projects/photo-of-the-month/.

The following description of the Spring is from Missouri Department of Conservation, 2023c: "*A large, beautiful, undisturbed spring and spring branch with associated aquatic plants and animals surrounded by forest in the Current River Hills region of the Ozarks. The Osage Indians reportedly called this spring "Spring of the Summer Sky" ("Do-Ge-Ke-Thabo-Bthi"). Spring water is actively dissolving away limestone and or dolomite as it moves through the earth. Springs are actually excavating new caves through this process. This dissolved limestone and or dolomite, along with the influence of the spring's depth and the blue of the sky, impart the blue color of the spring. The recharge area for the spring includes the headwaters of Logan Creek which is nearly 10 miles away. This part of the recharge area lies in the topographic watershed of the Black River despite the fact that the spring itself is feeding the Current River. In the spring pool star duckweed, American bur-reed, and water starwort occur. Giant cane, grass-of-Parnassus, and blue boneset grow along the banks of the spring run. Southern redbelly dace, Ozark sculpin, and rainbow darter inhabit the spring run. The rare Swainson's warbler has been found in giant cane stands along the Current River just upstream from the spring run. Other birds to look for include American redstart, northern parula, kingfisher, Kentucky warbler, red-shouldered hawk, ovenbird, wood thrush, and yellow-throated warbler.*

The surrounding area was used for a lodge and retreat until 1960 when it was sold to the Conservation Department. Just upstream on the Current River is Owls Bend. This area historically supported a mill that produced gunpowder and a river ferry which was the only way across the Current River here until 1975."

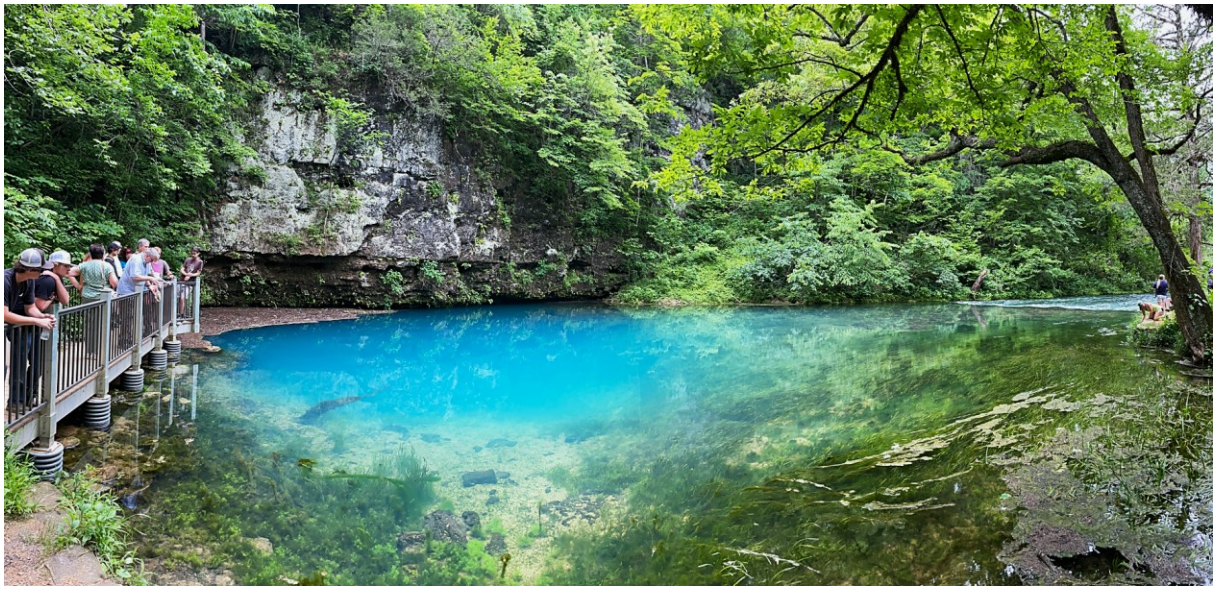


Figure 3.1.28 Blue Spring during high flow conditions in June 2025.

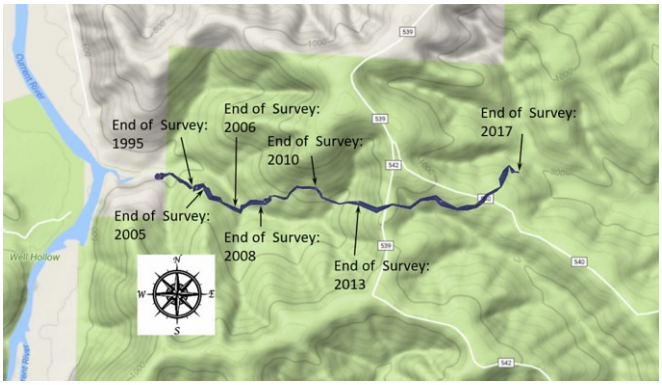


Figure 3.1.29 Mapped submerged passages of the Blue Spring cave. Courtesy of The Ozark Cave Diving Alliance; available at https://www.ocda.org/wp-content/uploads/2021/10/Blue-Spring-Progression.png

Alley Spring, Shannon County, Missouri

Alley Spring is located six miles west of Eminence, Missouri, on State Route 106. The grounds of the famous red Alley Mill powered by the Spring and the wider Spring area are part of the Ozark National Scenic Riverways operated by the National Park Service. The spring has an average daily flow of 134 cfs or 87 million gallons. Most often the water is a placid azure blue, gently welling up from below, but after a storm has swelled the underground conduits feeding the spring, it can gush forth in angry swells and splashes of brown.

Alley Spring discharges at the foot of a picturesque limestone bluff of Eminence Dolomite and flows past the Alley Mill before discharging into Jacks Fork, about 1/2-mile away.

Alley was home, farm, and school for people who lived there a century ago. Dances, baseball games, and roller skating were all part of Alley's busier days. John Knotts purchased the 80-acre site in 1902 and diversified the enterprises to include a well-stocked store and blacksmith shop. The mill was vital to community life, where grain was ground to provide the daily bread. It was also used for grinding cattle feed, operating a sawmill, and furnishing electrical power for nearby houses. The present building was constructed during 1893-1894 by George Washington McCaskill as a merchant mill. It was larger than most mills in the Jacks Fork area and replaced an earlier mill on this same site that was built by 1868. Originally unpainted, it was first painted white with green trim, then later the famous red color associated with Alley Mill today.



Figure 3.1.30 Beautifull red Alley Mill located on the Alley Spring basin's bank utilized the perennial flow of water to operate its turbine-powered flour mills. A control wheel on the porch allowed the miller to control how much water entered the turbine and thereby control its speed.

In 2001 the United States Geological Survey requested the assistance of The Ozark Cave Diving Alliance (OCDA) to obtain scientific data from within Alley Spring. To date the team has surveyed/mapped several thousand feet of passage, collected rock and sediment samples, and taken video and photographs within the system. Just under 3000 feet of the system has been surveyed with an average depth of 155 feet. This system is normally affected by poor visibility and high flow. A map is on display inside the Red Mill that was created by the survey information gathered by the OCDA (OCDA, 2023).



Figure 3.1.31 Discharge basin of Alley Spring at the base of the bluff in Eminence Dolomite. Photo courtesy of USGS.

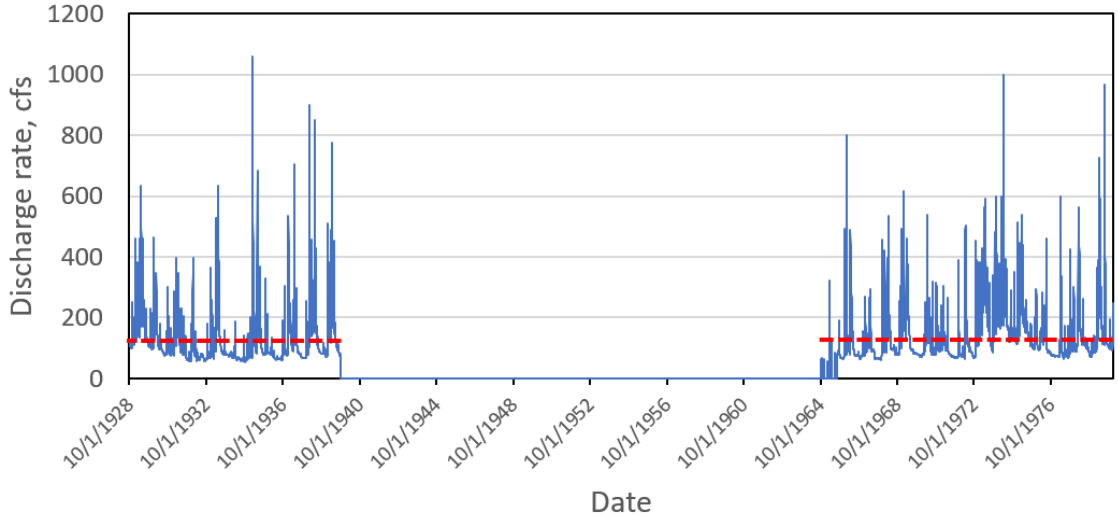


Figure 3.1.32 Hydrograph of the Alley Spring discharge rate recorded at the USGS gaging station 0706550 Alley Spring at Alley, Missouri. Average for the first period 10/1928 to 9/1939 is 124 cfs; for the second period 8/1965 to 12/1979 it is 141 cfs. For the entire period of record the average discharge is 134 cfs.

An interesting note is provided by Vineyard and Feder (1982) illustrating a textbook example of the mature classic karst nature of the Ozark aquifer: “Bridge (1930) reported that the spring was once observed to cease flowing for about 12 hours. The water level decreased rapidly and sank to about 5 feet below normal pool level. After about 12 hours the water level rose and flow resumed, but the spring was quite muddy for several days thereafter. At approximately the same time a large sink formed in the upland about 15 miles northwest of the spring. Bridge does not record the date of the temporary interruption of the flow, nor the location of the sink, but it was generally supposed that the problem was caused by temporary blockage of the spring supply channel by the sinkhole collapse.”

As described by the Missouri Department of Conservation (2023d), despite its historic use to power a mill, Alley Spring has retained its biological integrity. The cool waters issuing forth from Alley Spring flow through a spring branch for a half mile before entering the Jacks Fork River. The spring branch cool water (58 degrees Fahrenheit) provides habitat for colorful Ozark fishes including the southern redbelly dace, the Ozark sculpin, and the bleeding shiner.

On the dry rocky ridges of this natural area are some of the highest quality old growth stands of white oak and shortleaf pine woodland known in the Ozarks. These stands were spared the heavy, indiscriminate timber cutting of the Ozarks that occurred from 1880-1920. Looking up at the sentinel pines found on the ridges one can get a glimpse of what the six million acres of shortleaf pine woodlands that were in Missouri in 1860 looked like. Today only about 600,000 acres of shortleaf pine remain in the state. At one time the nation's largest sawmill was at Grandin, Missouri, where lumber production peaked at the beginning of the 20th century. In 1910 the mill was moved to West Eminence, just 4 miles from Alley, but by 1920 the woods were cut over and the mill was sold.

Also protected is Branson Cave, one of the most biologically diverse caves known in Missouri. The natural area is home to five species of conservation concern that depend on spring and cave natural communities. A wide spectrum of Ozark natural communities typical of the Current River Hills region can be explored from springs and associated mesic upland forests to cherty dry woodlands and dolomite glades. Visitors can see plants such as wild hydrangea and walking fern near the spring and then ascend the hill to see shortleaf pine and low bush blueberry on dry cherty slopes. Birds to look for in the woods include ovenbird, summer tanager, red-eyed vireo, and black-and-white warbler. Deep below the trails are cave passageways which support at least seven cave-adapted animals or "troglobites" including the grotto salamander, a species of conservation concern.



Figure 3.1.33 Intensely karstified dolomite on a trail around Alley Spring.

Rainbow Spring, and North Fork Springs, Ozark County, Missouri

Rainbow Spring, also known as Double Spring, is located 30 miles west of West Plains, Missouri and accessible via Route CC (west), State Highway 181 (south), and County Road 374 (east). It boils up around large blocks of dolomite and sandstone at the western base of a wooded bluff on the right bank of the North Fork River Spring and is in the uppermost Gasconade Dolomite. Ledges of sandstone of the Roubidoux Formation are exposed

in the bluffs upstream and downstream from the spring, and in the bed of the North Fork at North Fork Springs (Vineyard and Feder, 1982). The water bursts from the base of a wooded bluff with a boil that rises one to two feet above the surface in a rocky grotto green with ferns, mosses, and watercress.

The outflow from Rainbow Spring divides into two branches as it leaves the pool, one leading north and the other south, forming a long island between the spring branch and the North Fork. This is the reason why Rainbow Springs originally was called Double Spring. The south branch carries 70 to 80 percent of the flow. Dams are built across both ends of the island to connect it with the mainland.

Rainbow Spring has been measured 27 times between 1919 and 1966, with the average discharge rate of 127 cfs. The maximum flow of 232 cfs was measured in April 1965 and the minimum flow of 47 cfs was measured in November 1964 (Vineyard and Feder, 1982; Table 13, page 128). It is therefore not entirely clear why Vineyard and Feder give an estimated value of 100 million gallons per day (155 cfs) for the average flow of the spring in Table 1 of the same publication.

The spring, and thousands of wooded acres around it, is privately owned and access is restricted. It is still the best-kept secret among Missouri's natural wonders



Figure 3.1.34 Rainbow Spring (formerly better known as Double Spring) has an estimated average discharge rate of 127 cfs based on 27 measurements between 1919 and 1966. Left photo Courtesy of Missouri Department of Natural Resources; available at https://dnr.mo.gov/land-geology/geology/karst-issouri/springs. Right photo by W.H. Pohl, from Beckman and Hinchey, 1944.

North Fork Springs are about 750 to 1,000 feet upstream from Rainbow Spring. They consist of numerous openings on the left bank of the river, in the bed of the river, and in the gravel bar toward the right bank. The openings on the left bank and in the bed of the river are along joints and those in the river form lines of boils extending from the shore to the gravel bar; they are aligned with the joint system in the bank and bed of the river. The rises comprising North Fork Springs extend about 100 feet along the river. North Fork Springs were first measured in November 1964 and three measurements of the spring flow have been made since that time. North Fork River flows over a bedrock floor in much of the reach from North Fork Spring to Rainbow Spring.

As noted by Vineyard and Feder (1982), it is difficult to explain the correlative measurements of the outflow of Rainbow and North Fork Springs. Listed below are these measurements of the two springs. If the two springs are hydraulically connected, there should be closer correlation between the outflows of the two springs. It is suggested that if the two are hydraulically connected, North Fork Springs should have a constricted outlet opening allowing little variation in flow. Rainbow Spring, on the other hand, can discharge the greater volumes of water

available in the spring supply system during the winter and spring months. The similarity of the water from North Fork and Double Springs also suggests a common supply system. In April 1966 the springs were sampled, and the results show that their water is chemically almost identical (Vineyard and Feder, 1982).

	Nov 16, 1964	April 8, 1966	July 6, 1966	Oct 6, 1966
North Fork Springs	75.9 cfs	66 cfs	75.3 cfs	68.4 cfs
Rainbow Spring	47 cfs	180 cfs	150 cfs	132 cfs



Figure 3.1.35 Screenshot of the Missouri DNR GIS database with satellite image of the area of Rainbow and North Fork Springs in Missouri. Esri, NASA, NGA, USGS, FEMA|Esri Community Maps Contributors, Ozark County, Missouri Dept. of Conservation, Missouri DNR, © OpenStreetMap, Microsoft, Esri, HERE, Garmin, SafeGraph, GeoTechnologies, Inc, METI/NASA, USGS, EPA, NPS, US Census Bureau, USDA.

Welch Spring, Shannon County, Missouri

Welch Spring is located approximately 2 miles north of Akers on Highway K. The spring flows from the base of a wooded dolomite hill near the historic Welch Hospital, and then flows to the Current River over a rock impoundment. Welch Spring is part of Ozark National Scenic Riverways.

Vineyard and Feder (1982) estimated that the spring's annual mean discharge was 116 cfs. Aley (1976) estimated an annual mean discharge of 140 cfs, whereas Keller (2000) estimated an annual mean discharge of 229 cfs and suggested that a flood in 1985 changed the character of the spring channel such that some of the flow that had gone through the alluvial gravel before 1985 could now be measured, resulting in larger discharge measurements. This would rank Welch Spring as the third largest spring in Missouri and the second largest spring in the Current River Basin.

The Ozark Cave Diving Alliance (OCDA) was given the opportunity to explore Welch Spring in 2007. Inside the gated entrance is an underground lake 10-18 ft deep. There is a dry passage which extends into the bluff from

this lake area which has been mapped. The spring then descends to 149 ft and gently ascends to the 100 ft area at the current end of the line. The rather poor visibility in this system is typical of Missouri springs and varies with weather. Exploration of this system will hopefully continue in the future (https://www.ocda.org/exploration/projects/welch-spring/).

The area around Welch Spring was settled in 1855 by Thomas Welch whose family ran a gristmill for fifty years after the Civil War. The spring and 40 acres were sold in 1913 to Dr. Chrisian Diehl of Roxana, Illinois, who spent the next 30 years alternately practicing medicine in Illinois in the winter and building and promoting a sanitarium for asthma sufferers in the spring.

As described by NPS, “Dr Diehl believed that the spring water had healing properties and that cool, pollen free air coming from the adjacent cave would be beneficial for people with asthma, emphysema, and tuberculosis, which together were called "consumption" at the time. He said that it worked for him, helping him with a chronic case of hay fever. To tap this clean air resource, Dr Diehl built a hospital over the mouth of the cave. Welch Spring, which flowed from the cave, was dammed up so that water would close off the entrance. This was to force more air out through the cave opening into the hospital. In today's terms, it might be better called a "health spa" since there wasn't much in the way of formal medical treatment, just an invitation to breathe the fresh air of the cave.

Dr Diehl was not blind to the scenic values of the region either. He hoped to run a thriving campground resort to supplement his medical fees with tourist dollars. In time his healing resort expanded to a few small cabins, a campground, a show cave, and he even had an electric generator running off the spring. Visitors came from the local area and from as far off as Oklahoma and Illinois, but times were hard and travel to such remote places still difficult.



Figure 3.1.36 Welch Spring in Ozark National Scenic Riverways. The ruin of Welch Hospital is visible to the left of the bluff. Also visible are patches of watercress (Nasturtium officinale), one of the oldest known leaf vegetables consumed by humans; it was a favorite of Native Americans. Photo taken in June 2025 during high flow conditions.

Unfortunately, the hospital and resort were not a big success. Roads in the Ozarks were rough and unpaved, making it difficult to get into the Current River Country. Few tourists were willing to make the trip. The good doctor died in 1940, and his family did not have much interest in keeping up the resort, and it soon fell into ruin.

Boiling Spring, Pulaski County, Missouri

As described by Vineyard and Feder (1982), Boiling Spring near Powellvile, Missouri on interstate highway 44, is the largest in the Gasconade basin. It rises in the channel of the Gasconade River near the right bank, beneath a high bluff of Ordovician Gasconade Dolomite. On the surface of the river the spring appears as a very pronounced boil whose height above the surface depends upon the discharge of the spring and that of the river. Unlike the water in some resurgences, where flow lost upstream reappears downstream, the water in this spring has a bright blue color and its source is in the upland areas east and south of the spring. It is reported that it has been turbid only once in 30 years (Bretz, 1956, p. 412). At low river stage, when the bottom of the pool can be seen, it is filled with gravel continually churned by the rising current.

The Missouri Department of Natural Resources (DNR) lists the discharge rate of Boling Spring as 105 cfs (68 MGD) on one of their web sites where there is a GIS-based visual story titled “Follow the Water” (Table 3.1.1). Vineyard and Feder (1982) provide the same number as an estimate but also list the maximum measured discharge rate of only 45.3 MGD (70.2 cfs; Table 9, page 92). Incidentally, the official Missouri DNR GIS web site for all springs in Missouri lists the minimum and maximum discharge rates as 25,133 and 31,506 gallons per minute respectively (56 cfs and 70 cfs respectively). Because of these discrepancies, Boling Spring is not shown as a first magnitude spring on the map in Figure 1.3.8 of this book.

Years ago, the spring was used as a water supply for a bluff-edge lodge and cabins above the spring. A cable was stretched from the bluff to a point above the spring, and a bucket could be lowered into the spring and water drawn upward to the lodge (Vineyard and Feder, 1982). The land access to the spring is owned by Missouri Department of Conservation (MDC). On the other side of the river is a nearby Boiling Spring Campground (18700 Cliff Road, Dixon) from where kayakers often visit the spring today.



Figure 3.1.37 Powerful boil of Boiling Spring in the Gasconade River as seen from a kayak. Photograph by Sterling Adams. Available at https://www.facebook.com/sterling.adams

According to information and analyses presented in Master Theses by students at the Missouri University of Science and Technology in Rolla (Kaufmann, 2002; Montes Chamorro, 2024), dye-tracing has confirmed that Boiling Spring is part of a complex, interconnected karst network that includes Wilkins Spring, Onyx Spring, and Yelton Spring. Using a technique first derived in 1935 but expanded with additional data, the discharge of Boiling Spring has been determined to be closer to 130 cubic feet per second, making it a large first magnitude spring. The recharge area, which was verified to be an elongated south to southeastern region, has a complex relationship with Wilkins and Yelton Springs on Mill Creek and Relfe (Coppedge) Spring on Spring Creek. The recharge areas for these and other springs apparently overlap and form a tiered system with Boiling Spring being the lowest base level spring for the region.



Figure 3.1.38 View of Boiling Spring discharging in Gasconade River. From Montes Chamorro, 2024.

Roubidoux Spring, Pulaski County, Missouri

Roubidoux Spring (in the past also known as Big Spring and Waynesville Spring) flows from a rocky basin at the base of a high bluff of Gasconade Dolomite along Roubidoux Creek within the city limits in the southern part of Waynesville, Missouri. A county road passes between the spring and the bluff and is protected by a concrete retaining wall which prevents undermining of the road by the spring during periods of high discharge. The spring is easily reached from Interstate 44, 3/4 mile south of the spring, and from Business Route 66 north of the spring.

When Roubidoux Creek is in flood, the spring may be covered with 8 to 10 feet of water, but even when this occurs the discharge of the spring increases accordingly and is sufficient to produce a mound or boil of water that rises above the level of flood waters in Roubidoux Creek.

Rubidoux Spring has never been measured in periods of peak flow. During high flow, spring water is usually much clearer than Roubidoux Creek water. It is interesting to note that the maximum measured flow of Roubidoux Spring is greater than that of Boiling Spring in Pulaski County, but the minimum flow of Roubidoux Spring is far less than the lowest measurement of Boiling Spring. Based on 14 measurements from 1924 to 1970, the average discharge rate is 58.3 cfs; the minimum and maximum are 4.7 cfs and 192 cfs respectively (Vineyard and Feder (1982).

Recharge for Roubidoux Spring comes from the dry valley of Roubidoux Creek extending south and west for approximately 10 miles from Roubidoux Spring. Permanent flow begins in Roubidoux Creek about 1 mile southeast

of Roubidoux Spring. From this point toward the headwaters of the creek, Roubidoux Creek sinks into a wide gravel bed, and the stream is dry most of the year for approximately 15 river miles (Vineyard and Feder (1982).

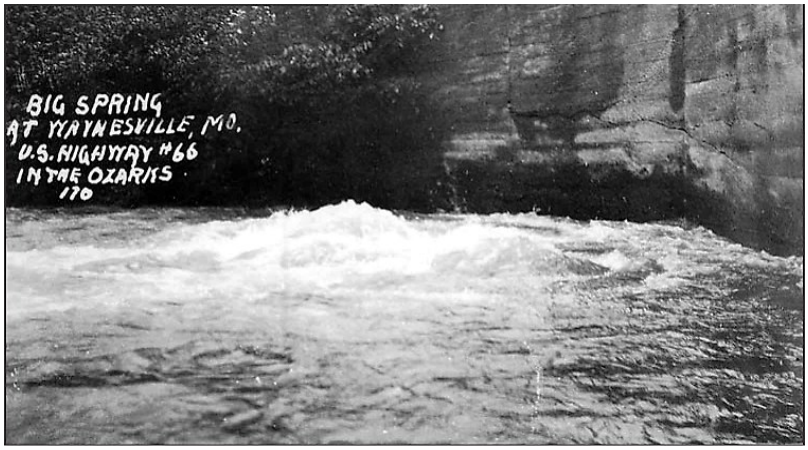


Figure 3.1.39 Historic photograph of Roubidoux Spring. During periods of high discharge, the flow coming from the underwater opening can boil three feet above the surface of the pool. Courtesy of Jan and Terry Primas. From "The Big Spring" by Terry Primas, Old Settler Gazette, 2008.

"The spring lies at the south end of Roy Laughlin Park on City of Waynesville property. It hasn't always been so. In the early part of the 20th century, the spring belonged to Frank O. Gray. In October, 1909, J. W. Wheeler, Administrator of the estate of Frank O. Gray, deceased, posted a notice in the *Democrat* that all of Gray's belongings, including a parcel of land containing the Big Spring, would be auctioned off on the south steps of the Pulaski County Courthouse. The *Democrat* reported on the impending sale, noting with some vision that "This is a remarkable piece of property and the citizens realizing the fact that it should belong to the town, have raised a fund to buy the tract of land on which it is situated. It may prove to be of great value to the city." (From *The Big Spring* by Terry Primas, Old Settler Gazette, 2008, p. 20-21.)

During the 1837-1839 Trail of Tears forced relocation of the Cherokee Nation, the area was used as an encampment along the 2,200-mile-long route from the southeastern part of the country to Oklahoma. The city of Waynesville erected memorial boards along the Roubidoux Walking Trail in Laughlin Park along with a Wayside Exhibit that can be found on the bank of the Roubidoux.

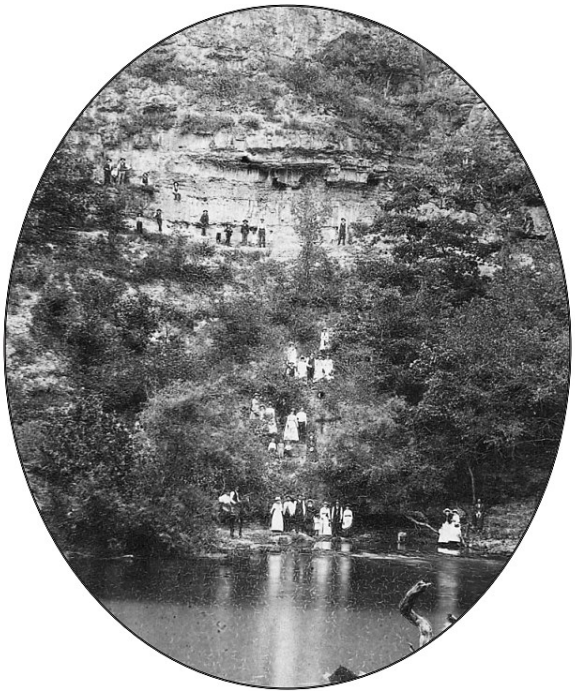


Figure 3.1.40 very early photograph of the Roubidoux Spring (known as Big Spring at the time) with citizens in their Sunday finery stairstepped up the bluff. Courtesy of Pulaski County Historical Society.

Waynesville was home to the 2015 National Speleological Society (NSS) Convention which showcased the Roubidoux Spring cave along with other cave systems in Missouri. For over 60 years, divers have braved the dark passages to a depth of 350+ feet. Depending on the rate of flow and temperature, which averages between 55-60° Fahrenheit, most divers experience a beautiful and breathtaking experience. As reported by The Ozark Cave Diving Alliance, with improved conditions and some real long weekends, the system has now been extended to 11,528 feet at a relatively modest depth of 145 feet.

The following text by Terry Primas, which accompanies the photograph in Figure 3.1.40, attests to the wise decision of the City to acquire the property: “Ninety-eight degrees in the shade and where to have a picnic? The Big Spring in Waynesville was a favorite public place on a sweltering day for local town dwellers. Folks from neighboring villages came to enjoy a stream-side lunch, too. The Big Spring (also known as Waynesville and Roubidoux Spring) boils out of an underwater cave with a water temperature of 58 degrees. The cool temperature moderates the heat around the spring.”



Figure 3.1.41 1960’s era photograph of Roubidoux Spring. Courtesy of Jo Schaper. Available at https://www.ocda.org/exploration/projects/roubidoux-spring/

Montauk Springs, Dent County, Missouri

Montauk Springs is in the Montauk Springs State Park, established in 1926, 15 miles southwest of Salem, Missouri via state roads 39 and 19. Montauk Springs was a single spring with a bedrock opening that became filled with gravel following a flood in the early twentieth century, causing the spring discharge to rise in several pools, gravel bars, and creek beds in the flood-plain alluvium of Pigeon Creek (Vandike, 1997). Visitors looking into the water from the banks of Pigeon Creek can see sand boils created by high-pressure water discharging out of small underwater caverns, pushing sand aside and creating the visual illusion of "boiling" sand.

Montauk Springs is the eleventh largest spring in Missouri, and the fifth largest spring in the Current River Basin (Vineyard and Feder, 1982). The annual mean discharge for the 3 years the USGS maintained a gage at Montauk Springs was 58 cfs in 1965, 73 cfs in 1967, and 86 cfs in 1968, but no longer term annual mean discharge value is available. The average of 28 discharge measurements made from 1980 through 1994 was 87.6 cfs (Vandike, 1997). Aley (1976) estimated the annual mean discharge from Montauk Springs of 100 cfs.

A dam on the southwest branch submerges the springs which feed it and forms Spring Lake, which is used to supply water to fish rearing ponds. The northeast branch is comprised of the flow from a number of spring outlets,

usually about seven in all, which emerge from gravel beds or from rocks at the base of a hill of Lower Gasconade dolomite. The flow of this latter branch is augmented by the smaller flow from Pigeon Creek above. The two branches unite near the park headquarters to form the headwaters of Current River.

The spring outlets are inconspicuous if compared with those of some other large springs in the State, but their combined flow (constituting the headwaters of Current River) forms an excellent stream for trout fishing. Montauk State Park is best known for being one of the three "trout parks" in the Missouri state park system. The Missouri Department of Conservation manages and operates the trout hatchery and offers tours as well.



Figure 3.1.42 Main Spring at Montauk Springs State Park. Photo by Sony Hocklander, Springfield Daily Citizen. Available at Why Montauk State Park is the perfect family getaway, whether you fish or not - Springfield Daily Citizen

Hahatonka Spring, Camden County, Missouri

Hahatonka Spring is in Ha Ha Tonka State Park which is located about 6 miles south of Camdenton, Missouri, on State Road D. With an average discharge of about 74 cfs it is the largest spring in Camden County. The spring discharges from a phreatic cave developed in the upper part of the Eminence Dolomite. Lower Gasconade Dolomite and the Gunter Sandstone member crop out in the valley walls around the spring branch. The spring rises at the head of a narrow, deep valley that is likely developed by collapse rather than by surface erosion (Vandike, 1992).

A bedrock island, containing several caves and heavily weathered bedrock, divides the spring branch a few hundred feet downstream of the spring. Beyond the island, spring flow enters the Niangua arm of Lake of the Ozarks.

Hahatonka Spring is one of many karst features occurring in the immediate area. Several major sinkholes, one containing a large natural bridge, lie within a few hundred yards east of the spring. River Cave, which pirates flow from surface drainage and channels it into the Hahatonka Spring conduit system, is 2,000 feet to the northeast.

Divers entering the spring have made the underwater connection with River Cave (Porter, 1990; personal communication; from Vandike, 1992). The abundance of solution features in the area is of interest because of the many indications of strong structural disturbance. Faulting is prominent in the area. Certain stream alinements and the offsetting of other streams suggest additional faulting or intense jointing, much of it still unmapped. Together, these features suggest the development of a large underground storage reservoir that keeps the flow of the spring at a high level (Vineyard and Feder, 1974).



Figure 3.1.43 Hahatonka Spring in Ha Ha Tonka State Park. Photo courtesy of Missouri State Parks

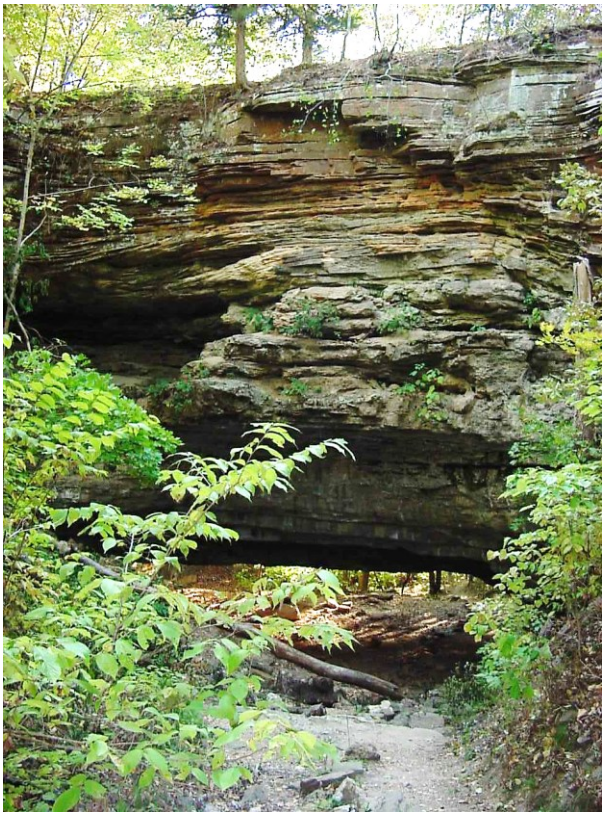


Figure 3.1.44 Natural bridge in Ha Ha Tonka State Park. Photo courtesy of Missouri State Parks.

Pulltite Spring, Shannon County, Missouri

Pulltite Spring is located close to Pulltite campground, about 14 miles north of Eminence, Missouri off highway EE. It emerges at the bottom of a cliff in Lower Ordovician Gasconade Dolomite, forming a small pool and branch that flows into the Current River. The spring was named Pulltight for it was a tight pull for the horses as they worked to ascend the steep hill with heavy burden after visiting the Watson Mill. It was hard work both ways. The way down to the mill was so steep that the driver would have to brace the back wheels of the wagon with a chain or tree limb. This prevented the wagon from running over the team of horses. To retain control, he would have to pull tightly on the reins to slowly guide the team and wagon down the hill. Therefore, the area became known as Pulltight. The spelling was shortened recently to "Pulltite. The spring was purchased by St Louis businessmen in 1911 and was run as a fishing resort, until the area was acquired by the National Park Service in 1967 when it became part of the Ozark National Scenic Riverways.

The estimated mean annual flow of the spring is 45.5 cfs, or 29.6 mgd (Aley, 1978). The estimated mean annual flow for Pulltite Spring and the many springs in its immediate area (Pulltite Spring Complex) is 223 cfs (Aley, 1978). Based on the mean annual flow, an approximation of the recharge area can be calculated to be about 223 square miles (Aley and Aley, 1978). However, using dye tracing a more accurate approximation of the recharge area for the Pulltite Spring Complex has been calculated to be about 161.5 square miles (Aley and Aley 1987).

NPS Pulltite Campground is one of the most popular hubs for river recreation in the Ozarks. The spring is located across the campground and can be reached only by boat or by wading across the Current River from the campground.



Figure 3.1.45 Pulltite Spring. Photo available on Facebook page of Ozark National Scenic Riverways. https://www.facebook.com/Ozarkriverways



Figure 3.1.46 Fire Hydrant Cave. Photo courtesy of https://currentriver.net/index.htm

The description below is an electronic reprint excerpt from "Caves of the Current River Valley", Journal of the Missouri Speleological Survey, Year 1982, Volume 22, Numbers 3-4, Page 89.

Fire Hydrant or (more properly) Barrel Spring, is a part of the Pulltite Spring complex. The spring gushes out into the river from a small opening about 3 feet above river level. The effect is of either a fire hydrant or a tipped-over water barrel, hence the names. Fire Hydrant Cave is not the spring orifice. Rather it is on the back side of the hill from which the spring emerges. The water for the spring does flow through the cave and thus the peculiar naming. The entrance is about 5 feet wide and 3 feet high. From the entrance the floor slopes down for 20 feet to a small room. Here the rapid cave stream flows in from the right and noisily splashes through the breakdown in the room. The room is less than 20 feet long and the water exits under a rock shelf to reappear at the spring orifice several hundred feet away. https://currentriver.net/pulltite_cave.htm

Round Spring, Shannon County, Missouri

Located approximately 13 miles north of Eminence on Highway 19 is Round Spring, one of the major attractions of the Ozark National Scenic Riverways (ONSR). The spring rises quietly in a nearly circular basin formed by the collapse of a cavern roof in Upper Cambrian Eminence Dolomite. Part of the roof remains intact as a natural bridge beneath which the waters of the spring flow toward nearby Current River. Boulders from the collapse of the cavern roof effectively block the tubular conduit of the spring, limiting divers to a depth of about 55 feet and preventing their access to the water-filled spring supply channels that are accessible in some other springs. Based on 20+ years of record by the USGS, the average discharge rate of Round Spring is 47 cfs or 30.4 MGD; it is 41 cfs for the first period of record (10/1928 to 9/1939), and for the second period 10/1965 to 12/1979 it is 51 cfs.

According to an NPS pamphlet on Round Spring, it is believed that a portion of the spring's recharge area is to the northeast of Spring Valley, which means that the water would have to flow under the Current River to reach the spring. Round Spring was one of the first parks in the Missouri state park system (1932).

The water temperature in the spring remains between 55- and 58-degrees Fahrenheit year around. Watercress abounds in the cold water, providing shelter for aquatic life including periwinkle snails, insect larvae, and a variety

of fish species. Large number of birds, deer and small mammals visit the spring. People, however, are prohibited from swimming in it.

A short walking distance from the spring is Round Spring Cave, considered the most spectacular cave within Ozark National Scenic Riverways. Large, varied formations fill over a mile of the passage, which has been well-preserved over the years. “There is no better way to learn about Ozark caves than through a tour! In summer, rangers conduct "lantern tours" of the cave, detailing local history and taking questions from visitors. Common topics include karst topography, geology, hydrology, cave wildlife, history, and more. Tours are limited to 15 visitors, allowing for a more intimate experience than typical show caves.” ADVISORY: To help control the spread of White Nose Syndrome, a disease that is killing bats but is harmless to people, please do NOT wear or bring anything into the cave that you have taken into other caves. This includes cameras, belts, footwear, etc. Thank you. (NPS; https://www.nps.gov/thingstodo/round-spring-guided-cave-tour.htm)



Figure 3.1.47 Two views of Round Spring. *Left*: Inside the spring’s collapsed roof section. *Right*: Outlet from the spring. Right photo courtesy of Explore White Plains; https://explorewestplains.com/round-spring/

Cave Spring, Shannon County, Missouri

As described by Vineyard and Feder (1982), Cave Spring issues from a large, near-vertical shaft that leads down to the subterranean conduits in Eminence Dolomite channeling groundwater to the spring orifice. This natural well permits access, with underwater breathing apparatus, to the water-filled supply network of the spring. Access is also available through a large sinkhole called Devils Well and through a small cave known as Wallace Well. The two wells and other similar, but hypothetical, features serve as direct routes and as reservoirs for recharge water entering the spring system. The fortuitous development of Cave Spring and its environs has permitted the observation of a cave system in transition from the deep phreatic zone beneath the water table into the vadose zone above the water table. This transition period is the stage of functioning as a spring. Cave Spring rises at the back of a short, discontinuous cave, from a vertical conduit about 30 feet in diameter and 155 feet deep (Fig. 3.1.53).

Aley (1976) estimated that the annual mean discharge from Cave Spring is 41.5 cfs. Missouri GIS database lists maximum flow of 32,538 gpm (72.5 cfs) and minimum flow of 7,046 gpm (15.7 cfs).



The spring system appears to derive most of its flow from conduits leading in from the north, which is supported by dye tracing results (Mugel et al., 2009).

Figure 3.1.48 Cave Spring discharges from a cave at the base of bluff in Eminence Dolomite and immediately flows into Current River in Ozark National Scenic Riverways (ONSR). Photo by NPS.

While situated within the National Scenic Riverways, Cave Spring itself is owned by the private L-A-D Foundation, which permits public access for viewing from the river https://ladfoundation.org/other-l-a-d-lands/cave-spring/

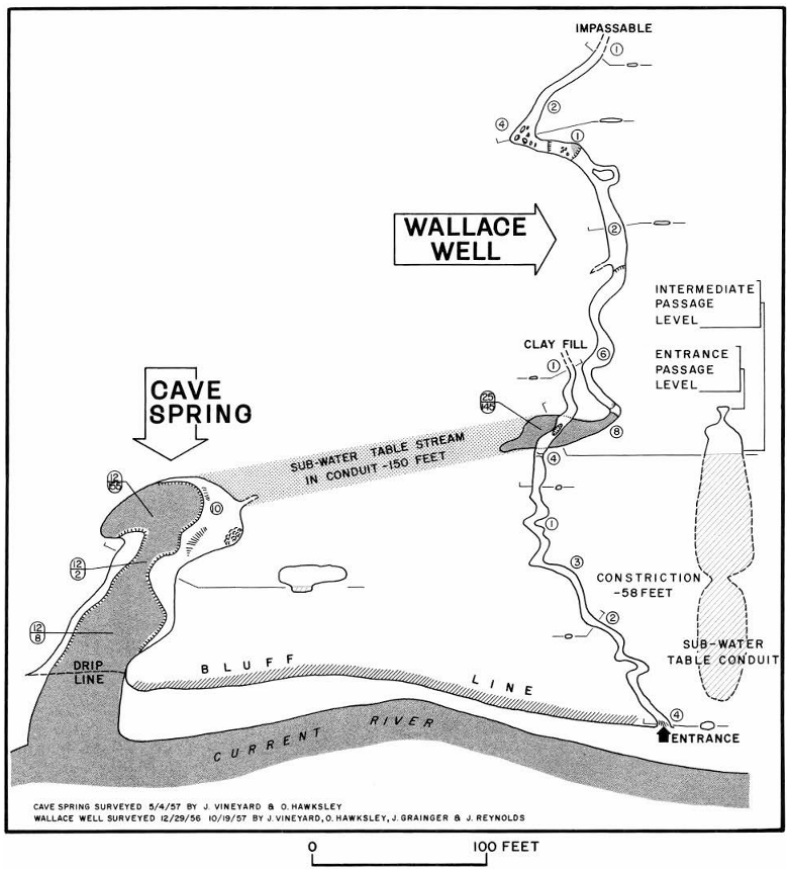


Figure 3.1.49 Map of Cave Spring and Wallace Well. Circled numbers are ceiling heights. Symbols with fraction numbers indicate ceiling height over water depth. From Vineyard and Fedder, 1982; Modified from Vineyard, 1963.
It is not known whether water enters the orifice area from south of the Current River. A passage extends beneath the river (William Cate, pers. comm., Vineyard and Feder, 1982), but it has not been explored by divers.

Devils Well is a large, conical sinkhole about 150 feet in diameter and 40 feet deep which Opens into a large chamber containing an underground lake (Figs. 3.1.50 and 3.1.51). The Gasconade Dolomite crops out in several ledges around the perimeter of the entrance sinkhole. The Gunter Sandstone Member of the Gasconade is exposed in the lower part of the sinkhole and the Eminence Dolomite begins immediately beneath the small rock span at the

base of the sinkhole. The Eminence is exposed for about 100 feet downward to the water table and is accessible to divers for an additional 200 feet. A vertical fault with 2 feet of displacement cuts across the Well, striking due east, displacing the lower part of the Gunter Sandstone and the underlying Eminence Dolomite.

Beneath the entrance slot in the bottom of Devils Well, the walls of an underground chamber diverge to form a bell-shaped profile. This chamber is about 400 feet long and 100 feet wide, with a total vertical range of approximately 300 feet. Water is continually being added to the supply in the lake by four or more waterfalls entering the Well from various points in the ceiling and walls.

Devils Well is one of the largest cave lakes in the nation. More than 400 feet long and reaching depths of over 200 feet, Devils Well is a major feature of the Cave Spring supply system.

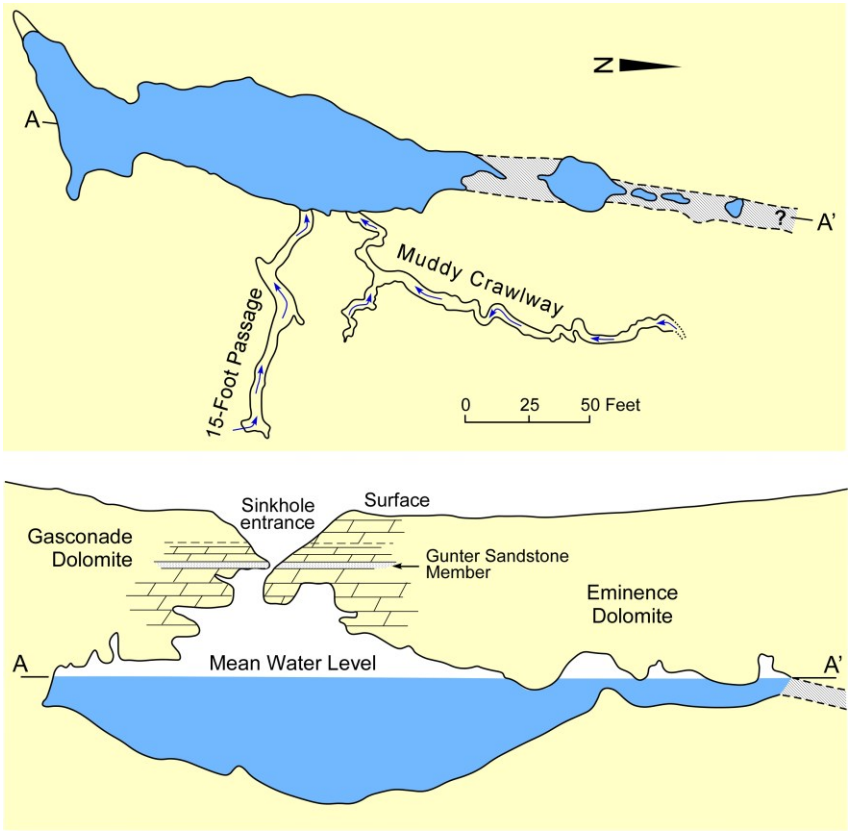


Figure 3.1.50 Map and cross sections of Devils Well, showing lake rooms discovered by divers. Entrance to the Well was by electric winch and bosun's chair through the bottom of a large sinkhole. Original survey after Vineyard (1963), with additions and cross section by Donald N. Rimbach and Michael Tatalovich, Ozark Spring Studies (1973). Modified from Vineyard and Feder, 1982.



Figure 3.1.51 Devils Well. Photo courtesy of Ozark National Scenic Riverways.



Figure 3.1.52 Thomas Hart Benton, one of Missouri's most famous painters, depicted the scene at the entrance to Cave Spring in a painting he completed in 1963. Benton's love for America was celebrated in his landscapes of the country and its people. Leo Drey, Founder of the L-A-D Foundation, purchased this area as he was assembling Pioneer Forest beginning in 1951. In 1970 Leo and Kay Drey donated Cave Spring to the Foundation.

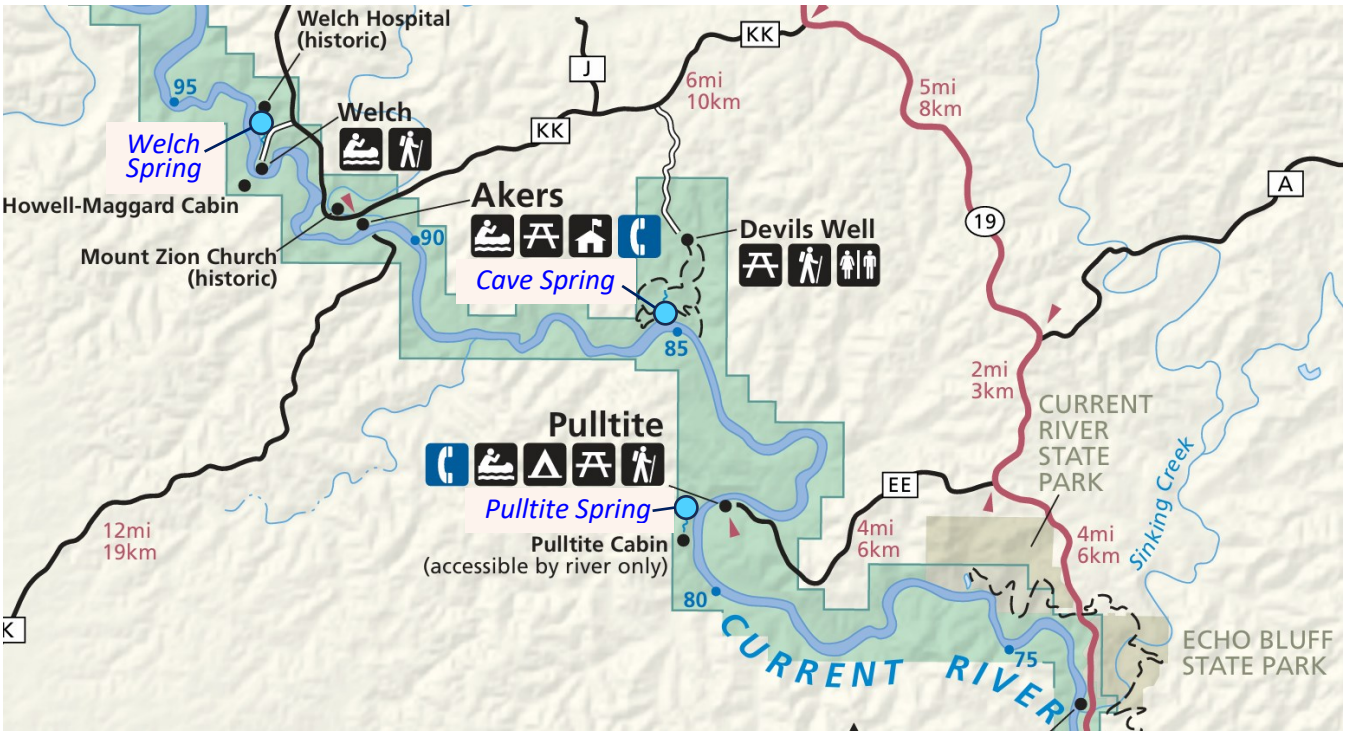


Figure 3.1.53 Part of the NPS map of Ozark National Scenic Riverways showing a 4.6-mile loop trail (dashed black line) that connects Cave Spring and Devils Well. Available at https://www.nps.gov/npgallery/GetAsset/0bc83b25-c1c5-4235-9cec-0b166ccbcb7d

Roaring River Spring, Barry County, Missouri

Roaring River Spring is in the Roaring River State Park, about 10 miles south of Cassville via State Highway 112. The park is surrounded by Mark Twain National Forest in the southwestern corner of Missouri.

Roaring River Spring flows from a pool at the base of a high bluff of Cotter Dolomite in a narrow, canyon-like valley and forms the head of Roaring River. A prominent vertical joint crevice, which has been slightly enlarged by dissolution and which extends upward in the face of the bluff above the spring outlet, suggests that the underground flow at this point may be directed in part by the pattern of jointing in the rocks.

As emphasized by Beckman and Hinchey (1944), a small spring is located about 90 feet directly above the Roaring River Spring. This upper spring flows out of a small, cavernous solution channel at the base of the St. Joe (Mississippian) limestone formation at its contact with underlying Kinderhook (Mississippian) shales.



Figure 3.1.54 Roaring River Spring in Roaring River State Park, Missouri.

Apparently, this small spring has its source in the overlying limestone and cherty limestone strata of Mississippian age and discharges water which has percolated from the surface downward through joints, crevices, and openings. Its further downward movement has been checked by the relatively impervious shales, upon which it has followed a lateral course along the top of the shale beds to emerge in the crevice above the big Roaring River Spring. Thus, the smaller flow is "perched" upon the impervious shale strata.

According to Vineyard and Feder (1982), the water quality data of Roaring River Spring indicate that much of the recharge area is in uplands underlain by Mississippian limestone.

Roaring River Spring has been used for many years as a source of water for one of the large state fish hatcheries. 216 measurements of the spring discharge rate were made between 1923 and 1966 with the average of 31.6 cfs. The maximum discharge was measured on April 6, 1965, when the spring flowed at a rate of 177 cfs. The minimum measured discharge was 8 cfs on October 31, 1963 (Vineyard and Feder, 1982).



Figure 3.1.55 Part of outflow from basins of the State Hatchery fed by the Roaring River Spring.

As of fall 2023, Roaring Springs holds the record for the deepest explored spring in the United States thanks to divers of the KISS (Keep It Super Simple) team (Fig. 3.1.56). The previous record was 462 set in 2013 at Phantom Spring Cave in west Texas. The following excerpts from the article *Down deep at 472 feet* by Flip Putthoff, published on September 13, 2022, in NWA Democrat-Gazette, illustrate this quest.

CASSVILLE, Mo. -- Scores of visitors admire the beautiful flow of turquoise water that emerges from Roaring River Spring at the heart of Roaring River State Park cradled in an Ozarks canyon south of Cassville, Mo.

From a sidewalk at the spring's outflow, it's easy to tell the clear water comes from deep in the Earth, but how deep? That's one fact a team of highly trained cave divers hopes to find out during extremely deep dives into the spring's dark, mysterious and dangerous depths.

These elite divers have reached a depth of 472 feet in the spring, but there's more cave below that, said chief diver Mike Young of Fort Smith. Their dive to 472 feet in November 2021 set the record for the deepest explored spring in the United States. The previous record was 462 set in 2013 at Phantom Spring Cave in west Texas.

Professional divers have explored the spring before -- two in the spring of 1979. A team in the early 1990s reached a depth of 225 feet. At that depth, there's a tight squeeze through a restriction that leads to more of the spring's cave. Bulky conventional scuba tanks prevented those divers from penetrating beyond the restriction.

Rebreather scuba technology the current team uses was developed in the late 1990s. Rebreathers are smaller than conventional scuba air tanks. They work by removing the carbon dioxide from a diver's exhaled breath so the diver rebreathes the oxygen.

The KISS (Keep It Super Simple) brand of rebreathers are manufactured in Fort Smith, Arkansas at Young's business, which builds several rebreather models. Special training is required before a diver can purchase a rebreather.

The smaller rebreathers allowed the divers to go through the restriction. Still, the squeeze is tight, said Randall Purdy of Kearney, Neb., the team's underwater photographer. Gesturing with his arms, he formed an opening about the diameter of a 55-gallon barrel. Wonders never before seen by humans unfolded when the divers emerged. They encountered a colossal water-filled cave room so large it could contain the 550-foot tall Washington Monument, according to Young

The rugged hills and valleys of Roaring River State Park tell a fascinating geologic story about the southwestern Ozarks of Missouri. The park is located on the edge of the Ozark Plateau where it is dissected by the White River. This natural erosion process over thousands of years has created an abrupt change in the landscape, from the flat plains of the plateau to deep valleys and steep hills. Visitors driving to the park are often surprised to find themselves going down and down into the park. The last mile of the drive into the valley drops 400 feet in elevation.

Roaring River is a tributary of the White River, which has cut into the flat plateau, creating deep valleys and revealing an unusual variety of rock formations. As the river cuts through the plateau, it exposes the different layers of sediment, including shale, limestone, dolomite, and chert. Roaring River State Park is one of only a few places in Missouri where visitors can see such a variety of rock formations within such a small area.

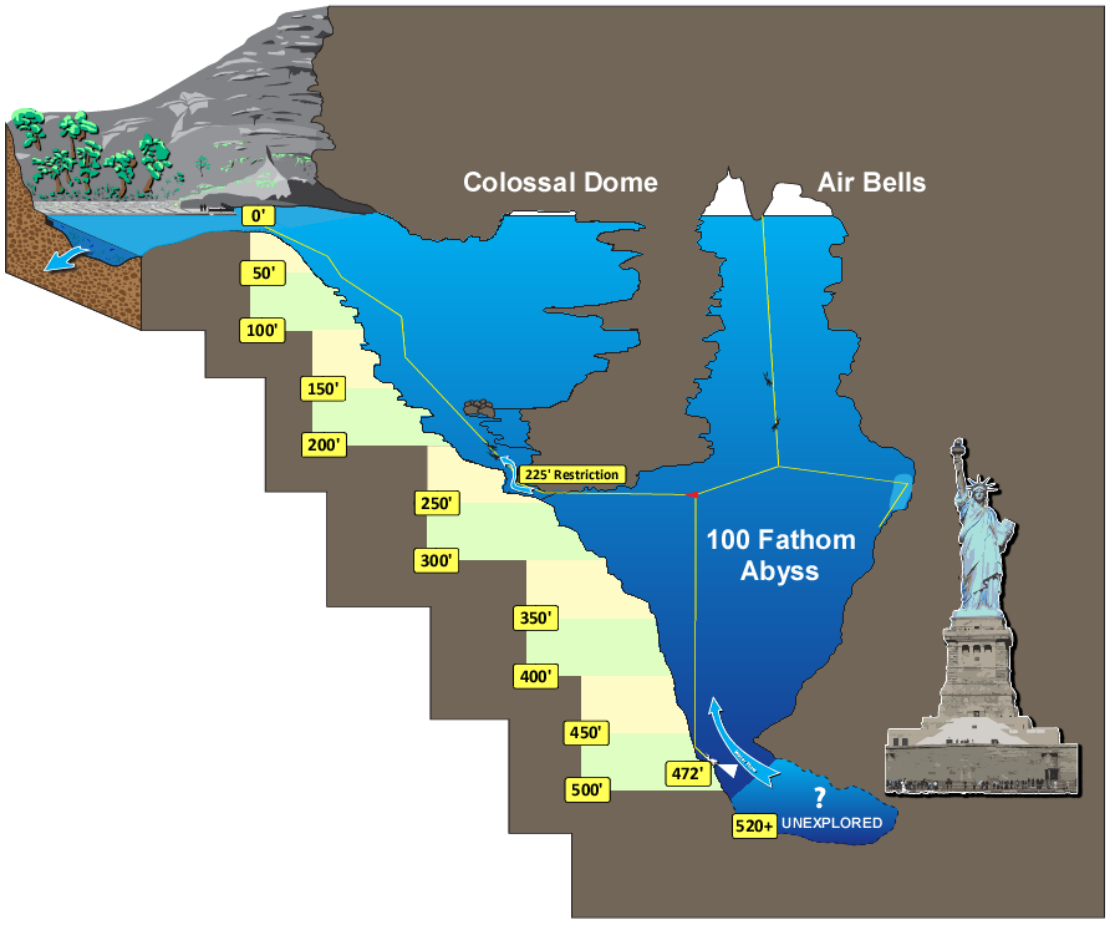


Figure 3.1.56 Profile of Roaring River Spring, at least 520 feet deep, based on the most recent information from the KISS team divers (2021-2023). Participating divers: Gayle Orner, Fernando Gutierrez, Jon Lillestolen, Randall Purdy, Neil Brownlow, Mike Henry, Greg Abels, Charles Walker, Bob Dankert, Eric Hahn, Charles Jay, Mike Young (Chief Diver), Tim Bass, Joseph Heinrichs, Matthew Westpheling. Surface: Colonel (Ret) Tony Bryant, Tom Creason, Adam McDowell, Tana Henry. Moral support: Sheri Young, Jennifer Ables, Angela Brownlow, Paige Purdy. Illustration by Curt Bowen. Courtesy of Mike Young and the KISS Dive Team. www.kissrebreathers.com/

Blue Spring, Carroll County, Arkansas

Blue Spring lies in a large, incised meander of the White River seven miles northeast of Eureka Springs via highway US-62W and County Road 210. This permanent ascending spring discharges from Cotter Dolomite through a near-vertical submerged narrow cave passage which extends deep into the underlying Jefferson City Dolomite of Lower Ordovician. The total depth of 229 feet below the spring pool level was reached by the cave diving team of Mike Young, Kendall Hopkins, and Mike Wright (the video of this cave diving exploration by T.L. Bass productions is available on YouTube). Although the diving results questioned previous information on the spring's depth (560 feet according to Sam Leath's article published in The Ozark Mountaineer in May 1955; or 510 feet as displayed on the information panel by the spring pool), Blue Spring is still one of the deepest and most notable in The Ozarks, with challenging narrow passages and depths yet to be fully explored.



Figure 3.1.57 Google Map of the Blue Spring area. Imagery ©2023 Arkansas GIS, Maxar Technologies, State of Arkansas, USDA/FPAC/GEO, Map data ©2023



Figure 3.1.58 Blue Spring in Blue Spring Heritage Center, Carroll County, Arkansas.

Although the spring is in an incised meander of the White River (Fig. 3.1.57), it apparently has no hydraulic connection with the river: the flow rate and physical characteristics of the spring (temperature is constant at 54 degrees Fahrenheit) are reportedly not affected by the river stage or changes in the river water characteristics.

Officially available period of record for the measured discharge rate of Blue Spring by USGS is 1951-1988. Based on 25 measurements, the average is 5.5 cfs or 3.5 million gallons per day (MGD). The official brochure of the Blue Spring Heritage Center states that the average daily flow of the spring is 38 MGD, without citing source of data. The spring flow originates beyond the White River valley, and the drainage area is unknown but must be of significant extent based on the spring's large average flow rate of 5.5 cfs. Curiously, the information panel at the spring states that "water from Blue Spring originates in the Pacific NW". Staff of the Blue Spring Heritage Center where the spring is located could not reference where this information came from, and were somewhat skeptical when told that, if true, this would constitute one of the most incredible hydrogeologic and natural phenomena known to mankind. Nevertheless, Blue Spring does present a puzzle given that it is surrounded by the White River and the depth of its explored submerged vertical feeder conduit is well over 250 feet below the river level. This means that the water discharging at the spring must flow even deeper beneath the river.

Crystal clear water pours from the Blue Spring circular pool into a trout-filled lagoon formed behind a wooden dam. The lagoon overflows into the White River, "replenishing the area with some of the purest water in the region according to a recent study."



For thousands of years, Native American elders have told stories of visits to Blue Spring and the important ceremonies held in the bluff shelter that served not only as a refuge, but also a sacred place for ritual. Osage Indians claimed the Blue Spring as their trading post. Early settlers nicknamed them "Strongboat Indians" and used their boats to float furs, bear oil, and beeswax down the old trade route of the White River to New Orleans.

Figure 3.1.59 Bluff Shelter at Blue Spring is listed on the National Register of Historic Places.

In March 1839, Blue Spring became a respite and stopover on the Trail of Tears for the Cherokee people during their forced march from Echota, Georgia. They knew of stories long told about the spring and spent time there in hope of healing on a devastating journey. The Cherokee are said to have camped under a rocky overhang called Bluff Shelter for several days on their march west toward Oklahoma.



Figure 3.1.60 Native American arrowheads found on grounds of Blue Spring, on display at the Heritage Museum. In 1971, Robert G. Chenall and his students from the University of Arkansas conducted an archeological dig of the Bluff Shelter next to Blue Spring. They found prehistoric artifacts, shellfish and the bones of deer, turtle, and other fauna. Some date back as far as 8000 BC.

Eureka Springs, Carroll County, Arkansas

Eureka Springs has welcomed visitors for hundreds of years. The legends of several Native American tribes spoke of a Great Healing Spring in the mountains of what later became known as "Arkansas." Newcomers thought this spring to be Basin Spring in the center of their small new settlement (now the area of Basin Park) but were also attracted by the number of other springs, believed to have therapeutic value for many of prevalent diseases. The springs medicinal qualities were “officially” discovered by Dr. Alvah Jackson in 1858. Apparently, Dr. Jackson did not reveal the source of his famous "Dr. Jackson's Eyewater" until 1879. In that year the doctor advised his good friend, Judge L.B. Saunders of Berryville, to try the springs as a final attempt in curing a bad case of erysipelas, a skin disease that caused extended inflammation. The positive results gained from the springs prompted Saunders to spread the word about the medicinal qualities of the water in and around the area. Promptly, the City of Eureka Springs was founded and named on July 4, 1879. It became a resort almost overnight. By late 1879, the estimated population of Eureka Springs reached 10,000 people and in 1881, the town was declared a "City of the First Class," the fourth largest city in Arkansas. Today, the entire downtown is listed on the National Register of Historic Places, boasting well-preserved 19th-century architecture. https://www.eurekaspringsar.gov/visitors/page/city-history

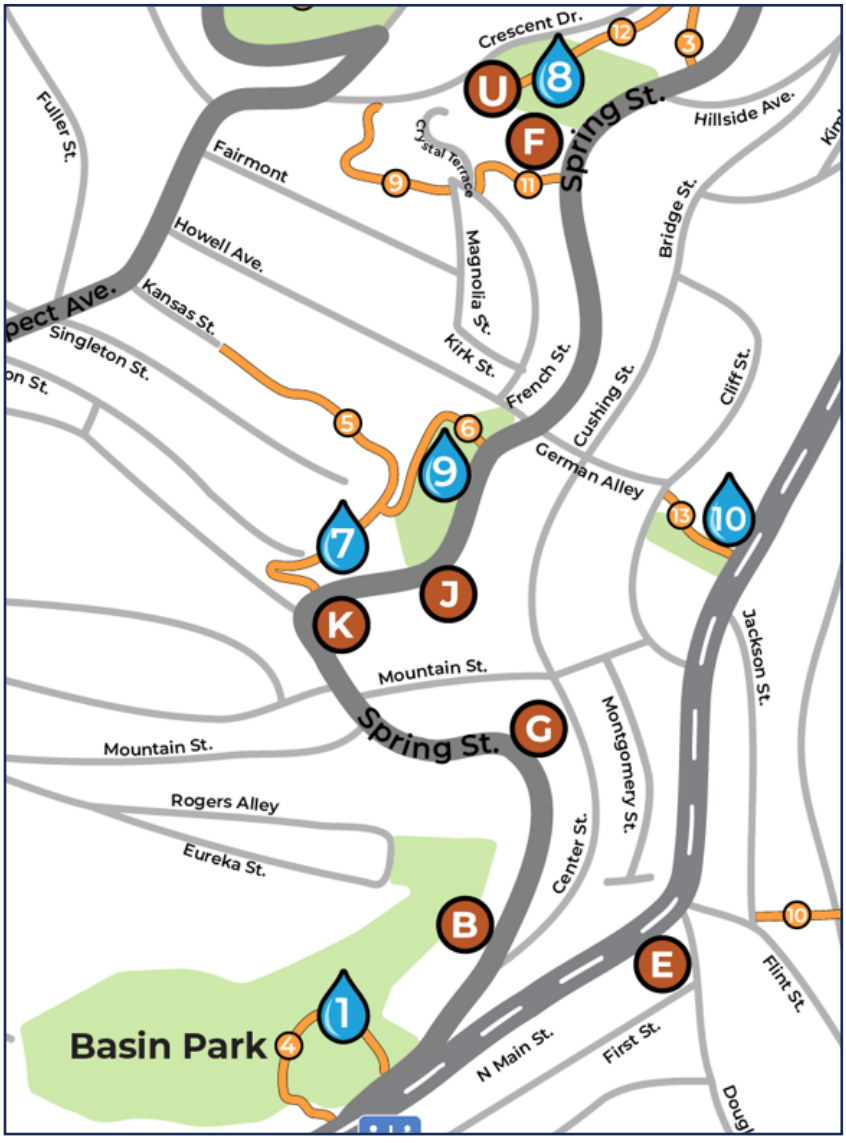


Figure 3.1.61 Part of Eureka Springs Downtown Parks, Springs & Trails Map. Springs are marked by blue water drops. Closest to the town center and most visited are (1) Basin Spring; (7) Sweet Spring; (8) Crescent Spring; (9) Harding Spring; (10) Gadd Spring. “*With over 60 natural cold springs, you're sure to find one (or more) that will steal your heart. From small caves to rock ledges to nooks and crannies, each spring has a unique charm. While the most popular springs can be found along Spring Street & Main Street, the majority of our springs lie just off the beaten path. So, make sure to pick up one of our Downtown Trail Guides and explore our beautiful town.*” https://www.eurekaparks.com/springs

"There is a story, very popular with the health-seekers camped around the spring they called The Basin, that Sioux Indians brought the young daughter of a great Chieftan to the Spring in search of healing. She suffered from an eye affliction which had taken away her sight, and the people were deeply saddened that such a fate should befall their beloved little princess. The young girl bathed her eyes in the waters, and within a short time her eyesight was fully restored, to the great joy of her people who told the story far and wide until it reached the ears of the first white men exploring the region."

--The Eureka Chronicle by June Westphal (https://www.eurekasprings.com/historical/).



Figure 3.1.62 Harding Spring issues at the base of a cliff in St. Joe Limestone.



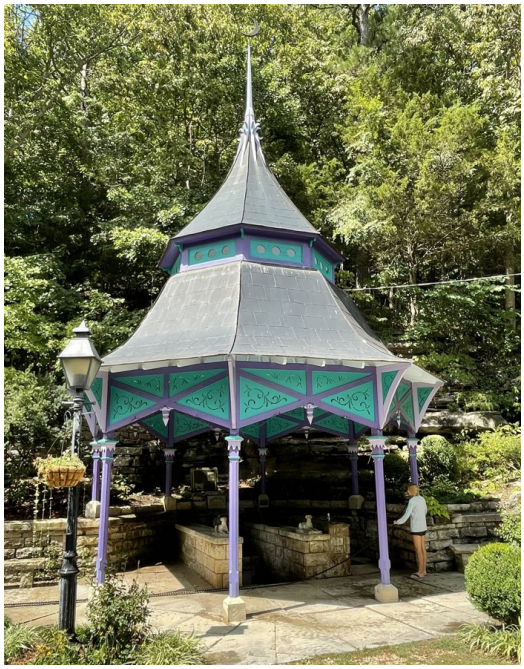


Figure 3.1.63 *Left*: Historic photo of Crescent Spring named for the crescent-shaped outcrop of St. Joe Limestone which can be seen to the left of the old pavilion. Courtesy of Eureka Springs Historical Museum. *Right*: New, restored purple and teal gazebo above the spring figures prominently on Spring Street.



The city of Eureka Springs lies on a steep escarpment between the Springfield Plateau and the Salem Plateau. The geologic formations are the Mississippian aged St. Joe Limestone member of the Boone Formation and older, Ordovician dolomites. These two units host the two regional aquifers in the area. Springs in this area have discharge variable with season and location, generally ranging from <1 to 5 cfs.

Figure 3.1.64 Historic houses on Spring Street are often built directly on top of St. Joe Limestone outcrops.

The St. Joe member of the Boone Formation is included in the Springfield Plateau aquifer. The St. Joe is a relatively pure phase limestone, with little chert. It is known throughout northwest Arkansas as a prominent cave former. Caves and karst in the St. Joe are largely aligned with solutionally enlarged fractures, which are arranged orthogonally and potentially were caused by regional uplift. The Chattanooga Shale comprises the Ozark confining unit, which separates the Springfield Plateau aquifer from the deeper Ozark aquifer. The Ordovician aged Cotter and Jefferson City Formations (Ozark aquifer) lie beneath the Chattanooga. Many prominent springs in the area flow from the Ozark aquifer. Locally, water from the Springfield Plateau aquifer passes though the Ozark confining unit by way of fractures. These fractures, along with direct infiltration into exposed Cotter and Jefferson City Formations, cause the Ozark aquifer to also be susceptible to contamination.

The springs and karst aquifer in Eureka Springs are heavily influenced by urban contamination. Of the known springs, none have acceptable water quality to serve as public drinking water supply. At their current levels of contamination, signs warn of the danger of drinking the water. Several of the springs are even considered public hazards by the Arkansas Department of Environmental Quality owing to their high indicator bacteria counts. Iron gates now restrict access to many springs and none of them are suitable for potable consumption without treatment.

"When the city was founded, the springs were the crowning jewel, and now their impurity is an embarrassment. The citizens of Eureka Springs are dedicated to the preservation and restoration of the springs. With the appropriate resources, these springs can once again be the pride of Eureka Springs."

Blanchard Springs, Stone County, Arkansas

Blanchard Springs is in the Sylamore Ranger District of the Ozark–St. Francis National Forest, fifteen miles northwest of Mountain View. As a natural outlet of the subterranean stream that formed the Blanchard Springs Cavern, the spring emerges from the mountainside in a waterfall and flows into a trout pond called Mirror Lake. According to the Office of State Geologist (www.geology.arkansas.gov/), discharge rate can vary between 1,000 to 103,000 gallons per minute (1.44–148 MGD) depending on local rainfall.

The spring was named for John H. Blanchard, who left his family's plantation in Kentucky and fought for the Confederacy, enlisting in the Kentucky Volunteers in 1861. Following bitter conflict at such battles as Chickamauga, where he was wounded, Blanchard sought peace after the war ended by homesteading 160 acres in the tranquil Ozarks. There, he built a gristmill powered by the falling spring which now bears his name.

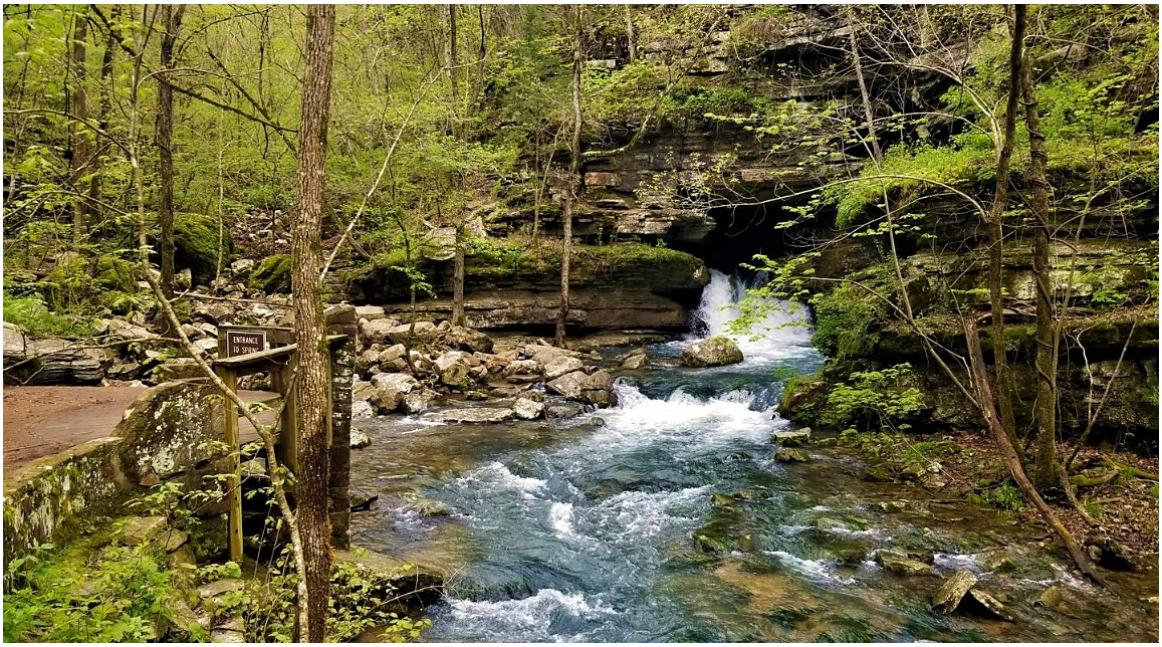


Figure 3.1.65 Blanchard Springs in Stone County, Arkansas issues from a cave. Courtesy of U.S. Forest Service.

Blanchard Springs Caverns (BCS) is one of the most spectacular and carefully developed caves in the United States and the only one operated by the US Forest Service. The active cave, with its glistening formations, stalactites, stalagmites, and flowstones is constantly changing, providing return visitors something new to see each time. Beautifully lighted rooms, accentuating the cave's features, are equipped with handrails and paved trails for comfortable walking. An undeveloped section of the cavern allows visitors an opportunity to climb over rocks, crawl through tight spaces and slide on red clay mudslides. (https://www.recreation.gov/ticket/facility/233266)

BCS is a cavity formed primarily in layers of limestone with thin layers of shale and sandstone which collapsed (https://www.blanchardsprings.org/geology-of-blanchard-caverns/). The cavern zone lies between bedded chert of the Boone Formation, and St. Peter Sandstone, which is also insoluble.

In December 2025, the State of Arkansas started a process to designate Blanchard Springs Caverns, "a jewel of the Ozarks", as a state park, to "preserve its cultural and environmental legacy while creating new opportunities for tourism, jobs, and year-round adventure."



Figure 3.1.66 Photographs from the newest Arkansas State Park, Blanchard Cave Caverns, taken by Katie Fite in support of the designation process. The powerful flow of the spring is best appreciated at the dam built by CCC. These and other fascinating photographs by Katie Fite are available at https://arkansas.app.box.com/s/4hzuip35ld9a3j50bwoktaxcdnpq4jmz.

3.2 Texas

Just like the state of Florida, Texas is rapidly destroying its two most valuable natural resources –its aquifers, including and especially the karstic Trinity-Edwards Aquifers, and the springs. And just like in Florida, the egregious mismanagement of the Texas groundwater resources is a runaway train. Except that this train is a Texas-size train, and the final inevitable train wreck will be the biggest in the country ("Everything's bigger in Texas").

Various government agencies, legislators, and a sizeable portion of the public in Texas are well aware of the deeply troubled situation. Below are excerpts from the 2027 State Water Plan "*Water for Texas*" by the Texas Water Development Board (TWDB), released on July 23, 2026, and available at https://www.twdb.texas.gov/waterplanning/swp/2027/index.asp

- Texas' population is anticipated to increase 53 percent between 2030 and 2080—from 34.2 million to 52.3 million—with more than 80 percent of this growth occurring within Regions C, G, H and L (*author's note: Region L is "South Central Texas" and encompasses the Edwards Aquifer*).
- Total water demands—including those associated with municipal use—are projected to increase by approximately 6 percent between 2030 and 2080, from nearly 17.4 million to nearly 18.4 million acre-feet per year. This slower rate of growth for the aggregated total demands, as compared to population growth, is influenced by declining irrigation demand projections over the planning horizon.
- Texas' existing water supplies—those that can already be relied on in the event of drought—are projected to decline by approximately 10 percent between 2030 and 2080. This is a decline from 15.5 million to nearly 14 million acre-feet per year primarily due to depletion of aquifers (emphasis added) with relatively smaller losses in reservoir yield due to sedimentation.
- Water user groups face potential water shortages of 3.6 million acre-feet per year in 2030 and 5.8 million acre-feet per year in 2080 in drought-of-record conditions. Municipal shortages account for 20 percent of needs in 2030, increasing to 57 percent by 2080.
- If Texas does not implement the water supply strategies and projects in the state water plan, a severe drought could cause an estimated $91 billion in economic damages in 2030, and those potential damages are projected to increase to $177 billion per year by 2080.

One of the key points highlighted in this plan by the TWDB (on page v) is that

"*The TWDB has no regulatory authority, related to the issuance of permits or groundwater management decisions or enforcement of drought contingency plans. The TWDB does not allocate or regulate water resources or supplies, nor is it involved in utility water supply contract decisions with their customers. The TWDB does not provide wholesale or retail water service, nor does it operate water supply infrastructure. The Board provides financial assistance for infrastructure development, collects water-related data, and administers the state's water supply planning process*."

In other words, the plan is just a plan and nothing more. Texas has a myriad of entities that are supposed to help manage its groundwater resources including "Authorities" (such as Edwards Aquifer Authority), "Boards" Groundwater Conservation Districts, and Groundwater Management Areas. Unfortunately, to no fault of their own, it is painfully obvious that they, for the most part, are underfunded, lack professional groundwater expertise, and, most importantly, have no real authority accompanied by strict legal enforcement, if not a mandate for criminal persecution.

Following are quotes from the Trinity Edwards Springs Protection Association (TESPA) webpage accessed in March 2023. This information, as of July 2026, has been removed from their website (https://tespatexas.org/). The quotes illustrate just one example of the unfortunate mismanagement of the Texas groundwater resources and the catastrophic failure to protect Texas springs which continue to dry up at an unstoppable pace.

Our work is far from over. The Hill Country is hot and dry right now. In this summer of 2023, Jacobs Well does not flow, again. Yet, development continues and investor-owned utilities like Aqua Texas continue to violate their groundwater permits. Local groundwater conservation districts do their best to enforce and apply penalties while avoiding lawsuits that tight budgets cannot support (emphasis added).

Our water means everything—our health, our economy, our connection to nature—everything. We must all stand together to ensure our water is managed sustainably. Now is the time to step up and take bold action. The situation is urgent, and YOUR voice is needed TODAY.

Aqua Texas failed to follow drought restrictions and Hays Trinity Groundwater Conservation District (HTGCD) rules and they failed to fix leaky pipes, wasting millions of gallons of our precious groundwater.

During the drought of 2022, Aqua pumped nearly 90 million gallons more than their permit allowed and they continue to pump. Aqua's noncompliance is unacceptable.



Figure 3.2.1 Jacob's Well spring, a popular swimming hole for generations of Central Texans, was dry in August 2023 because of over-pumping of the aquifer by Aqua Texas. Photo Credit: Dylan Baddour/Inside Climate News.

Aqua Texas has refused to settle its notice of alleged violation (NOAV) of the HTGCD rules and has failed to fix the leaky infrastructure.

Aqua Texas is owned by a multi-billion-dollar company that has prioritized profits for its shareholders at our community's expense. It's time to hold them accountable to state and local rules and guidelines for responsible management of our water.

Now Aqua is suing the HTGCD in federal court, arguing that they should not have to comply with rules and regulations, or the limitations of their permit.

In the words of Charlie Flatten, General Manager for HTGCD:

"[The District] has never threatened Aqua's or the public's access to water...As is the case with every utility (corporate or local), Aqua is required to follow the rules that they agreed to in their permit. Some of those terms include adhering to a drought contingency plan if necessary. Aqua has not shown that they adhered to that plan in 2022. [The District's] lawful mandate is to protect the aquifer for ALL users. During emergency drought conditions, we must all conserve—and Aqua did not do that in 2022 and 2023."

A Google inquiry about the Aqua case returned the following in July 2026:

- Aqua Texas says it will continue providing water to its customers despite the expired permit, citing a regulatory obligation to serve within its certificate of convenience and necessity area (Aqua).
- Settlement attempts: Negotiations occurred, but no resolution was reached. The District has not renewed Aqua's permit (Aqua).
- Court action: The case is ongoing; no final judgment has been issued, and the District is seeking to enforce its authority over groundwater use.



Figure 3.2.2 In June 2023, the water was deep enough for visitors to jump into Jacob's Well. Photo by Ian Dille, Courtesy of Texas Highways https://texashighways.com/

In 1991, The Sierra Club filed a lawsuit under the Endangered Species Act against the State of Texas that resulted ultimately in the creation of the Edwards Aquifer Authority (EAA). The Texas Legislature directed the EAA to regulate, among other things, pumping from the Aquifer, to implement critical period management restrictions, and to pursue a program "to ensure that the continuous minimum springflows of the Comal Springs and the San Marcos Springs are maintained to protect endangered and threatened species to the extent required by federal law" (EAA Act § 1.14(h)).

The following are among the major functions of the EAA as established by the EAA Act (EAA, 2024):

- Manage and control withdrawals of water from the Aquifer through the issuance of permits and the registration of wells.
- Protect the water quality of the Aquifer.
- Protect the water quality of the surface streams to which the Aquifer provides springflow.
- Achieve water conservation.
- Maximize the beneficial use of water available for withdrawal from the Aquifer.
- Protect aquatic and wildlife habitat.
- Protect species that are designated as threatened or endangered under applicable federal or state law.
- Provide for in-stream uses, bays, and estuaries.
- Protect water supplies.
- Protect the operation of existing industries.
- Protect the economic development of the state.

Less than 25 years since the Sierra Club won the lawsuit and the two largest springs in Texas seemed safe, the runaway train has already left the station as attested by the photographs in Figs. 3.2.3 and 3.2.4.



Figure 3.2.3 The most iconic of Comal Springs (Spring 1) has been dry multiple times over the last decade or so. Right photo taken on September 6, 2013. Photographs courtesy of USGS.



Figure 3.2.4 *Left*: Comal Spring 1 Run; courtesy of USGS. *Right*: Spring Run on February 10, 2026. From BIO-WEST, 2026.

Las Moras Springs, Kinney County

An Iconic Texas Spring Was Dry for Months. Is Mother Nature or Man to Blame?

By Forrest Wilder, Texas Monthly, October 2023. This article originally appeared in the October 2022 issue of Texas Monthly with the headline "Gone With the Flow." Below are several excerpts (full article is available at https://www.texasmonthly.com/news-politics/why-did-las-moras-fort-clark-springs-dry-up/).

Plans were underway to revive tourism at Fort Clark Springs in southwest Texas. But then, in a scenario increasingly common across the state, the water stopped flowing.

April 8 is significant to Travis Huey for two reasons. It's the date when, in 2024, the narrow path of a total solar eclipse will pass over Brackettville, the small southwest Texas town where he lives. Huey owns the local

newspaper, the Kinney County Post, and he's the president of the association that runs Fort Clark Springs, a private resort and historic site in Brackettville that has struggled to attract visitors in recent years. Huey has big plans for the fort that include drawing 5,000 visitors to the four-minute, eighteen-second eclipse.

But April 8, 2022, complicated his plans. That's the day when Las Moras Springs, which typically pumps out 12 to 14 million gallons of cool, clear water a day—enough to fill two dozen Olympic size swimming pools—stopped flowing into the fort's swimming pool, the third-largest spring-fed pool in the state and Fort Clark's main attraction.

Conoly said he believes the board is making moves to abandon healthy spring flow as a planning goal and instead do what many other Texas districts do—assume a future in which the aquifer is depleted.



Las Moras Springs is on the property of Fort Clark in Brackettville. The springs rise under artesian pressure from the Edwards and associated limestones and pass through a fault in the overlying formation, emerging at an elevation of about 1096 feet.

The springs fill a large walled-in area which spills into a 300-foot-long swimming pool, with excess flows diverted to a bypass channel around the pool. Just below the pool, both discharges combine and form the headwaters of Las Moras Creek.

Figure 3.2.5 An aerial view of dry spring basin and swimming pool at Fort Clark Springs (Las Moras Springs) in Brackettville, Texas, spring 2022. Photograph by Tamir Kalifa. Courtesy of Forrest Wilder, Texas Monthly.

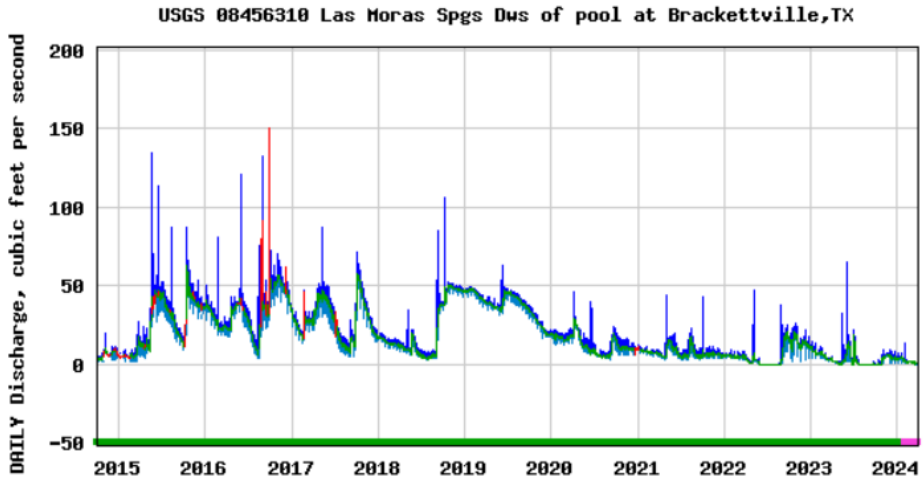


Figure 3.2.6 Discharge rate of Las Moras Springs recorded by USGS. Small oscillations are caused by irrigation wells in the springs' area being turned on and off.

Spanish explorers began to camp at the Springs in the 16th century and gave them their name, which means "The Mulberries". In 1840 a cavalry unit drove Comanches from their village at the springs. At that time, many bison, antelope, and wild mustangs roamed the area (Brune, 1981).

As explained by Forest Wilder, Fort Clark was established in 1852 as a frontier military outpost along the San Antonio–El Paso Road, an important trade route that facilitated westward expansion into Indian Territory and guarded against a restive Mexico to the south. Black Seminole scouts served at the fort from 1872 until 1914, and one of the last horse-mounted cavalry divisions trained here during World War II. After the war, the fort was deactivated, sold to Brown & Root, the Texas construction behemoth, and later converted into a guest ranch. Today its sprawling grounds—all run by a homeowners' association—include a golf course, almost nine hundred homes, a motel, nature trails, an airport, an RV park, and more than thirty historic buildings, some of which date back to the fort's founding.



Figure 3.2.7 The Fort Clark Springs swimming pool, fed by Las Moras Springs, in 2018. Photograph by Wynn Myers. Courtesy of Forrest Wilder, Texas Monthly.

Phantom Spring, Jeff Davis County

Located 6 miles west of Toyahvale, Phantom Spring and cave is owned by the Federal Bureau of Reclamation. Access to the cave is tightly restricted and requires a special scientific permit. No other access to the cave or the surrounding property is allowed under any circumstances. Only two permits are issued at a time due to concerns about overuse of the fragile ecosystem.

Fascinating facts about Phantom Spring (also referred to as Phantom Lake Spring), for a long time known as the deepest spring in the country (462 feet) before losing that title to Roaring River Spring in Missouri (>520 ft; see Fig. 3.1.56 in Chapter 3.1) are available at the excellent website of ADM Foundation. Am abundance of information, fascinating photographs and video footage are provided about a spring cave diving expedition, the

science behind it, ecology, biology, geology, and the preservation of endangered species. Visiting this webpage is a must (http://www.admfoundation.org/projects/phantomcave2013/phantom2013.html).

Phantom Spring was declared clinically dead in 1999 (see graph on Fig. 2.2.15, in Chapter 2.2). Although one would never find a stamped, official death certificate listing the cause of death in any publication of various districts and agencies responsible for the well-being of Texas groundwater and springs, there is little doubt in anyone's mind what it was.

This quote is from the 2012 narrative available at the ADM Foundation website: "*In addition to groundwater extraction from municipalities and irrigation, these desert ecosystems are faced with the threat of hydraulic fracturing, also known as fracking. Serving as the primary method to extract petroleum and natural gas from deep shale formations, fracking activity is increasing at a rapid rate in Texas. More than 10 natural gas wells have been reported between Phantom Springs and the nearby town of Pecos. Over a well's lifetime, it can require up to several million gallons of water. In addition to requiring large quantities of water, fracking fluid is composed of chemical additives including known carcinogens such as benzene, lead, and boric acid. Disposal of fracking fluid often consists of injecting this polluted brine into disposal wells around the area.:*



Figure 3.2.8 The mouth of Phantom Lake Spring after the 2022 monsoon season. Photo courtesy of Texas by Nature. More photographs and descriptions of six springs in the Balmorhea area are provided at https://texanbynature.org/projects/water-in-the-desert-the-story-of-san-solomon-springs/

Phantom Lake Spring is still the official home to five endangered species: Comanche Springs pupfish (*Cyprinodon elegans*), Pecos gambusia (*Gambusia nobilis*), Phantom tryonia (*Tryonia cheatumi*), Phantom springsnail (*Pyrgulopsis texana*), and diminutive amphipod (*Gammarus hyalelloides*). Comanche Springs located within the city limits of Fort Stockton, Texas went dry long time ago resulting in the loss of the pupfish from this locality. It is now only known from a group of springs around Balmorhea including Phantom Springs. Likewise, the Pecos gambusia was once more widely distributed, but today is only found in springs surrounding Balmorhea.

As part of their 2012 exploration, the ADM Foundation team laid down sediment traps, collected water quality data, and surveyed over 8,000 feet of the Spring cave. They also did dye tracing to determine where the water in Phantom Cave was traveling. Two days later, the dye came out in San Solomon Springs near Balmorhea. It is therefore postulated that, since the water is not discharging at Phantom Springs anymore, most if not all of it is now helping San Solomon Springs survive the continuing onslaught on its aquifer.

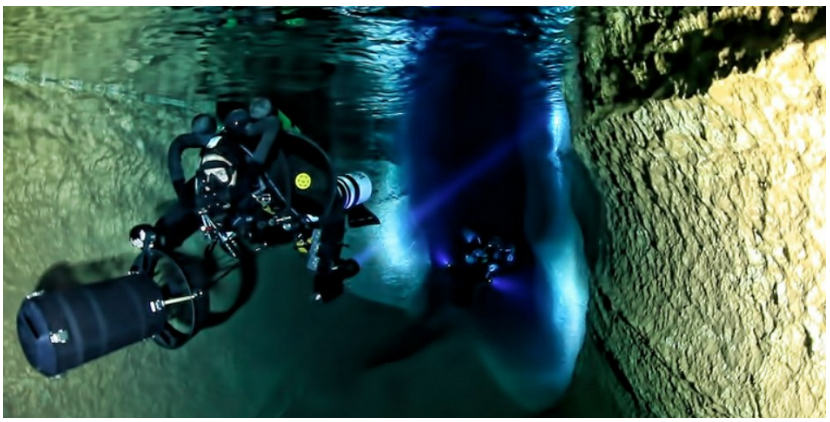


Figure 3.2.9 Explorers Brett Hemphill and Curt Bowen illuminate the furthest reaches of Phantom Spring for videographer Andy Pitkin. Photo courtesy of ADM Exploration Foundation.

The big springs of Balmorhea and Fort Stockton are believed to have their source in an extensive network of fissures and solution passages in limestones of the Comanche series of Lower Cretaceous age, which are the oldest strata exposed near Balmorhea. The Comanche section includes, besides the limestones, interbedded calcareous shale and a basal unit composed of sand or sandstone.

The limestones are mainly massive, thick-bedded, and very fossiliferous. They weather characteristically to dark-colored craggy or solution-rounded ledges and low rolling to hilly topography.

The uppermost limestone of the Lower Cretaceous series is well exposed near Phantom Lake Spring in a ridge that extends northwest from the spring along the south side of the Old Spanish Trail Highway. This limestone contains several deep, cavernous channels and crevices in this general vicinity, and from one of these crevices emerges the water that feeds Phantom Spring (White et al., 1941).

San Solomon Springs, Reeves County

San Solomon Springs are in Balmorhea State Park, four miles southwest of Balmorhea on State Highway 17, in Toyahvale. “Dive into the crystal-clear water of the world’s largest spring-fed swimming pool in a cool oasis in the high desert. Swim, scuba dive, or just relax under the trees at this historic park in arid West Texas.” (https://tpwd.texas.gov/state-parks/balmorhea).

In an old-fashioned, informative, and engaging USGS style of the past, White et al. (1941) describe the geology and groundwater resources of the Balmorhea Area in west Texas where the San Solomon Springs system is:

Balmorhea is the center of a thriving farming community, the lands of which are irrigated with water derived chiefly from large springs but partly from the storm flow of Toyah Creek. Balmorhea is situated near the foot of the Davis and Barrilla Mountains and along the southwestern margin of the Toyah Basin (see Fig. 3.2.10)*. The mountains and adjacent basin are drained by Toyah and Limpia Creeks.*

The group of springs around Balmorhea occur in the floor of the valley of Toyah Creek. They have been divided into artesian springs—Phantom Lake, Giffin and San Solomon Springs; and gravity springs—Toyah Creek, Saragosa, East Sandia and West Sandia Springs. The combined discharge of the springs during dry years is about 23,000 gallons a minute, of which the artesian springs supply more than 90 percent.

The underground reservoir which supplies the artesian springs is the fractured and cavernous Lower Cretaceous limestone. This limestone, about 500 feet thick, is underlain by impermeable rocks, probably of Permian

age, and is overlain by impermeable Upper Cretaceous strata that have a maximum thickness of about 500 feet. These are in turn overlain in the mountains by Tertiary lava and on the plains by gravel and other surficial deposits.

San Solomon and Giffin Springs rise from gravel deposits in the floor of the valley near Toyahvale, about 4 miles southwest of Balmorhea. The water originally came to the surface, forming a swamp of considerable size, which drained into Toyah Creek. San Solomon Spring now issues from gravel in the bottom of a large well-built bathing pool. The water is beautifully clear, and the discharge at all times is that of a good-sized creek. It is not surprising, therefore, that this spring is one of the leading scenic attractions of the region.

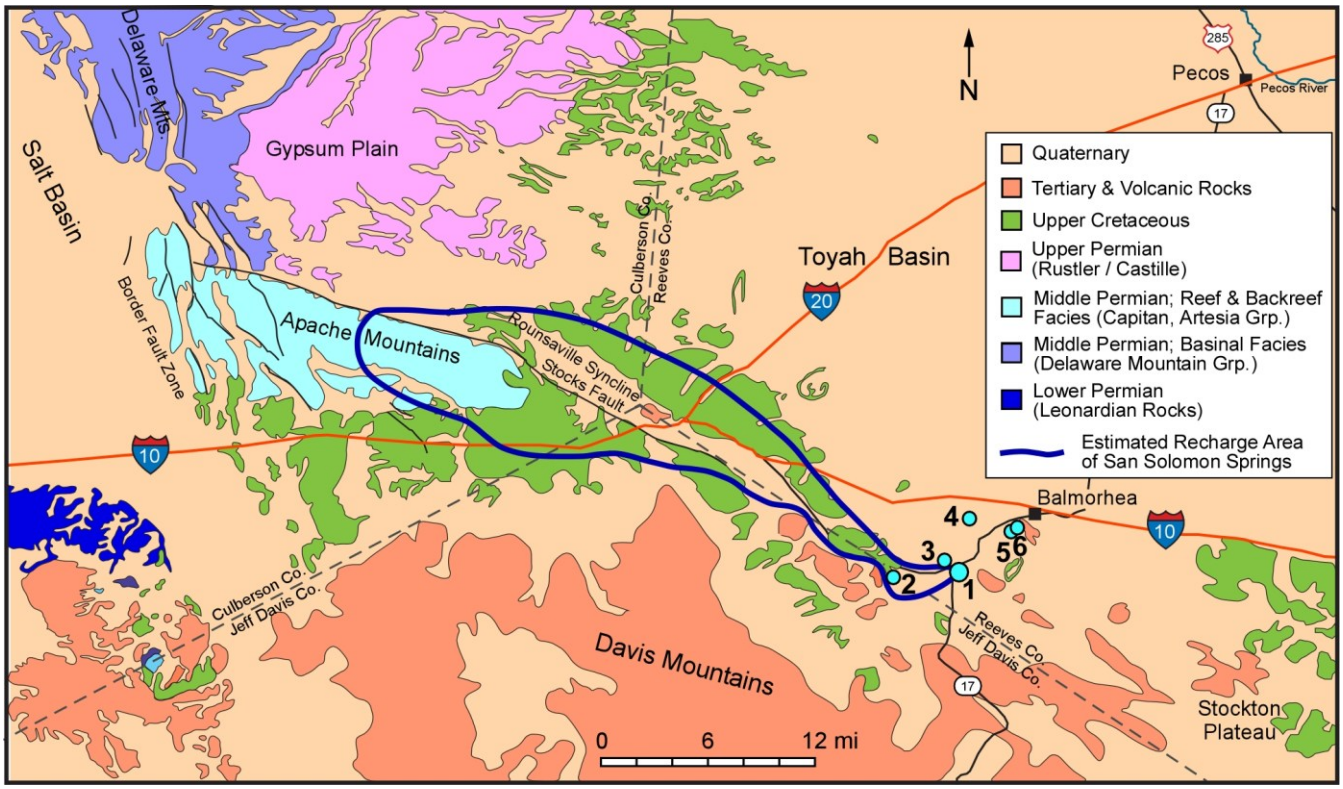


Figure 3.2.10 Hydrogeologic map of the San Solomon Springs system area. (**1**) San Solomon Springs; (**2**) Phantom Lake Spring; (**3**) Giffin Spring; (**4**) Saragosa Springs; (**5**) West Sandia Springs; (**6**) East Sandia Spring. Modified from Land and Veni, 2018. (Veni, 2013; based on Barnes, 1975; 1976; 1979; 1982.)

The lowest discharge of San Solomon Spring recorded by the Geological Survey was 26.5 second-feet on April 26, 1923 (p. 121), and the highest was about 71 second-feet on October 7, 1932 (p. 131). The lowest recorded flow of Phantom Lake Spring was 10 second-feet on October 16,1931 (p. 142), and the highest 114 second-feet October 2 and 3,1932 (p. 127). The discharge of the Giffin Springs is relatively small, the smallest recorded daily flow being 2.9 second-feet March 4, 1925, and the largest between 6 and 7 second-feet in October 1932.

As explained by Land and Veni (2018), Trans-Pecos Texas experiences an average annual precipitation of less than 300 mm; surface-water resources are scarce and brackish in this semi-arid region. Groundwater from wells and springs thus plays an important role in municipal, domestic and irrigation water supplies in the area (Sharp et al., 2003).

The springs of the San Solomon Springs group were used by Native Americans since prehistoric times; their irrigation canals were visible as late as 1898 (Hutson, 1898). The springs became known to European explorers and

settlers in 1583 when Spanish explorer Antonio de Espejo made the historic discovery of San Solomon Spring (Castañeda, 1936). In 1849, the springs became known as Mescalero Springs for the Mescalero Apache who watered their horses there.

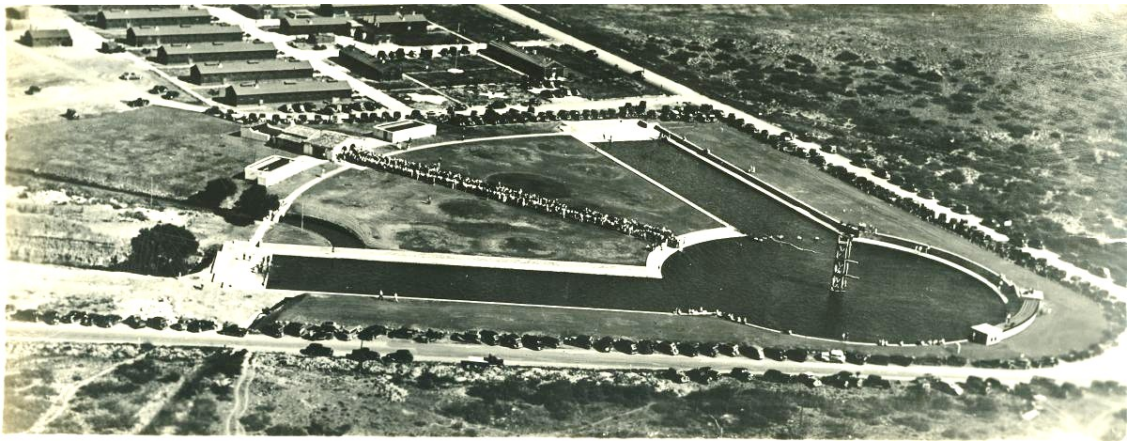


Figure 3.2.11 Historic photograph of San Solomon Springs swimming pool. Courtesy of Texas Parks and Wildlife Department. Available at https://texanbynature.org/projects/water-in-the-desert-the-story-of-san-solomon-springs/. Seeing the headwaters of San Solomon Springs as an attractive location for a park, the State Parks Board acquired the land in 1934. The Civilian Conservation Corps (CCC) arrived in July 1934, first building barracks, mess hall and a kitchen to support 130 to 200 men. The iconic double-wing swimming pool was built between 1935 and 1940. The pool is up to 25-feet deep, covers 1.3 acres and holds 3.5 million gallons of water. The largest spring, San Solomon Spring (singular) discharges into the pool.

Mexican farmers called the springs "San Solomon Springs." They dug the first canals by hand and then used the water to irrigate crops. They sold those crops to residents of Fort Davis. With plentiful water and the arrival of the railroad, a cattle ranching industry emerged in the 1880s. The Town's and Park's name comes from four men's surnames: E.D. Balcom, H.R. Morrow, Joe Rhea and John Rhea: Bal-mor-hea. These men formed an irrigation company in the area in the early 20th century.

The State Parks Board acquired nearly 46 acres around San Solomon Springs in 1934. Civilian Conservation Corps Company 1856 built the park between 1935 and 1940.

On the State Park's website (https://tpwd.texas.gov/state-parks/balmorhea/protecting-springs) one can read:

Water gushing from San Solomon Springs fills the pool at Balmorhea State Park, then flows downstream through cienegas and canals in the park and beyond. Not only does the water refresh hot, dusty Texans, it is also home to several rare and endangered species.

Texas Parks and Wildlife Department's mission is to protect the springs and this historic state park.

Oil and Gas Production

Oil and gas wells have dotted far West Texas over the last 50 years. But recent innovations such as hydraulic fracturing or "fracking" have changed how companies extract fuel from the ground.

As a result of these innovations, the Apache Corporation announced in fall 2016 it had identified a new area for pumping – the Alpine High Play. This play includes southern Reeves County, home to Balmorhea State Park.

Apache estimates it will have 2,000 to 3,000 well locations in this play, but possibly up to 5,000. Apache Corporation is not the only company operating in this area. Drilling may last for more than 20 years.

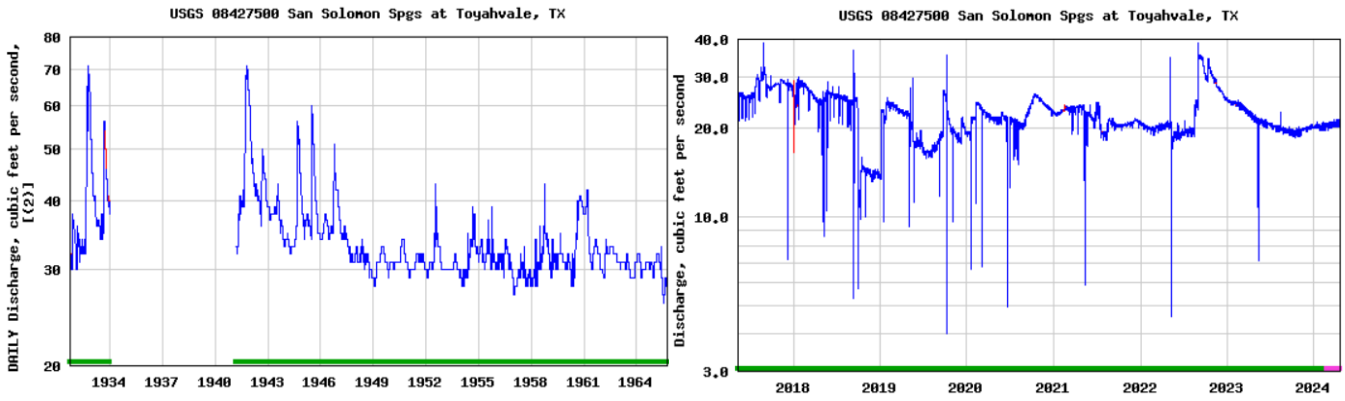


Figure 3.2.12 Hydrographs of daily discharge rate of San Solomons Springs. For October 1931–September 1965 (left) the average is 34 cfs. For October 2017–April 2024 (right) the average is 22 cfs.

Concerns

How might oil and gas production impact the springs?

One concern is that water use might diminish spring flows. Fracking to extract oil and gas requires 2 to 5 million gallons of water per well. Apache intends to reuse water as much as possible during the fracking process. It's undetermined how much water from the local area oil and gas producers will use.

Diminished spring flows or lower water quality would very likely harm the species that depend on the San Solomon Springs for survival. Either of these could also make the pool unsuitable for swimming.



Figure 3.2.13 Swimming pool in Balmorhea State Park fed by the main San Solomon Spring. It is the largest spring-fed pool in the world. Photos courtesy of Texas Parks and Wildlife Department, https://tpwd.texas.gov/state-parks/balmorhea.

Langford Hot Springs, Brewster County, Big Bend National Park

Langford Hot Springs is in Big Bend National Park (https://www.nps.gov/rigr/index.htm) in remote southwest Texas, about four miles upriver from Boquillas Canyon and the Mexican village of Boquillas. The crystal-clear water of the spring issues, at near constant temperature of 105 degrees Fahrenheit (40.5°C), from Boquillas limestone on the north (left) bank of the Rio Grande River, near the confluence of Tornillo Creek and the Rio Grande.

Below are a few excerpts from the fascinating story of Langford Hot Springs as told by Park Ranger Jeannette Woerner of the National Park Service.

"The hot springs of Big Bend have long been a focal point for tourists and travelers in the desert. But these hot mineral waters represent more than just a destination for visitors to the National Park. For J.O. Langford and his family, they represented health, business and home. And for several decades in the early 20th century, the hot springs were also the center of a desert community.

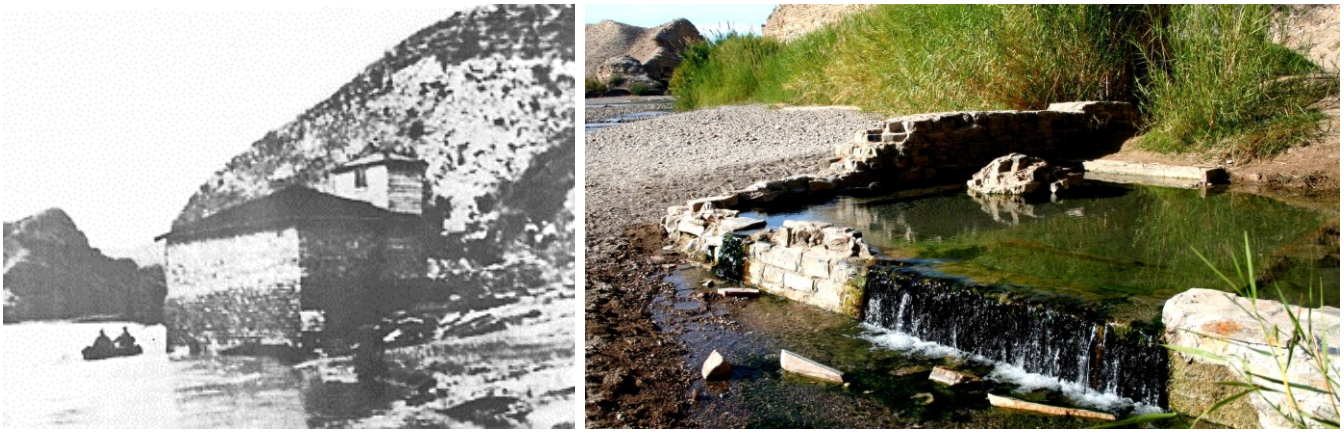


Figure 3.2.14 *Left*: Historic photograph of the original bathhouse built by Langford over the main hot spring. *Right*: Today only remains of the foundation are visible. Photo by Matthew Yarbrough, courtesy of NPS.

In 1909, J.O. Langford was sitting in the lobby of the Alpine Hotel when he overheard someone discussing the miraculous healing properties of the hot springs in Big Bend country. Another person might have been skeptical of claims that the mineral water could cure any disease, but Langford had been ill since contracting malaria as a child and had been looking for relief from his pains and sickness ever since. In fact, he was so interested in these springs that he immediately set out to purchase the land surrounding them, down in a remote, lonely country that he'd never even laid eyes on.

Langford set out two weeks later with his pregnant wife Bessie and their young daughter Lovie, and traveled for ten days before finally reaching their new home on the Rio Grande. Regaining his health was a priority, so as soon as they arrived, Langford began his treatment: For 21 days, he drank and bathed in the mineral water of the spring, which at its source reaches 105 degrees, and soon he felt healthier and stronger than ever. Langford recognized the value of the springs and he immediately set up a business. He built a bathhouse (Figure 3.2.14) and developed his land into a desert health resort. Word-of-mouth recommendations soon brought people from across the country to find health at the springs, and many claimed to have been cured of skin diseases, rheumatism, stomach troubles, and chronic pain.

Langford and his family lived and prospered in the Hot Springs area until 1942, at which time he sold the land to the state of Texas. The purchase was made in hopes of one day protecting the area, and two years later, Big Bend National Park was established.

While the buildings have been abandoned since the 1950s, they still stand as a reminder of the community they once supported. To this day, you can visit the remains of the general store and motor court. A walk just a quarter mile down the river will bring you to what is left of the main bathhouse, and you too can relax and soak in what J.O. Langford and many other West Texans considered to be healing waters."

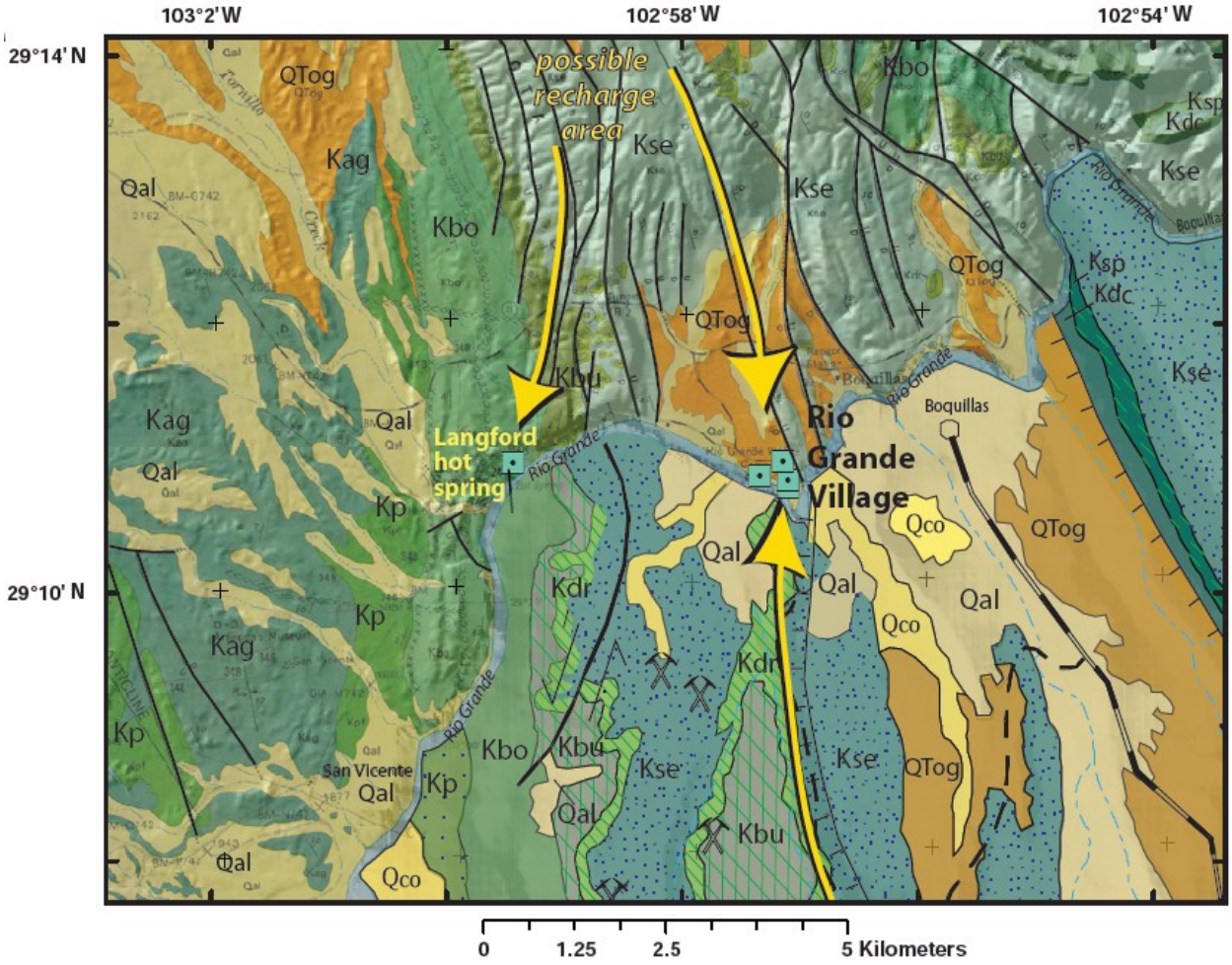


Figure 3.2.15 Geologic and shaded relief map of the Rio Grande Village (RGV) area. Turquoise square symbols indicate water samples. Geologic units are: Qal, Quaternary alluvium; Qco, Quaternary colluvium; QTog, Quaternary to Tertiary older gravels; Tacb, Tertiary Alamo Creek Basalt; Tasb, Tertiary Ash Springs Basalt; Tig, Tertiary igneous rocks; Tch, Tertiary Chisos Formation; Tcf, Tertiary Canoe Formation; Tbmb, Tertiary Bee Mountain Basalt; Tbr, Tertiary Brown Rhyolite; Twsf, Tertiary Wasp Spring Flow Breccia; Kag, Cretaceous Aguja Formation; Kp, Cretaceous Pen Formation; Kbo, Cretaceous Boquillas Formation; Kbu, Cretaceous Buda Formation; Kse, Cretaceous Santa Elena Formation; Kdr, Cretaceous Del Rio Formation; Kdc, Cretaceous del Carmen Formation; Ksp, Cretaceous Sue Peak Formation; Kjf, Cretaceous Javelina Formation. Big Bend National Park geology from Maxwell and others (1967). Mexican geologic mapping from Servicio Geológico Mexicano (2003). Note location of Langford hot spring, west of RGV, the hottest thermal water spring in the Park at 40.1°C. Water samples at Rio Grande Village are all thermal waters similar in composition, but somewhat more diluted than Langford waters. Hot spring water is probably recharged about 5 km north of RGV in higher ground of the Cretaceous Santa Elena limestones and flows along N-S fault zones (yellow arrows) to the hot spring sites. Possible recharge sites to the south in Mexico also are shown. Mine symbols (crossed picks) indicate fluorite mines. From Pat Shanks et al., 2008.



Figure 3.2.16 Langford Hot Spring in Big Bend National Park in November 1997 (left), and March 2017 (right).

Native Americans used the hot springs area as a homestead well before Europeans arrived in this part of Big Bend. The steep rocky cliffs provided shelter, as well as a canvas. Rock art, including pictographs and petroglyphs, adorn the limestone walls. Aprons of rock jutting into the Rio Grande reveal bedrock mortar holes once used for grinding mesquite beans and other seeds. J. O. Langford found that Native Americans had bathed in the hot springs where the hot water "...poured off a lip of rock and into the casket-like bathtub that the Indians built sometime in the past by chipping out and enlarging a fissure in the flat layer of sedimentary rock." The later construction of the bathhouse covered that historic tub (NPS).



Figure 3.2.17 Cliff of Boquillas limestone near Langford Hot Spring.

Comal Springs, Comal County

Comal Springs in Central Texas are in northwest New Braunfels, in Landa Park which is owned by the city. They are the largest springs in the southwestern United States and issue from various points along the base of an escarpment formed by the main Balcones normal fault in the area, named Comal Springs fault (see Fig. 1.25 in Chapter 1). The seven major and many smaller spring outlets occur for a distance of 4300 feet along the fault. Many of the springs lie beneath Landa Lake or along its edge. Upthrown to the fault, Edwards carbonate rocks are exposed in Panther Canyon along the nature trail. The same rock layers are over 800 feet below the surface of Landa Lake.

The springs give rise to the Comal River, which runs for about three miles before joining the Guadalupe River and is known as the shortest significant river in the United States. The steady flow of water was developed for water power beginning in 1850s. Mills were built downstream of the springs, where the natural spring lake was enlarged by a low dam.

As explained by LBG-Guyton Associates (2004), discharge measurements by Ogden et al. (1985) and BIO-WEST (Ed Oborny, personal communications, 2003) indicate that during periods of normal to high flow only about one quarter of the total spring discharge (as measured at the USGS gage) is from the western group of large springs shown in Fig. 3.2.18.

The measurements also show that under conditions of lower flow, the contribution of this western group of springs, which are at higher elevations than the others, is reduced until these springs "go dry," at which point the lake springs are the sole components of total flow.

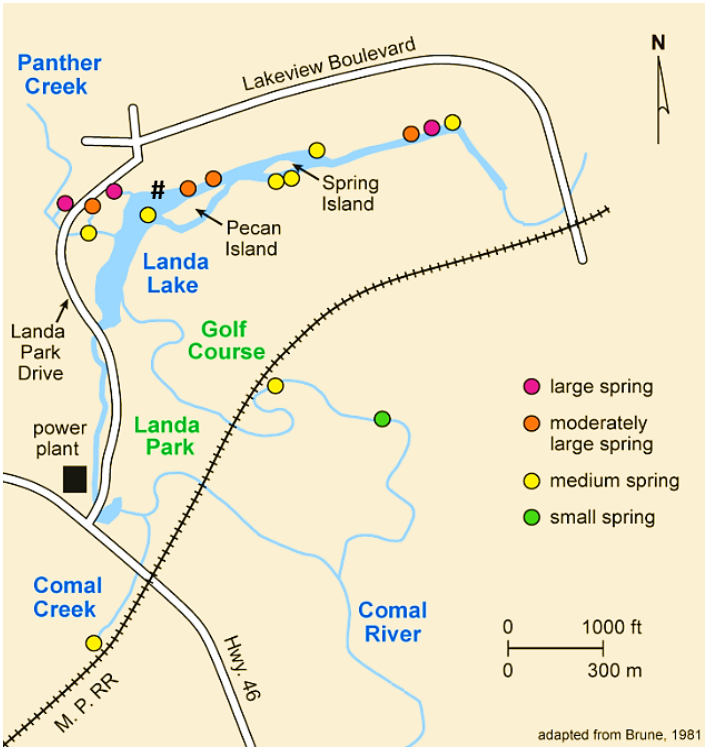


Figure 3.2.18 Locations of Comal Springs. The largest spring, commonly referred to as Spring #1 (spring "l" in Brune, 1981). Courtesy of Gregg Eckhardt. https://www.edwardsaquifer.net/

Discharge of springs in the lake is evidenced by small sand boils, saltation of sand grains, and areas where extensive clouds of gas bubbles can emanate from the lake bottom. Gas bubbles emanating from the bottom are considered evidence of groundwater discharge. The gas bubbles occur because the discharging groundwater is oversaturated with dissolved atmospheric gas. As Edwards groundwater rises from greater depths in the aquifer and discharges into the lake or individual springs, it depressurizes because of the decreasing hydrostatic head as it approaches land surface. As the water rises these gases come out of solution forming bubbles and bubble trains (LBG-Guyton Associates, 2004).



Figure 3.2.19 Landa Lake in Landa Park, owned by the City of New Braunfels, is formed by numerous springs discharging around the lake or directly into it. The average discharge rate of all springs combined, for the period of record, is 280 cfs.

Regional flow paths originating in the western part of the Edwards Aquifer are generally understood to supply discharge at Comal Springs which have a small range of hydrologic and geochemical variability (Musgrove and Crow, 2012).



However, only a portion of the regional flow paths originating in the western part of the Edwards Aquifer are generally understood to supply discharge at Comal Springs (Musgrove and Crow, 2012). Groundwater from the up-thrown block of the Comal Springs Fault appears to be discharged from the western group of orifices at Comal Springs. The remainder moves eastward and discharges at San Marcos Springs (LBG-Guyton Associates, 2004).

Figure 3.2.20 *At three-miles long and a cool 70-72-degree temperature year-round, the spring-fed Comal River in New Braunfels provides a scenic, three-hour float. Tube gently along crystal clear waters amidst a scenic landscape with occasional strong currents. Easy float for families.* Text and photograph from *10 Of The Best Lazy Rivers To Go Tubing In Texas* by Colby Smith, Secret Houston, April 24, 2024. Photo credit: @garytray https://secrethouston.com/best-lazy-rivers-texas

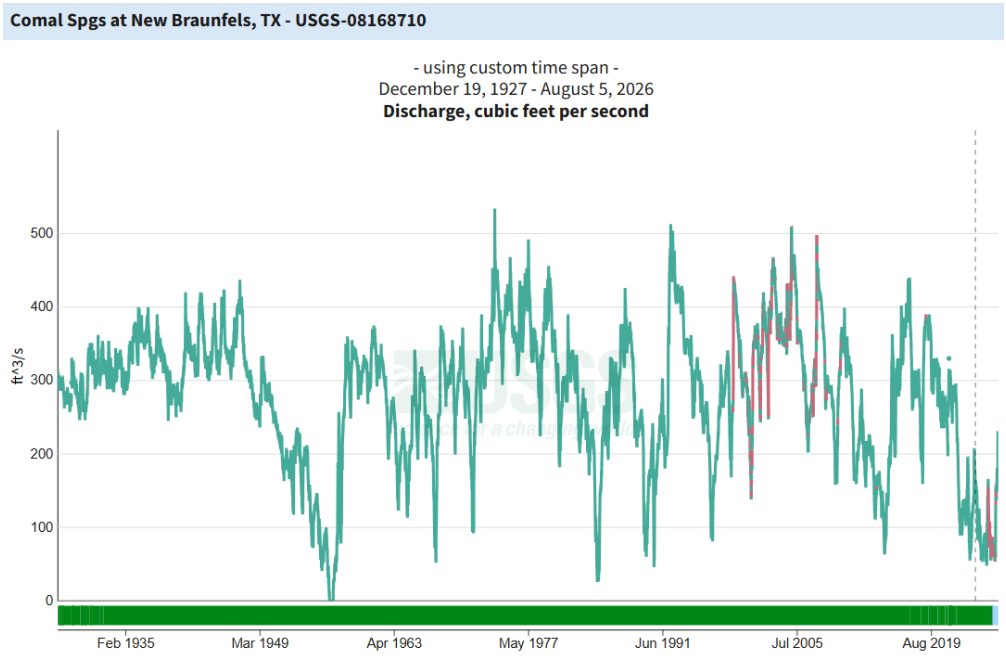


Figure 3.2.21 Discharge rate of Comal Springs for the period of record. The average is 280 cfs. All spring vents went dry in the summer of 1956 during the most severe drought ever recorded in Texas.

San Marcos Springs, Hays County

San Marcos Springs in the City of San Marcos in southeast Hays County are the second largest spring group in Texas and form headwaters of the San Marcos River. The springs (also known as Aquarena Springs) issue from the Edwards aquifer at the bottom of Spring Lake. The lake was built in 1848 on the site of a marsh formed by the springs to provide hydropower to a mill. The San Marcos River originates from the 15-acre lake and flows approximately 4 river miles south to its confluence with the Blanco River.



Figure 3.2.22 Historic photo of Spring Lake built in 1848, showing power plant and western end of the lake. Photographed about 1900. From Taylor, 1904.

Sink Creek flows into Spring Lake from the east. Two connected dams control outflow from the lake on the western end.

The springs at San Marcos Springs are all within Spring Lake. The names of the individual springs were given over the years by the glass-bottom boat guides at Aquarena Springs. There are no springs discharging at land surface and flowing to the lake as occurs at Comal Springs (LBG-Guyton Associates, 2004).

The thermally consistent water from the San Marcos River has long been noted and generally varies by less than 1 or 2°C in the headwaters at any time during the year. The stability of this stream, both in terms of flow dependability and thermal characteristics, is thought to have provided the appropriate ecological conditions necessary to allow the extreme endemism of the San Marcos River biota (Edwards et al., 1980).

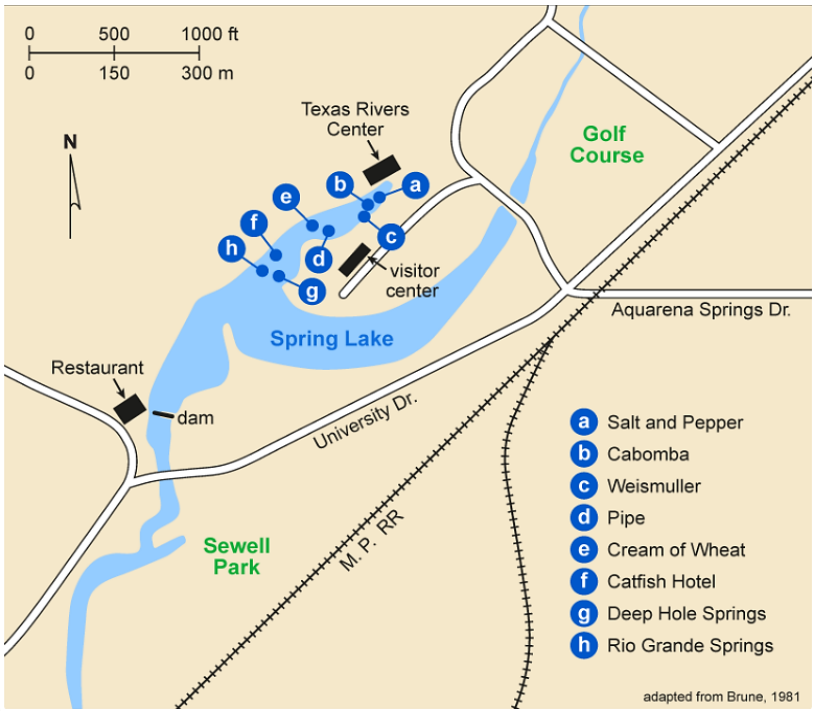


Figure 3.2.23 Locations of larger San Marcos Springs. Courtesy of Gregg Echardt https://www.edwardsaquifer.net/

A brief sketch of the earliest inhabitants and visitors to the San Marcos Springs and surrounding vicinity is provided by Brune (1975):

When the Spanish explorers discovered these springs in 1743, they estimated there were 200 springs. From 1755 to 1756 a mission was located there. The springs were an important stop on El Camino Real from Nacogdoches to Mexico. In 1840 Bonnell described them as "the most pleasant and delightful situation in the Republic." Power plants, gins, corn mills, and an ice factory used the water power. The springs were a stop on the Chisholm Cattle Trail from 1867 to 1895.

The city of San Marcos developed around the headspring area following early flooding and Indian attacks at the original site of the city several kilometers downstream from the headsprings. An early Federal fish hatchery was established near the springs (Jordan and Gilbert 1895). In the late 1920s the springs became a part of a privately owned resort hotel and later an amusement water park, Aquarena Springs, which operated between 1950 and 1996. The property of Spring Lake was purchased by Texas State University in 1994.

Since 2012, the Spring Lake is managed by The Meadows Center for Water and the Environment of Texas State University (https://www.meadowscenter.txst.edu/about.html). The Center offers glass-bottom boat tours, guided snorkel tours, guided stand-up paddling tours, nighttime light up kayak tours, scuba diving, and various educational and training programs.

According to LBG-Guyton Associates (2004), discharge from the identified spring areas within Spring Lake occurs as (1) discrete flow out of a defined orifice, such as a fissure; (2) diffuse flow out of the rubble of angular blocks; and (3) strong sand boils or sand geysers out of sand plains composed of white and tan, fine-grained quartz and calcite sands and black organic materials.

The measured velocity of groundwater discharging from the John Weissmuller orifice was 2.79 feet per second (ft/sec); the discharge rate was estimated at 15 cubic feet per second (cfs). Velocity of water discharging through unconsolidated sand sediments, which cover the conduits in Edwards Limestone, could not be measured and was estimated at 0.03 ft/sec (2,592 ft/day) in the areas of fascinating sand boils and geysers, with the total discharge rate of 142 cfs. The average discharge rate of San Marcos Springs, for the period of record, is 170 cfs.



Figure 3.2.24 Two western outlets from Spring Lake, San Marcos Springs, February 2015.

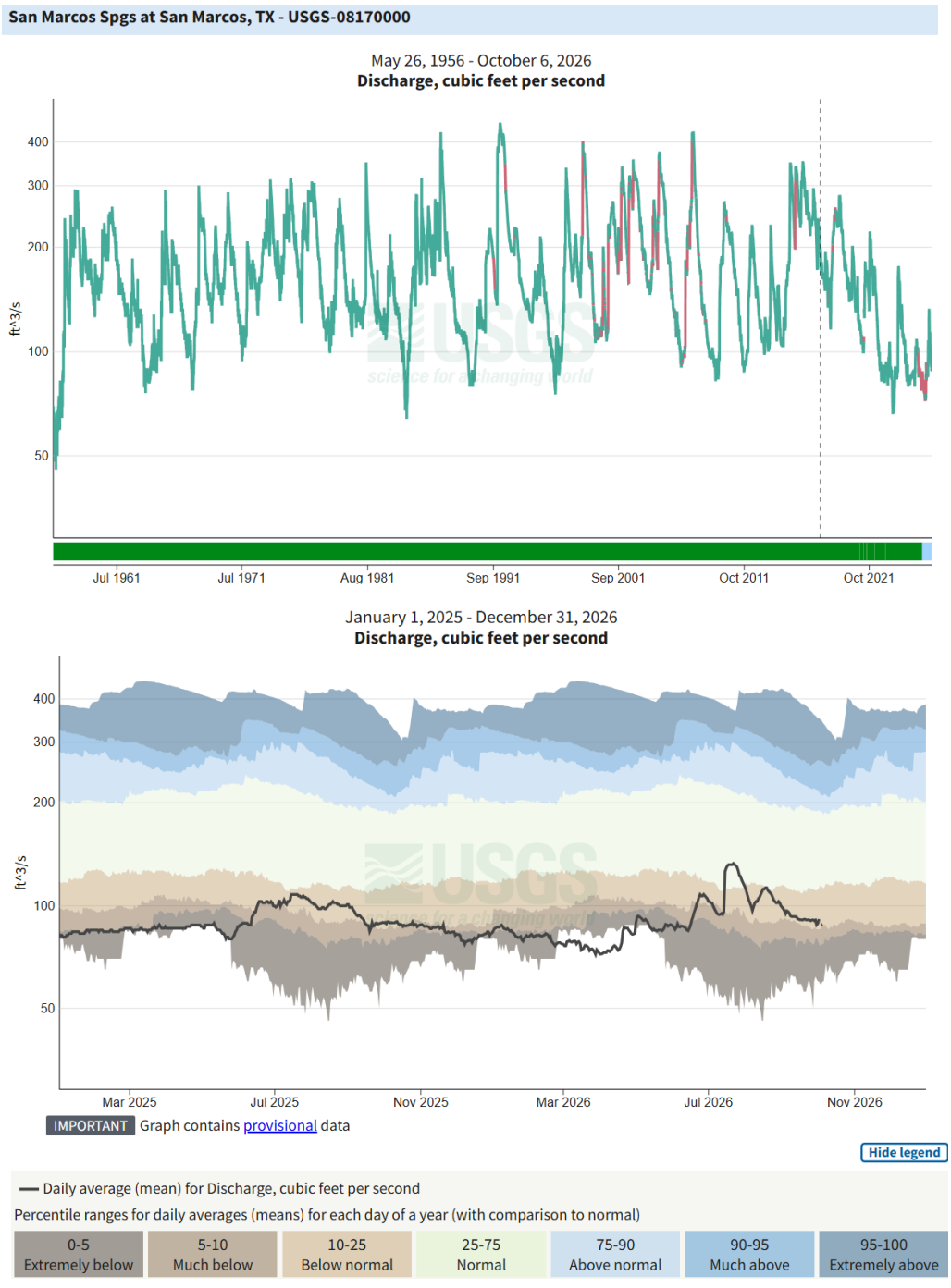
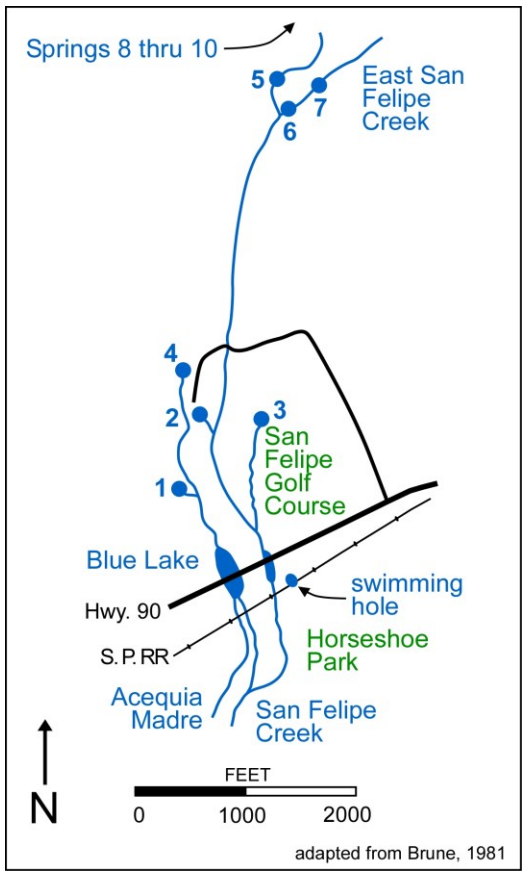


Figure 3.2.25 USGS graphs illustrating the ongoing demise of San Marcos Springs over the last 10 years or so. *Top*: Discharge rate for the period of record (May 1956 – October 2026), the average is 170 cfs. *Bottom*: New type of USGS graph showing daily average mean for the last water year, during which the discharge rate was below normal most of the time, or even off the chart for the *extremely below* category. The rate increased somewhat by the end of July 2026 due to two severe storms: on July 14-15 a powerful storm system dumped massive amounts of rain across the region, with radar and local weather reports indicating 7.5 to over 9 inches of rain falling right in the San Marcos area. The relentless downpour caused creeks like Purgatory Creek to rise rapidly, prompting city officials to shut down low-water crossings near downtown due to rising waters and heavy ponding on the roads. On June 11-12, 2026, severe storms pushed through Hays County, triggering flood watches, closing low-water crossings, and knocking out power to more than 2,000 utility meters in San Marcos.

San Felipe Springs, Val Verde County

San Felipe Springs, on the outskirts of Del Rio, are the third largest in Texas. There are a group of ten or more springs that extend for over a mile along San Felipe Creek on the grounds of the San Felipe Country Club and on several ranches to the north (Fig. 3.2.26). A groundwater divide near Brackettville separates this portion of the Edwards Aquifer from the central portion and the larger springs in San Marcos and New Braunfels.

In 1849, Captain S.G. French described San Felipe Springs and correctly noted they are the most western of a whole series of outflows from the Edwards. He wrote (French, 1850):



To the north of the road, and half mile distant, there is a beautiful spring of water, fifty feet in diameter at the surface, the sides of which incline towards a centre, like an inverted cone, and then, sinking in a cylindrical form to a depth of twenty-eight feet, through a soil of hard clay, afford a passage for the water to rise. The water comes to the surface with a slight ebullition, and flows off in a volume that would fill a cylinder two feet in diameter. This spring is the source of the San Felipe; as it flows on, the volume of its water is increased by other large springs, on either side; until it becomes a creek, when it empties into the Rio Grande, eight miles below the crossing, some thirty feet wide and several feet deep. Near its junction with the Rio Grande, its banks are shaded with large groves of pecan, maple, elm, and mulberry trees.

Figure 3.2.26 Map of San Felipe Springs based on Brune, 1981. Courtesy of Gregg Eckhardt.

Hill and Vaughan (1896) give the following description of the springs:

The westernmost of the line of fault springs are the San Felipe springs near Del Rio. They break out at the edge of the Edwards Plateau, 2 miles northeast of Del Rio and about 5 miles from the Rio Grande. The pool is almost as large as that at the head of the San Antonio River. From the deep-seated rock at its bottom the water can be seen welling up in a great column, and has the same peculiar greenish-blue color as that of the other streams of this class. No trees surround it; it is alone - a fountain in the desert. The rocks from which it bursts - the Fort Worth limestones - have the same kind of joints and faults as are found at San Antonio and Austin. The outflow from the pool forms a bold, rushing stream that runs off to the Rio Grande, some 5 miles distant. The spring stream, in addition to running a mill and supplying the village with water, is partially utilized to supply 15 miles of irrigation ditch and to irrigate 5,000 acres, and can furnish water for the irrigation of several thousand acres more. Mr. Babb's measurements make a total discharge of 19 second-feet, or about 12,000,000 gallons per day.

Although much of the area downstream from the Springs is now urban, about 3,000 acre-feet per year are still delivered via the irrigation canals and used for agriculture. Swimming holes at Horseshoe Park and Lions Park are very popular cooling-off spots.



Figure 3.2.27 *Left*: Spring #2, circa 1906 A very rare *Real Photo* postcard of Spring #2 before the municipal water supply pumphouse was added. Real Photo postcards were introduced in 1906 by Eastman Kodak and made it possible for anyone to have a postcard made using their own photograph. As such, they were usually produced in very limited numbers. The style of the back of the card dates it to pre-1907. *Right*: Postcard view of the pumphouse at Spring #2, circa 1930s (courtesy of Gregg Eckhardt, https://www.edwardsaquifer.net).



Figure 3.2.28 *Left*: Blue Lake, early 1990s. Water leaving Blue Lake, which is fed by Springs 1, 2, and 4. Discharge from the Lake is split: water flowing to San Felipe Creek is running over the spillway on the right, and the Acequia Madre Canal is on the left. Photo courtesy of Gregg Eckhardt. *Right*: San Felipe Springs that run alongside the 2nd hole of San Felipe Golf Course. Courtesy of City of Del Rio.

Historical Marker for the Canals System of Del Rio says:

Crude irrigation systems, drawing water from San Felipe Springs and Creek, were first devised by Indian and Spanish inhabitants of this area. Anglo-American settlers also saw the need for irrigation in this arid region, and about 1869 a group of landowners formed the San Felipe Agricultural, Manufacturing & Irrigation Company. Among early stockholders were W. C. Adams, Donald Jackson, Joseph Ney, Randolph Pafford, James H. Taylor, and A. O. Strickland. They dammed San Felipe Creek just below the Springs, and by 1871 had built canals diverting water to 1,500 acres of land.

In addition to promoting agricultural development, the work of the San Felipe Company stimulated the growth of Del Rio, since the irrigation canals provided water to the city as well. Today this vital water supply system is still in operation.

Barton Springs, Travis County

On the *Visit Austin* web page (https://www.austintexas.org/listings/barton-springs-pool/4687/), there is the following description of Barton Springs: "Within Zilker Park's 358 acres lies one of the crown jewels of Austin – Barton Springs Pool. The pool itself measures three acres in size and is fed from underground springs with an average temperature of 68-70 degrees, ideal for year-round swimming. Over the years, Barton Springs Pool has drawn people from all walks of life, from legislators who have concocted state laws there to free-spirited, topless sunbathers who turned heads in the 1970s. Robert Redford learned to swim at the pool when he was five years old while visiting family in Austin. Today, Barton Springs still attracts a diverse crowd of people and has seen record-setting numbers of visitors nearing 800,000 in recent years."

Barton Springs is the largest natural discharge point of the Barton Springs segment of the Edwards Aquifer (see Fig. 2.2.9 in Chapter 2.2). It consists of four major outlets, the largest discharging directly into Barton Springs Pool, a major recreational attraction of the City of Austin. Each of the spring outlets provides habitat for the federally-listed endangered Barton Springs Salamander.

As described by Smith et al. (2009), the formation of the aquifer was influenced significantly by fracturing and faulting associated with the Miocene-age Balcones Fault Zone (BFZ) and dissolution of limestone and dolomite units by infiltrating meteoric water. Faults trend predominantly to the northeast and are downthrown to the southeast, with total offset of about 1,100 ft across the study area. Dissolution along fractures, faults, and bedding plane partings and within certain lithologic units has created numerous sinkholes, sinking streams, springs, conduits, and caves.



Figure 3.2.29 Aerial photograph of the Barton Springs Pool. From Smith et al., 2009.

Groundwater generally flows west to east across the recharge zone, converging with preferential groundwater flow paths subparallel to major faulting, and then flowing north toward Barton Springs. Rates of groundwater flow along preferential flow paths, determined from dye tracing, can be as fast as 4 to 7 mi/day under high-flow conditions or about 1 mi/day under low-flow conditions (Hauwert et al., 2002).

It is estimated that 85% of recharge to the aquifer occurs along its six major (ephemeral) losing streams that cross the recharge zone, and the remaining recharge occurring in the upland areas of the recharge zone (Slade et al., 1986). The amount of cross-formational flow (sub-surface recharge) occurring through adjacent aquifers is unknown, although it is thought to be relatively small on the basis of water-budget analysis for surface recharge and discharge.

Unfortunately, behind the upbeat description of the Barton Springs on the *Visit Austin* web page, there is a grim reality described by Gregg Eckhardt on his web page (https://www.edwardsaquifer.net/barton.html). Below are only few of the alarming excerpts that may inspire some (or many) readers of this book to act and speak up for Barton Springs and all other springs in Texas.

Citizens have been fighting for decades against insensitive development that threatens Barton Springs. In the 1990s, residents overwhelmingly passed a Save Our Springs ordinance that would have implemented strict development controls. It was subsequently nullified by the state legislature, which passed a law allowing any development plat already on file to be completed without regard to the new controls.

In 2006, the United States Geological Survey published a Scientific Investigations Report that summarized water quality sampling performed from 2003 to 2005. Barton Springs was found to be affected by persistent low concentrations of atrazine (an herbicide), chloroform (a by-product of drinking water disinfection), and tetrachloroethane (a solvent).

In 2008, the fight to preserve Barton Springs was the subject of The Unforseen, a documentary co-produced by Robert Redford, who learned to swim there as a child. The movie uses the struggle over development in the Barton Creek watershed to illustrate the many clashes between private property rights and resource protection that are occurring across the country.

In January of 2019 the University of Texas at Austin released a study that showed 13 central Texas salamanders are in danger of `extinction due to overexploitation of groundwater. These include three at Barton Springs: the Austin Blind Salamander, the Jollyville Plateau Salamander, and the Barton Springs Salamander. In an interview for The Daily Texan, research fellow Thomas J. Devitt said urbanization has caused a decrease in the salamander populations. He said "Central Texas has some of the fastest-growing cities and counties in the nation, putting a strain on natural resources, especially water, and the environment. Overpumping of groundwater has caused (aquifer water levels) to drop and springs to stop flowing. This causes habitat loss for the salamanders."



In June of 2019 an environmental group, the Center for Biological Diversity, filed a lawsuit against the Fish and Wildlife Service, claiming federal officials are failing to protect the Georgetown and Salado salamanders. The group alleges the federal government has not taken steps to protect the habitats of the two salamanders since they were first listed as threatened in 2014.

Figure 3.2.30 Barton Springs pool, Austin, Texas. Courtesy of Austin Convention and Visitors Bureau.

Today, Austin's natural beauty is a shadow of its former self, and the future of Barton Springs seems dim. Texas is struggling to find the right balance between protection of common resources and private property rights, and currently the pendulum is swinging strongly in favor of landowners, not residents.

Gregg Eckhardt

Obituary

In some way, one is compelled to make an analogy between a dry Texas spring and a dead canary in a coal mine. Except that, in the case of Texas, there are already lots of dead canaries but no one in power, either politically or financially, seems to care about the inevitable explosion in the mine and what will happen to the hard-working miners in it. And of course, everyone knows that Texas has quite a few billionaires.



An example that ties it all together, i.e., Texas aquifers, springs, billionaires, and the hard-working miners (people) of Texas is provided courtesy of various local media including Channel 8 abc news (Dallas's Leading Local News: Weather, Traffic, Sports and more | Dallas, Texas | wfaa.com), as well as newsletters of big law firms in Dallas.

A recent high-stakes legal battle over East Texas groundwater rights is talk of the town in Texas. It is between prominent Dallas investor Kyle Bass, whose real estate entities filed a federal civil rights and property-taking lawsuit on July 7, 2026, and the Neches & Trinity Valleys Groundwater Conservation District.

The core of the dispute centers around a temporary moratorium enacted by the district on May 21, 2026, which blocks new non-exempt drilling, operating, and water-transfer permits. Bass's companies—Redtown Ranch Holdings LLC and Pine Bliss LLC—allege that the moratorium unconstitutionally targets them and effectively "confiscates" their private property rights to the water beneath their combined 11,500+ acres of land across Anderson and Henderson counties.

Filed in the Tyler Division of the U.S. District Court for the Eastern District of Texas, the lawsuit argues that according to the Texas Constitution, landowners own the groundwater beneath their property. Bass's legal team asserts that by freezing permits, the government is executing a regulatory "taking" of private property without just compensation.

Local officials, including Crockett City Administrator John Angerstein, have warned that Bass's broader plans could eventually extract anywhere from 10 billion to 16 billion gallons of water annually from the Carrizo-Wilcox Aquifer.

Critics argue that none of this water is designated for the local rural communities. Instead, they allege it is a speculative venture intended to export water away from East Texas to support heavy urban development and sprawl in the Dallas-Fort Worth metroplex or the Texas Hill Country. (The author of this book adds hundreds of AI data centers, lured to Texas by its current governor Greg Abbott and the government, to the list of likely users of the enormous volume of groundwater to be extracted.)

What all this means is that, once the aquifer is depleted (mined), all that is left to the Texans that depend on it is to wait and see what will happen with the Great Master Plan. This grandiose master plan is already being circulated by the rich and powerful in Texas, and outside Texas: build a network of state-wide water pipelines, desalinate sea water from the Gulf, desalinate highly mineralized, deep groundwater found everywhere in the state of Texas, and transfer all that water around as needed. And naturally, all this will be paid for by the taxpayer's money.

In the meantime, all Texas springs will, sooner rather than later, be dead.

In concluding this obituary for Texas springs and groundwater resources, the author reminds readers of a sad truth: history repeats itself, and humankind never learns. Excerpts below are from the anthological work by Gunnar Brune published in 1975 and titled *Major and Historical Springs of Texas*.

Springs were vital to the survival of Texas' earliest inhabitants, over 30,000 years ago. At an archeological site near Lewisville in Denton County, radiocarbon analysis has dated the remains of these early new-world humans at 37,000+ years old, including crude sculptures, spears, and spear throwers (Newcomb, 1961). These early Americans always made their campgrounds near water, whether it was a spring, spring-fed stream, a river, or a lake. Bedrock mortars or rock mills were worn into the rock by the Indians as they ground stool, acorns, and other nuts, mesquite beans and grain. These mortars can still be seen at many Texas springs. It is also noteworthy that the Pueblo Indians of west Texas used spring water for irrigation of crops long before the arrival of the Europeans (Taylor, 1902; Hutson, 1898).

Because the springs were so vital to the life of both the Indians and the white man, it is not surprising that many battles were fought over their possession. In 1650 when Spanish explorers first visited Big Spring in Howard County, they found the Comanche and Pawnee Indians fighting for its possession. When a network of forts was strung across Texas, they were, in nearly all cases, located near springs in order to have a reliable supply of pure water. Later the covered-wagon and stagecoach routes came to rely heavily upon springs. For example, the "Camino Real" or King's Highway, completed by the Spanish colonists about 1697 from Natchitoches, Louisiana, to San Antonio and Mexico, passed 13 major Texas springs and many more minor ones.

Many of the early settlements relied entirely on spring water. At least 200 towns were named for the springs at which they were located. About 40 still are shown on the official Texas State Highway Map, but many of the springs have dried up.

Throughout the long period during which various Indian tribes occupied Texas, spring flow remained unchanged except as affected by wet and dry climatic cycles. At the time of Columbus' epic voyages Texas abounded with springs which acted as natural spillways to release the excess storage of underground reservoirs. Early explorers described them as gushing forth in great volume and numbers. The very early accounts usually describe not springs but "fountains". This is an indication of the tremendous force with which these springs spouted forth before they were altered by modern man. As an example, less than 100 years ago Big Boiling Spring, one of the Salado Springs (Bell County) was still described as a fountain rising 5 feet high. Such natural fountains ceased to exist in Texas many years ago.

Probably the first effect upon ground-water tables and spring flow was the result of deforestation by the early white settlers. Deforested land was placed in cultivation or pasture. The deep open structure of the forest soils was altered as the organic matter was consumed and the soils became more impervious.

In the middle 1800's deep wells began to be drilled. It was found that flowing wells could be brought in nearly everywhere.

Although the effects of flowing wells upon spring flow were severe, there was more to come. When the wells ceased flowing, pumping began. Ground-water levels were systematically drawn down, as much as 700 feet in some areas. At first pumping for municipal and industrial use was primarily responsible. In recent years tremendous quantities of ground water have been withdrawn for irrigation, amounting to about 80 percent of the total ground water used in Texas. As a result, some streams which were formerly "gaining" streams, receiving additional water from streambed seeps and springs, are now "losing", and many streams have ceased flowing. Thousands of small springs have dried up, and the larger springs have generally suffered a decrease in flow.

3.3 Montana, Wyoming, and South Dakota

3.3.1 Montana

The locations of four large springs in Montana described by Meinzer in his seminal work *Large Springs in the United States* (1927) are shown in Fig. 3.3.1, together with three other notable, non-thermal springs: Blaine, Armstrong, and Nelson Springs. What they all have in common, except possibly for Armstrong and Nelson Springs, is that they drain the deep-seated carbonate Madison Aquifer which is recharged in the highland areas and plunges

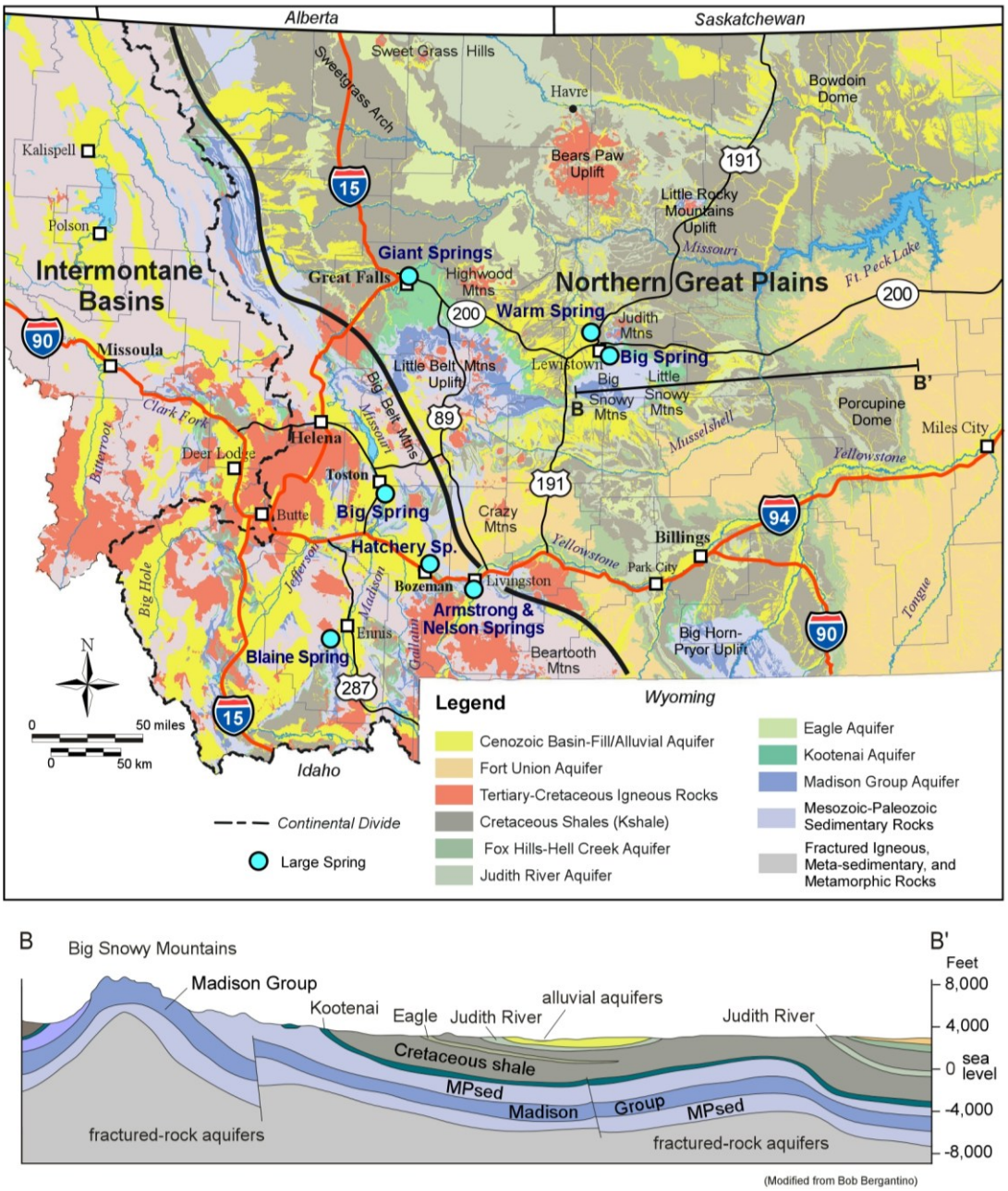


Figure 3.3.1 Locations of large springs in Montana. Aquifer map and cross-section from John I. LaFave, 2024: Principal Aquifers of Montana. MBMG Special Publication 122: Geology of Montana, vol. 2: Special Topics; available at https://mbmg.mtech.edu/pdf/geologyvolume/LaFavePrincipalAquifersFinal.pdf

below a series of thick, less permeable sediments that extend into the Northern Great Plains, with most of the discharge from the system occurring in eastern North Dakota and in the Canadian Province of Manitoba.

As discussed by LaFave (2024), the Madison Group aquifer consists primarily of Mission Canyon Limestone. The thickness of the Mission Canyon ranges from 300 to 600 ft (Miller, 1981). In central Montana, the Madison Group crops out mainly along the northern flanks of the Little Belt and Big Snowy Mountains. It is also prominently exposed along the northeast flank of the Pryor Mountains in the Big Horn-Pryor Uplift, and in narrow exposures in mountain ranges in southwest Montana. The Madison typically dips downward steeply away from the mountain fronts and decreases in slope with distance from them. It is deeply buried (5,000–8,000 ft below the surface) under most of eastern Montana. The limestone is relatively impermeable; however, where fracturing and karst features are well developed, there is the potential for large well yields and fast advective flow paths. The permeability and well yields are variable; in places closely spaced wells have very different reported yields.

In Cascade County, near Great Falls, the Madison is about 350 ft below the land surface; however, it is exposed about 25 mi to the south–southeast along the north flank of the Little Belt Mountains (Smith, 2008; see Fig. 1.55 in Chapter 1.2). More than 900 wells (roughly 75 percent of all Madison wells statewide) use the Madison aquifer, mostly between the Little Belt Mountains and the Missouri River near Great Falls. Several streams lose flow to the Madison aquifer in outcrop areas, and groundwater moves from the mountain recharge areas, in the Little Belt Mountains, towards the Missouri River, including the largest spring in Montana, and one of the largest springs in the country, the Giant Springs near Great Falls (Feltis and Shields, 1982; Madison, 2016).

Chlorofluorocarbons (CFC) age dating of Madison aquifer water at Giant Springs and a nearby Madison well returned an apparent age of 23 and 26 years respectively, suggesting that flow rates between the Little Belt Mountains (recharge area) and the Springs and the Missouri River (discharge area) could be as much as 15 ft per day (LaFave, 2012).

Four of the largest springs in Montana, which drain the Madison, supply water for fish hatcheries: Giant Springs in Cascade County near Great Falls, Big Spring in Fergus County near Lewistown, Hatchery Spring in Gallatin County near Bozeman, and Blaine Spring in Madison County near Ennis. They all have a consistent flow of relatively low-TDS water with similar water chemistry (LaFave, 2024).

Giant Springs, Cascade County

Giant Springs were discovered by Captain Clark, of the Lewis and Clark Expedition. On June 18, 1805, Clark described Giant Springs as "*the largest fountain or Spring I ever Saw, and doubt if it is not the largest in America Known, this water boils up from under th rocks near the edge of the river and falls imediately into the river 8 feet and keeps its Colour for ½ a mile which is emencely Clear and of a bluish Cast*."

On June 29, Lewis and Drouillard hiked six miles from their upper portage camp to visit the spring after hearing Clark's description. They found it "*much as Capt. C; had discribed & think it may well be retained on the list of prodegies of this neighbourhood towards which, nature seems to have dealt with a liberal hand*." (National Park Service, https://www.nps.gov/places/giant-springs.htm).

As explained by Cassius A. Fisher of USGS in 1909, "*According to measurements made by E. T. Nettleton* (in 1892) *the flow of these springs is approximately 638 cubic feet per second, an amount which, converted into gallons, is the equivalent of over 400,000,000 gallons every twenty-four hours a veritable underground river. The fact that the water of Giant Springs issues from rocks at the water's edge and in the bed of the river renders it difficult to measure their exact flow. In order to ascertain this amount, measurements were taken of the total flow*

of Missouri River above and below the springs; the difference between these two measurements is assumed to be the quantity furnished to the river by the springs. It is readily seen from the above figures that these springs rank among the largest in the United States. The water, which boils up with considerable force, is clear, blue, and relatively pure, containing no more dissolved salts than the average well water of the region. It has a temperature of about 50° F.

No spring deposits occur in the immediate vicinity of Giant Springs, and the water is not generally regarded as possessing therapeutic value. It is not utilized at present, but is allowed to flow into the river. There are, however, a few improvements, such as sidewalks, etc., which make it possible for tourists to view the springs from the most advantageous points."



Figure 3.3.2 Giant Springs. Photo by Alex Wiles, National Park Service.

Unfortunately, no similar measurements were conducted since or made publicly available to the best of author's knowledge. However, based on various accounts over the years, the spring discharge rate does not seem to change significantly from season to season and is always described as powerful. Whatever the case may be, Giant Springs, if not the largest, likely belongs to the group of 5 largest springs in the country.

Hydrogeologically, it is a unique spring draining the deeply seated karst aquifer in Madison Limestone. This ascending spring emerges from the large joint planes in a medium to coarse grained Kootenai sandstone which overlies Morrison Formation and Madison Limestone (see Fig. 1.55 in Chapter 1.2). For a short distance on either side of the main spring there are smaller springs flowing from joint planes. In the bed of the Missouri River, there is another large spring which is only apparent during low water seasons. The spring recharge area is 30-40 miles to the south where the Madison Limestone outcrops on the Little Belt Mountains.

As discussed by LaFave (2012), samples from Giant Springs and a nearby Madison well were analyzed for environmental isotopes to determine apparent groundwater age and residence time. Chlorofluorocarbons (CFC's) and tritium were detected, and the results returned "CFC ages" of 23 and 26 years, respectively. The ages are consistent with the tritium values of 10.1 and 10.2 TU. Radiocarbon dating of samples from the well and Giant

Springs returned values of close to 50 Percent Modern Carbon (47.48 and 49.38 PMC, respectively), reflecting that half of the DIC is derived from atmospheric CO_2 with modern ^{14}C activity, and half from the carbonate ($CaCO_3$) that has no ^{14}C activity.

Giant Springs is in Giant Springs State Park, the most visited state park in Montana. The springs are a source of the Roe River (only 201 feet long, before it empties into the Missouri River), once listed as the shortest river in the World by the Guinness Book of World Records before Guinness eliminated the category. The park hosts a trout hatchery, established in 1928, which uses about 15,000 gallons of water per minute by diversion from the springs. Water temperature is constantly 54 degrees Fahrenheit. A drinking water bottling company uses water sourced directly from the springs via a protected intake.



Figure 3.3.3 Drone photo of Giant Springs discharging into the muddy water of the Missouri River. Courtesy of Giant Springs State Park.

Big Spring, Fergus County

The following description of Big Spring, which is near the Town of Lewistown, is from the Town's' website (https://enjoylewistown.com/big-spring-creek).

"Big Spring Creek originates seven miles southeast of Lewistown where it bubbles out of the ground from one of the world's largest fresh water springs. Historically known as a sacred site to Native American tribes, and later recorded as a valued resource for Lewistown's early settlers, Big Spring Creek remains one of Lewistown's best kept secrets. This artesian spring emerges from the Madison-Limestone aquifer formation on the north slopes of the Big Snowy Mountains and pumps from 50,000 to 64,000 gallons per minute (equivalent of 110 to 134 cfs) of 52-degree water, all year round. From here, Big Spring is diverted in part, and serves as a state fish hatchery in which thousands of brown, rainbow and cutthroat trout are raised annually.

From it's origin near the state fish hatchery, the creek runs almost thirty miles before joining the Judith River, a main tributary of the Missouri River. The creek flows through and under the City of Lewistown for a few city blocks, via a man-made channel.

Big Spring Creek is a stream of extraordinary beauty and is a superb fishery, known for its year-round, red-ribbon fly-fishing. Montana Fish, Wildlife & Parks maintains six seasonal public fishing access sites along Big

Spring Creek. Lewistown is one of the few towns of its size where you can often catch rainbow and brown trout from within the city limits.

Aside from the world class fly-fishing and recreational opportunities, Big Spring is the main water source for the City of Lewistown, providing crystal clear, mineral-rich water to 6,000+ city residents, with little purification needed. This crystal-clear mineral water is also bottled and shipped throughout the country.

The largest cold-water trout hatchery in the state, Big Spring Trout Hatchery is located seven miles southeast of Lewistown next to Big Spring, the originating point of Big Spring Creek. The public can view fish in the hatchery raceways but are encouraged not to feed or bother them."





Figure 3.3.4 *Left*: Big Spring near Lewistown, Montana. Photo by Mike Cline taken on 16 August 2015, licensed under the Creative Commons Attribution-Share Alike 4.0 International license. Available at https://commons.wikimedia.org/wiki/File:Big_Spring_Creek_Lewistown_Montana_29.JPG. *Right*: Trout Hatchery at Big Spring. Photo courtesy of Montana Fish, Wildlife & Parks.

Warm Spring, Fergus County

Warm Spring is located 15 miles northwest of Lewistown on the Denton Highway (driving directions: 9.5 miles north of Lewistown on US Highway 191, turn left at State Highway 81 and drive about 5 miles, turn right at the Warm Springs sign). It is the principal source of Warm Spring Creek, a tributary of Judith River.

According to Meinzer (1927), the first reported discharge measurement of the spring was 140 cfs in 1907, followed by 180-190 cfs in 1923. The water was used in part by the Barnes-King Development Co. for power in mining and milling operations, and some of the water was also used for irrigation. Today, the spring is on a private ranch owned by Vanek Family since 1940 and feeds a large 5,000-square-foor pool of water at constant temperature of 68°F (20°C). "*On a typical summer day, you can find folks of all ages swimming, picnicking, playing volleyball or horseshoes, and fishing in Warm Spring Creek. You can relax in the shade of one of many willow shoots that surround the spring. Warm Springs charges $4 for age 14 and over, $3 for ages 6- 13 and age 5 and under are free. Sorry no dogs allowed!*" (https://enjoylewistown.com/exploring-lewistown-montana/wide-open-places/warm-springs).

Several opinions on the spring's origin and the unusually high (for non-thermal springs in Montana) water temperature of 68°F are provided by Meintzer (1927) and Calvert (1909). In the light of newer, now widely accepted

understanding that this first-magnitude spring drains the deeply seated Madison limestone aquifer, the opinion by Calvert that the temperature is not result of recent magmatic activity, seems right on target:

If the spring does represent the last phase of such activity, however, it would seem highly improbable that it should be the only one in the entire district whose temperature is noticeably above the average. Likewise, it would seem strange that not even this spring is depositing lime. In fact, field analysis of its waters shows that they contain no more than the average of carbonates for that region. It might also be argued that the temperature is a result of recent faulting, and the fact that the spring is directly on a fault line gives weight to that assumption. However, numerous other faults, apparently contemporaneous, occur in the region, and since springs near them are not affected as regards temperature, it would seem that some other cause should be sought to account for the exceptional temperature of the water of Warm Spring. A much more probable theory regarding the elevated temperature of the spring water is that it has come from a considerable depth.



Figure 3.3.5 Photo from an article by Elizabeth Hawley *Guide to the Best Gigantic Warm Springs Experience* published in Pocket Montana, updated on March 13, 2024.

3.3.2 Wyoming

Wyoming has diverse karst areas underlined by limestone, dolomite, and gypsum. While the state is famous for its wide-open basins and dramatic granite peaks, the karst is equally fascinating and includes both classic carbonate karst and evaporite karst.

The primary driver of Wyoming's karst is the Paleozoic Madison Limestone, exposed by tectonic uplift along mountain ranges. The state has a number of world-class limestone karst features including:

- Thermal karst springs in the Yellowstone National Park.
- Thermal karst springs in Thermopolis
- The largest periodic (“ebb-and-flow”) karst spring in the World, *Periodic Spring,* near Afton.
- Natural Trap Cave, just northeast of the town of Lovell in Big Horn County, contains one of the most complete Pleistocene-to-Holocene fossil records in North America.

- Bighorn Cavern, a massive cave system with over 14 miles of mapped passages located within the Bighorn Canyon National Recreation Area, and extending into Montana.
- Sinks Canyon State Park near Lander, home to Wyoming's most famous sinking river and river rise.
- Big Spring Scenic Backway in Lincoln County, named after one of the most scenic karst springs in the West.

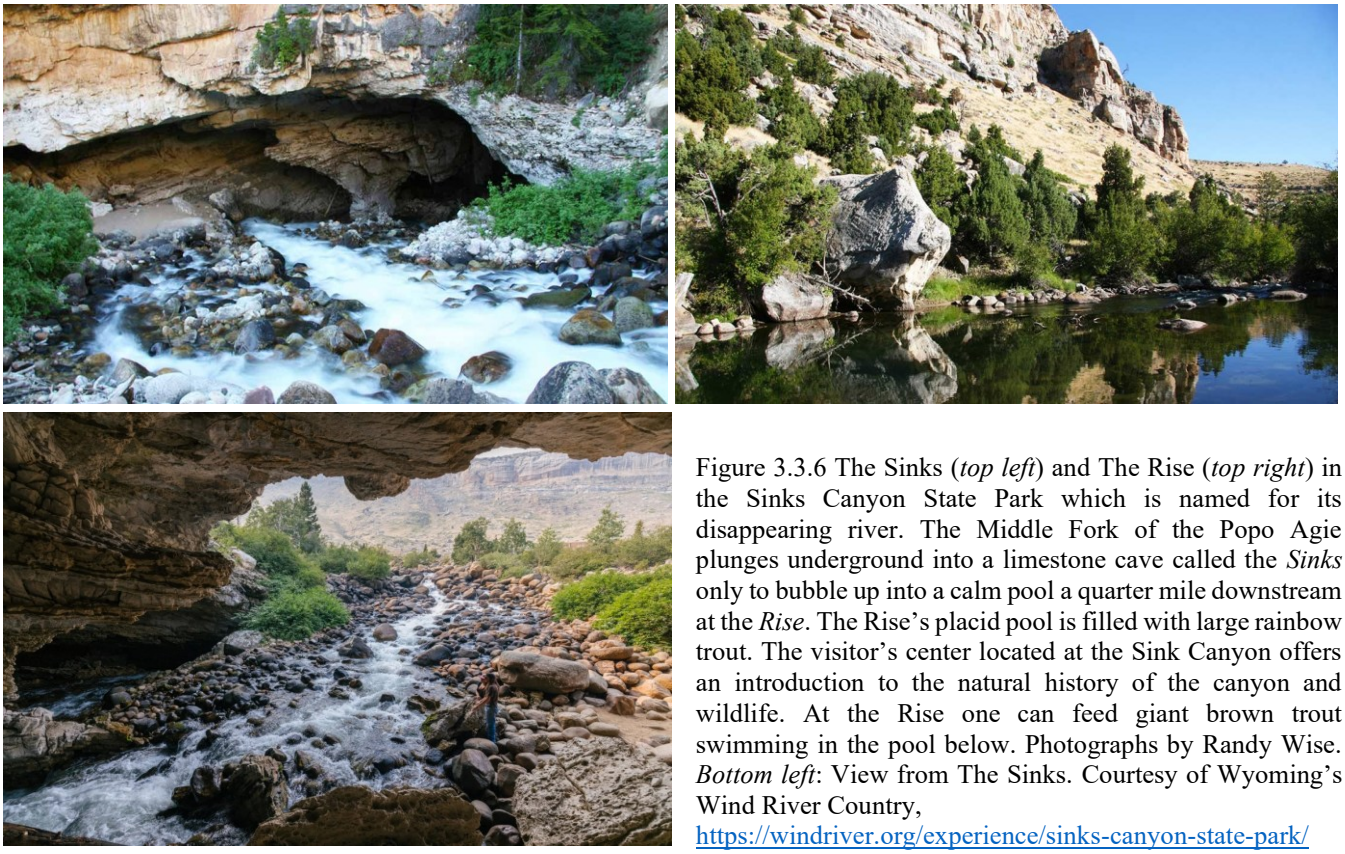


Figure 3.3.6 The Sinks (*top left*) and The Rise (*top right*) in the Sinks Canyon State Park which is named for its disappearing river. The Middle Fork of the Popo Agie plunges underground into a limestone cave called the *Sinks* only to bubble up into a calm pool a quarter mile downstream at the *Rise*. The Rise's placid pool is filled with large rainbow trout. The visitor's center located at the Sink Canyon offers an introduction to the natural history of the canyon and wildlife. At the Rise one can feed giant brown trout swimming in the pool below. Photographs by Randy Wise. *Bottom left*: View from The Sinks. Courtesy of Wyoming's Wind River Country, https://windriver.org/experience/sinks-canyon-state-park/

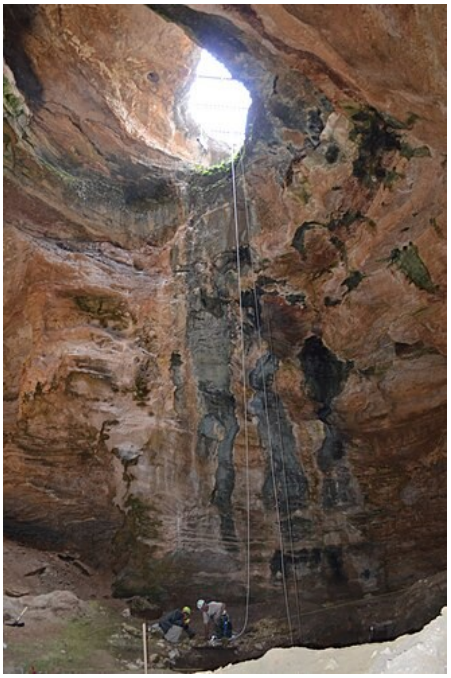


Figure 3.3.7 Natural Trap Cave is an 85-foot-deep, bell-shaped limestone pit with a narrow 15-foot-wide opening at the top. Because it is flush with the ground, it is virtually invisible, acting as a natural pitfall for wildlife for over 40,000 years. The cave itself is strictly closed to the public to preserve its world-class paleontological resources. A heavy metal grate covers the entrance to prevent people and wildlife from falling in. Because the cave remains a constant refrigerator-like temperature (under 42°F), it has preserved a massive cache of Ice Age fossils—including mammoths, short-faced bears, American cheetahs, and lions. Photograph by Brent Breithaupt, courtesy of Bureau of Land Management.

Mammoth Hot Springs, Yellowstone National Park

As explained by the scientists of USGS (Bargar, 1978; White et al., 1975; Sorey et al., 1991; Morgan et al., 2017; and Shanks, 2021), Mammoth Hot Springs is the only major thermal area in Yellowstone National Park dominated by deposition of hydrothermal $CaCO_3$ (travertine) rather than siliceous sinter. Samples of hot spring fluids at Mammoth Hot Springs, including from USGS drill hole Y-10, show that measured near-surface temperatures never exceed 73°C (163.4°F), and the chemistry of the water suggests that subsurface fluid temperatures have not exceeded 100°C (212°F). Compared with alkaline-chloride waters, like those from Old Faithful, thermal waters at Mammoth Hot Springs are rich in calcium, magnesium, bicarbonate, and sulfate.

The alinement of young volcanic rocks along a fault zone between Norris Geyser Basin and Mammoth Hot Springs zone suggests that the thermal water that eventually reaches Mammoth may be heated by partly molten magma within the fault zone. During transport along the fault, the water is enriched in calcium and bicarbonate, the two ingredients necessary for the formation of travertine. The trend of White Elephant Back Terrace and the tension fractures to the southwest of this fissure ridge suggest that the thermal water moves up through the old terrace deposits along preexisting vertical linear planes of weakness. As the water reaches the surface, pressure is released, carbon dioxide escapes as a gas, and bicarbonate in the water is partitioned into more carbon dioxide and carbonate; the carbonate then combines with calcium to precipitate pure white calcium carbonate, forming travertine of famous terraces of Mammoth Hot Springs.

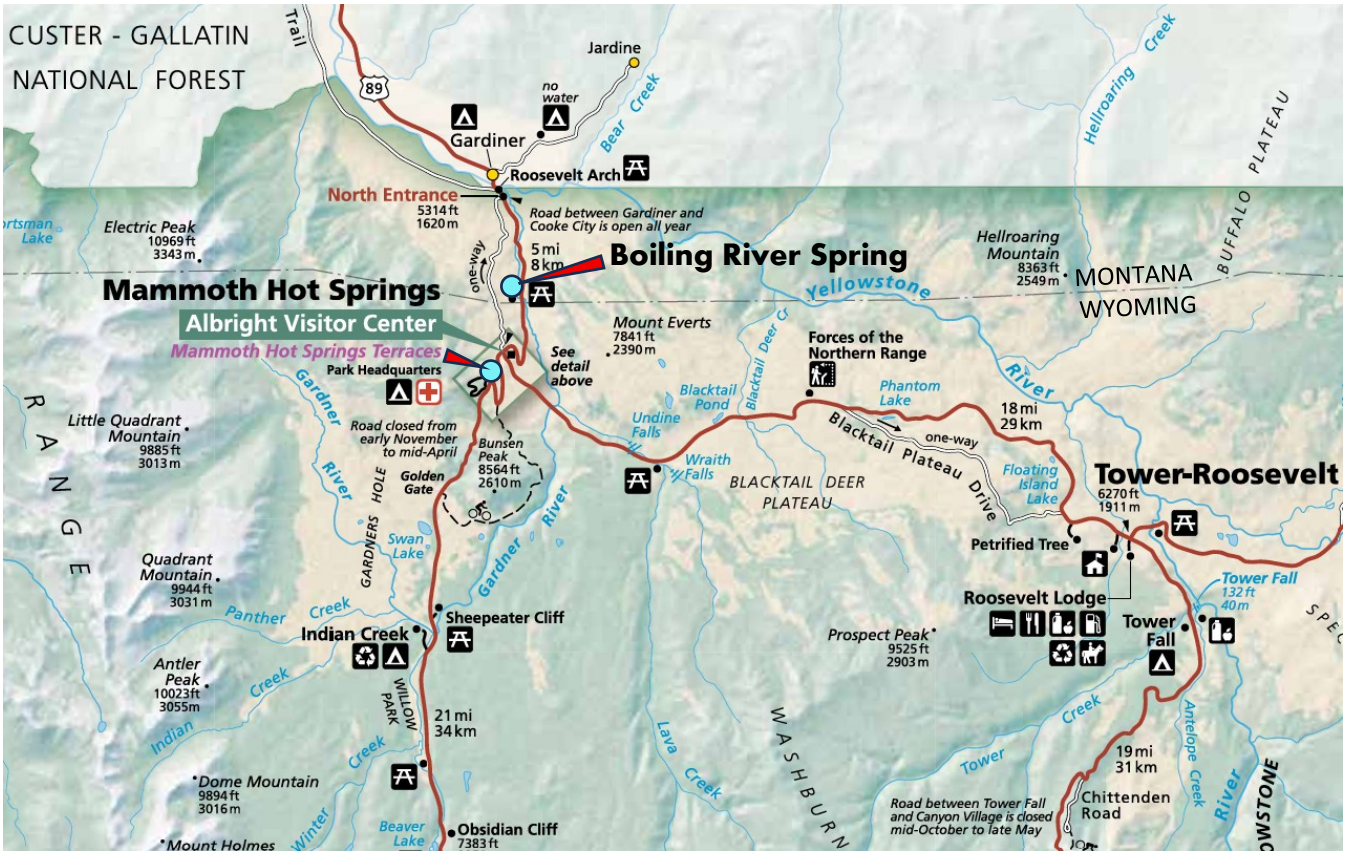


Figure 3.3.8 Part of the NPS' Yellowstone National Park map with locations of the only two thermal calcium carbonate type (karst) springs in the park: Mammoth Hot Springs, and Boiling River Spring. Fluids of all other thermal features and springs in the park are either alkaline-chloride or acid-sulfate type.

Sulfide-oxidizing Aquificales bacteria contribute to the bright yellow, orange, and red coloration of the pools and channels. When discharge of a hot spring ceases, the microorganisms die, and the travertine deposits become white again and then dingy gray. For hundreds of years, Shoshone and Bannock people collected minerals from Mammoth Hot Springs for white paint.

Most travertine has precipitated rapidly from solution and is lightweight and porous; however, a few fissure ridges contain denser, vertically banded travertine layers that line the fissure. These vertical bands were apparently deposited over a long period of time. Dense travertine also forms beneath the surface of the terraces by deposition in the pore spaces of older deposits.

With the passage of time, the surface travertine deposits develop a loose soil capable of supporting the growth of small plants and eventually large trees, such as occur on Pinyon Terrace.

The active Mammoth Hot Springs area is an elongate northeast-southwest depositional mound that hosts approximately 100 active hot springs. The total thickness of travertine deposits can be approximated using results from drill hole Y–10, which was drilled near Bath Lake in 1967 to 1968, to a total depth of 113 m (371 f). Travertine comprises the upper 77 m (253 ft) of the section, with Mississippian Madison Formation dolomitic limestones below the travertine.

Some hot springs are pools with little or no deposition of travertine, whereas others form magnificent arrays of terracettes, cone-shaped deposits, or linear mounds called fissure ridges. Fissure ridges occur in numerous sizes and shapes. In some places water from new hot-spring activity becomes ponded behind fissure-ridge barriers or darns, with the result that travertine deposits eventually form large Rat terraces.

Most of the thermal water from Mammoth Hot Springs re-enters the subsurface through karst terrain and re-issues along the Gardiner River in a series of springs. The largest of these is known as the Boiling River Spring.



Figure 3.3.9 Travertine terraces and deposits at Mammoth Hot Springs.

Boiling River Spring, Yellowstone National Park

In the general area of Mammoth Springs, the portion between Hotel Terrace and the Gardner River has few small thermal springs and one spring with a very large rate of discharge which is the source of Boiling River (historically also referred to as Hot River). The spring emerges from a small cave beneath an old, partially collapsed

travertine ledge near the level of the Gardner River (Fig. 3.3.10-*Left*). The discharge channel is about 9 feet wide and 2 feet deep, and the stream flows for about 400 feet before emptying into the river over a series of cascades (Fig. 3.3.10-*Right*). Unfortunately, in June 2022 Yellowstone floods shifted the geothermal vents and eliminated the cold-water mixing zone that previously made soaking safe. The hot spring water now flows at approximately 140 degrees Fahrenheit. Immersion in this water would cause severe and potentially fatal burns. The NPS has officially deemed the site unsafe with no reopening timeline as of August 2026.

The underground stream conduit can be followed upstream for an additional 420 feet through a series of collapse depressions features in travertine, one of which (MHS-1) contains visible flowing water.

According to Bargar (1978), the spring which feeds Hot River has the greatest discharge of any hot springs in Yellowstone National Park. Several discharge measurements reported by Allen and Day (1935) show that the flow of Hot River ranges from about 33 to 40 m^3/min (19.4 to 23.5 cfs), with the average flow being slightly greater than 38 m^3/min (22.4 cfs). A measurement by USGS in 1967 yielded about 41 m^3/min. Allen and Day also suggest that the variation in temperature of Hot River from an average of about 71°C (160°F) at different time of the year may be caused by dilution with cold meteoric water.



Figure 3.3.10 Pre-2022-flood photographs of Boiling River Spring. *Left*: Spring orifice. *Right*: Boiling River, seen in the center left, used to flow for about 400 feet before emptying into Gardiner River over a series of cascades. Boiling River water is too hot (~160°F) for direct bathing, so the visitors were soaking in the mixing zone with cold water of Gardiner River. This was only one of two places in Yellowstone National Park where bathing and swimming were allowed.

As discussed by USGS authors in a publication edited by Sorey (1991), hot-spring discharge from the Mammoth hydrothermal system totals 590 L/s (20.8 cfs), approximately 10 percent of which flows from springs in the Mammoth Terraces. Most of the remaining 90 percent flows into the Gardner River from Boiling (Hot) River Spring. Mammoth thermal water is derived from some combination of lateral flow moving northward in the Norris-Mammoth corridor and deep circulation originating from more local sources (for example, the Gallatin Range to the west). The main feature of hydrothermal system at Mammoth Hot Springs is flow through pre-Cretaceous sedimentary rocks in the Gardiner syncline where reservoirs for thermal fluids and sources of dissolved chemicals are most likely provided by Paleozoic carbonate rocks, such as the Mississippian Mission Canyon Limestone which is part of the upper Madison Formation Limestone.

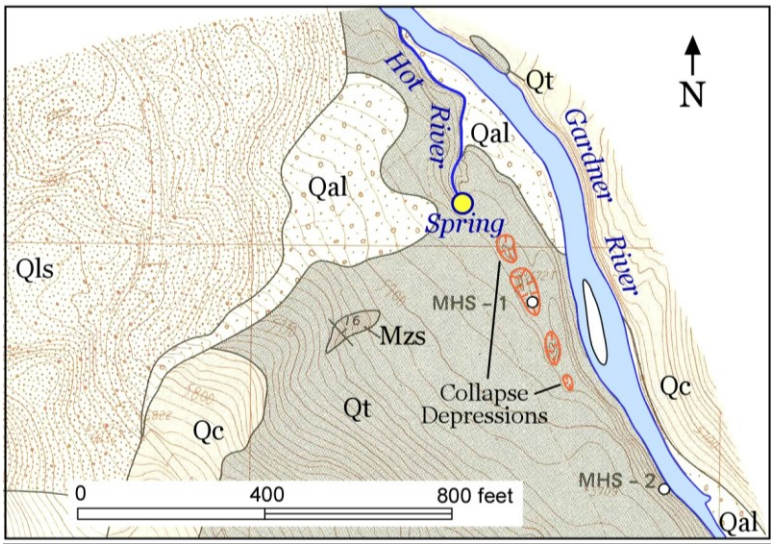


Figure 3.3.11 Geologic map of the Hot River Spring area. From Bargar, 1978 (part of Plate 1; modified for clarity). Qal–Alluvium; Qc–Colluvium; Qt–Travertine; Qls–Landslide deposits; Mzs–Sedimentary rocks; MHS–New unnamed spring.

Reservoir temperature estimates of 100°C (212°F) based on chemical geothermometers are consistent with such a flow system. Minimum depths of fluid circulation required to reach these temperatures range from 2-3 km (approximately 1-2 mi). The Paleozoic carbonate rocks occur at such depths under the Sepulcher structural low west of Mammoth and beneath the Yellowstone River valley between Bear Creek and La Duke (Fig. 3.3.12).

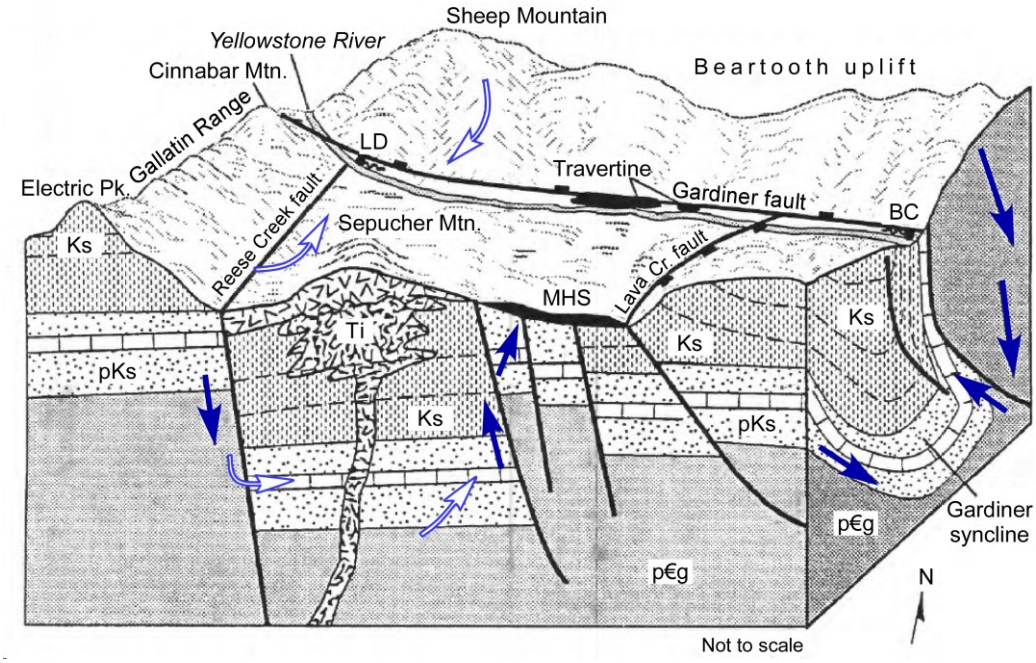


Figure 3.3.12 Conceptual model of possible flow systems associated with thermal areas at Mammoth Hot Springs (MHS), Hot River (HR), Bear Creek Springs (BC), and La Duke Hot Spring (LD). Arrows indicate general directions of fluid flow (solid in exposed planes of the diagram and open along zones unseen in exposed planes of the diagram). Faults shown as heavy solid lines, with block teeth on overriding plate of reverse faults. Simplified stratigraphic units are labeled Ks (Cretaceous sedimentary rocks), pKs (pre-Cretaceous sedimentary rocks), Ti (Tertiary intrusives), and p€g (Precambrian gneiss and schist). Approximate position of the Mississippian Mission Canyon Limestone (upper formation component of the Madison Limestone), a regionally continuous aquifer, is shown by the blocky pattern. From Sorey et al., 1991; modified for clarity.

Thermopolis Hot Springs, Hot Springs County

Hot springs in Hot Springs State Park in Thermopolis (Greek for "hot city") is the largest group of thermal springs in the country outside Yellowstone. Thermopolis is 214 miles west of Jackson, Wyoming via US-26E, WY-133N to US-26E, and US-20N.

The Hot Springs State Park (https://hotspringsstatepark.com/) is renowned for its spectacular *Rainbow Terraces* of travertine (Figure 3.3.13) where more than 1.8 million gallons flow over the terrace every 24 hours at a temperature of 128 degrees Fahrenheit. A winding, concrete walkway atop the terraces provides a view of algae and mineral formations on the bluffs overlooking the Big Horn River.

The park's free State Bath House has both indoor and outdoor pools open to the public year-round where the water is maintained at 104°F for therapeutic bathing. In addition to the State Bath House, Hot Springs State Park also has two commercial hot springs pool areas: Hellie's Tepee (https://tepeepools.com/) and Star Plunge (https://starplunge.com/). Both facilities have indoor and outdoor pools and hot tubs, baby pools, waterslides, and steam rooms. Each establishment owns its own hot mineral water swimming pools which are open to the public.



Figure 3.3.13 Rainbow Terraces in Hot Springs State Park Photo courtesy of Hot Springs Travel & Tourism; available at https://thermopolis.com/attractions/hot-springs-state-park/

As explained on the Park's website, the terraces were known and used for many years by Native Americans, who believed that the waters were beneficial to health and that they could make a warrior invincible in battle. Chief Washakie of the Shoshone tribe, who built a personal bath house there, and Chief Sharp Nose of the Arapaho tribe, sold the hot springs to the United States in 1896 with the provision that a portion should be forever reserved for the use and benefit of the public. Today nothing is left of Washakie's bath house, although a small marker may be found at the site. At Thermopolis each year in early August the presentation of the springs to the white man is re-enacted in the "Gift of the Waters Pageant." When the United States released a one-square mile tract of land to the state in 1897 to establish Big Horn Hot Springs State Reserve, the Reserve became the first of Wyoming state parks.

Breckenridge and Hinckley (1978) provide a description of the hydrogeologic setting of the Thermopolis spring system and the characteristics of six main springs including the largest and most impressive, Big Spring (Fig. 3.3.14-*Left*), which at the time had flow rate between 2,280 and 3,173 gpm over a period of 8 years, with an average of 2,648 gpm. The water temperature remained constant throughout the 20th century, in the 133-135°F range.

"Big Spring forms a 25-foot diameter pool through vents in the pool bottom. Inflow is accompanied by constant gas bubbling which produces a churning, boiling effect. Two manmade ditches and a buried pipeline accommodate outflow. One ditch flows into a swimming pool 170 yards south while the other, larger channel feeds water across 200 yards of extensive and spectacular terrace formations (Rainbow Terraces), then into the State Bathhouse and a second swimming pool. In addition, the underground pipe system serves the pools and bathhouse. Outflow from all these springs finally enters the nearby Bighorn River. Spring waters piped to commercial establishments at the south end of the park have a vapor vent in the form of a standpipe. Around this pipe mineral deposition has created a 20 feet high cone known as Teepee Fountain (Fig. 3.3.14-*Right*)" (Breckenridge and Hinckley, 1978).



Figure 3.3.14 *Left*: Big Spring in Hot Springs State Park, Thermopolis. Photo by Julie Henning; available at https://www.roadtripsforfamilies.com/hot-springs-state-park-thermopolis-wyoming/. *Right*: Teepee Fountain, a cone of travertine formed by mineral deposition around a standpipe. Photo courtesy of Hot Springs State Park.

Breckenridge and Hinckley (1978) describe the hydrogeology of the Thermopolis springs system as a nearly ideal natural artesian system where recharge occurs at exposed portions of Paleozoic formations in the Owl Creek uplift, and discharge at the crest of the Thermopolis Anticline cut by Bighorn River. The distribution of travertine deposits in the area indicates that the springs have not always issued from their present locations and have slowly migrated from approximately three miles northwest of Thermopolis (still at the crest of the anticline) towards their present site.

All the springs now flow from outcrops of the red Chugwater Formation at the crest of the anticline. However, based on the formation's low permeability and water chemistry (the Chugwater characteristically has Total Dissolved Solids >30,000 ppm), the authors conclude that the source of water is deeper: "Because strata are bent so sharply at the anticlinal crest, many have fractured and shifted, providing a ready water pathway from much lower formations."

Breckenridge and Hinckley find that the most reasonable hypothesis is that the water comes in large part from the deeply seated Madison Limestone, probably modified in chemistry as it cools and rises through rock units above it. They emphasize the main problem in explaining the Thermopolis spring system: the source of heat. A variety of possibilities have been proposed, including chemical or frictional heating which may contribute some energy, but

to what extent this might supplement thermal gradient heating is unknown. If the waters are assumed to circulate into the Madison and deeper, a locally high geothermal gradient may account for the observed temperatures. Until all these factors are more clearly defined, no explanation can be completely accepted (Breckenridge and Hinckley, 1978).



Figure 3.3.15 Travertine deposits above White Sulphur Springs, Hot Springs State Park, Thermopolis. © 2024 Hot Springer Ken. https://hotspringers.net/white-sulphur-spring/

Periodic Spring, Lincoln County

Periodic Spring discharges from a solution cave in the Madison Limestone along the axis of the Periodic anticline in the Absaroka thrust sheet of the Salt River Range of western Wyoming (Figs. 3.3.16 and 3.3.17).

The spring turns on and off with variable cyclicity during low discharge stages. The intermittent yield is caused by the mechanical operation of a siphon in the cave behind the spring (see Fig. 1.57 in Chapter 1.2). The cold, good quality water discharged from the spring serves as the Town of Afton municipal water supply. The major component of recharge to the spring originates on elevated outcrops of the Madison Limestone four miles east of the spring on the eastern side of the topographic divide of the Salt River Range. Smaller amounts of water are derived from infiltration through fractures along the axis of the Periodic anticline between Swift and Cry Creek canyons. The Spring varies from low flows of 5 cubic feet per second (cfs) to high flows of up to 100 cfs (Huntoon and Coogan, 1987).

Periodic Spring Trail can be reached from Highway 89 in Afton, heading west on County Road 138 that continues as Forest Service Road 10211 at the Bridger-Teton National Forest boundary. This gravel road winds through the Swift Creek canyon for about four miles where it ends. The trail follows the creek side for a little over 1/2 mile to the Periodic Spring.

The actual spring orifice has been modified to prevent rocks falling into the vent. It has a concrete roof, metal baffles and assorted pipework. A rockslide that occurred in late May 2024 has partially covered the Periodic Spring and the spring trail.

The water typically flows from the spring for 15 to 20 minutes, decreases in intensity during the next few minutes and stops completely for several minutes more. The flow resumes quickly, preceded by muffled subterranean noises, though without the sudden surge that characterizes a geyser. The cycle is best observed during mid to late summer, as at other times the periodicity is less evident.

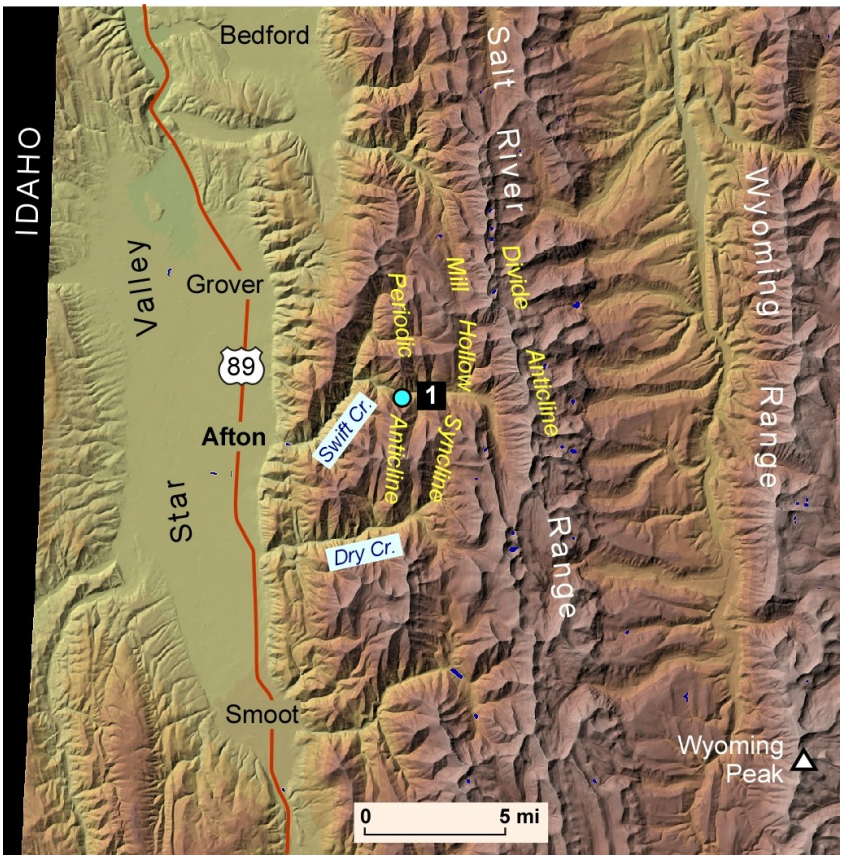


Figure 3.3.16 Periodic Spring (#1) can be reached from Highway 89 in Afton, Wyoming heading west on County Road 138 which continues as Forest Service Road 10211 at the Bridger-Teton National Forest boundary. This gravel road winds through the Swift Creek canyon for about four miles where it ends. The Periodic Spring Trail follows the creek side for a little over 1/2 mile to the Periodic Spring. Shaded relief map courtesy of USGS.

Several descriptions of Periodic Spring including mechanism(s) that are thought to cause its intermittent ("ebb-and-flow") nature, as well as photographs and videos taken before the recent rockslide can be found at https://www.worldrecordacademy.org/2023/11/worlds-largest-rhythmic-spring-world-record-near-afton-wyoming-423543

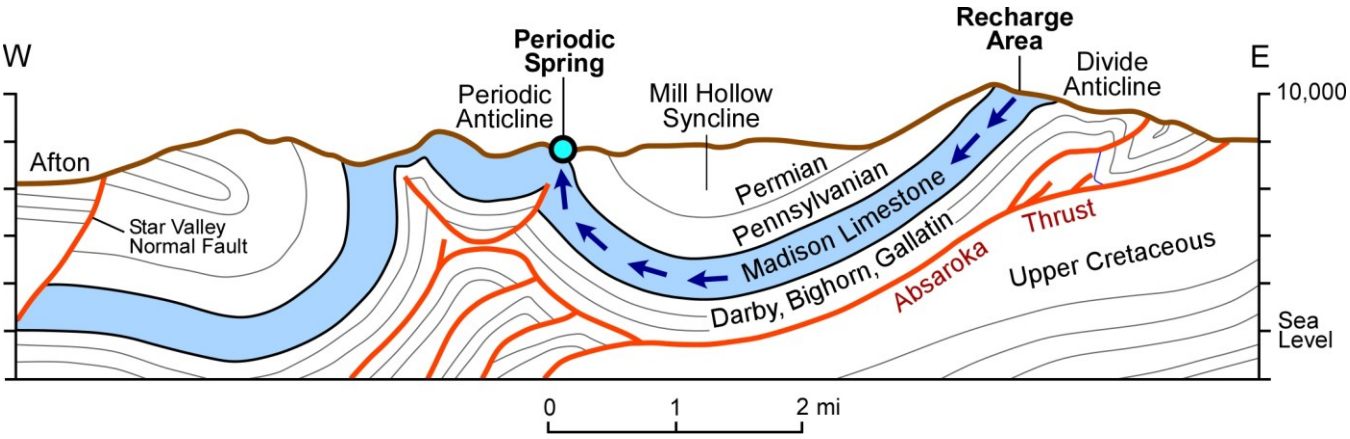


Figure 3.3.17 Portion of an East-West cross section through the overthrust belt developed by Huntoon and Coogan (1987). Arrows denote the probable flow of groundwater in Madison aquifer (colored blue) from the recharge area on the eastern topographic high of the Salt River Range (Divide Anticline) through Mill Hollow Syncline to Periodic Spring. Faults and Absaroka Thrust are shown in red. Modified from Blanchard et al., 1990.



Figure 3.3.18 *Left*: Photograph of Swift Creek as it tumbles out of the Periodic Spring, near Afton, Wyoming, as seen from the bottom (taken in September 2009). Author: Ninjatacoshell. The file is licensed under the Creative Commons Attribution-Share Alike 3.0 Unported license. Available at https://en.wikipedia.org/wiki/File:Periodic_Spring_(from_the_bottom).JPG. *Right*: A rockslide that occurred in late May 2024 has partially covered the Periodic Spring Photo by Eric Buehler taken on June 1, 2024. More photographs of the slide can be found at the Afton's Star Valley Independent News website: https://svinews.com/blog/2024/06/02/photos-intermittent-spring-rock-slide/



Figure 3.3.19 Periodic Spring prior to water supply development. Photo courtesy of Star Valley Historical Society https://starvalleyhistoricalsociety.org/artifacts/item/19994

Big Spring, Lincoln County

The Big Spring Scenic Backway, located entirely within Lincoln County in southwestern Wyoming, is named after Big Spring, one of the most scenic karst springs in the West. Historically, this Backway was part of the Oregon Trail. It spans 68 miles connecting the towns of Kemmerer and Cokeville. The route follows Wyoming State Highway 233 (Hams Fork Road) out of Kemmerer, transitions to Forest Road 10062 across the mountains, and links to Wyoming State Highway 232 (County Road 204) into Cokeville. Big Spring is located about 21 miles from Cokeville and 44 miles from Kemmerer on USFS Road 10062. The spring is located across West Hams Fork Creek within the Bridger-Teton National Forest. The parking lot and access road are east of the West Hams Fork Creek while the spring and picnic area are west of the creek. A trail and bridge connect these two areas over the creek (Fig. 3.3.20).

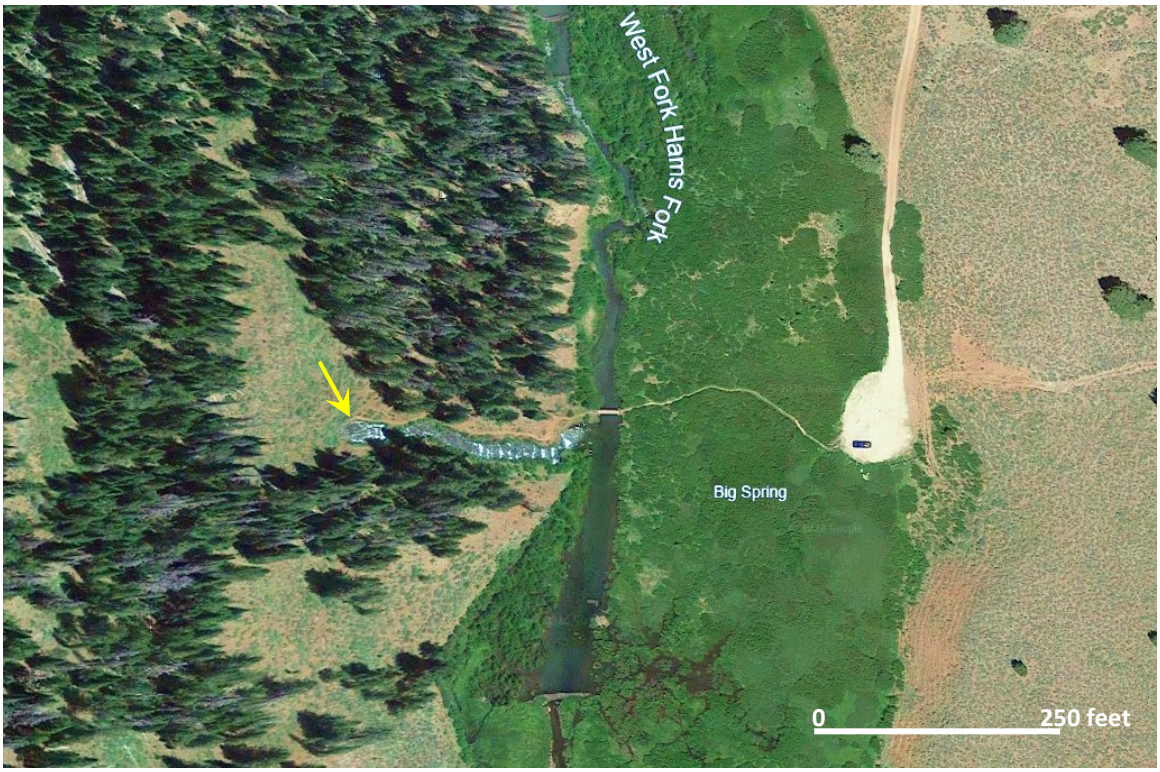


Figure 3.3.20 Google Earth image (acquired on July 27, 2026) of the Big Spring stop on The Big Spring Scenic Backway in Lincoln County. The spring, pointed out by the arrow, was referred to as Ladder Springs in older reports. The name comes from a ladder-like appearance of the cascading water on its way to the West Fork Hams Fork. The cascades are formed by differential erosion, by the flowing spring water, of thin Lower Triassic Woodside Red Beds consisting of siltstone, shale and argillaceous sandy limestone.

The Backway cuts through the southern portion of the Bridger-Teton National Forest, plunges deep into the Tunp Mountain Range, and passes right near the Fossil Butte National Monument just west of Kemmerer. Large stretches of this route consist of unimproved dirt and gravel Forest Service roads. High-clearance vehicles and good tires are strongly recommended for the mountain pass sections (https://fossilbasin.org/2023/05/23/a-guide-to-wyomings-big-spring-scenic-backway/).

The bedrock where Big Spring emerges in the Tunp Mountain Range is primarily composed of Triassic marine sedimentary rocks. Mapped on geologic maps as Ladder Spring (e.g., Rubey et al., 1980), Big Spring issues from

the Lower Triassic Thaynes Limestone at the contact with the underlying Lower Triassic Woodside Red Beds consisting of siltstone, shale and argillaceous sandy limestone.

Big Spring is a typical overflowing contact spring, with the less permeable base of the karst aquifer sloping away from the contact, with an angle of dip between 20 and 45 degrees to the west (Fig. 3.2.22).



Figure 3.3.21 Big Spring on Big Spring Scenic Backway in Lincoln County. *Left* photo courtesy of Fossil Basin website (https://fossilbasin.org/2023/05/23/a-guide-to-wyomings-big-spring-scenic-backway/). *Right* photo courtesy of Lincoln County.

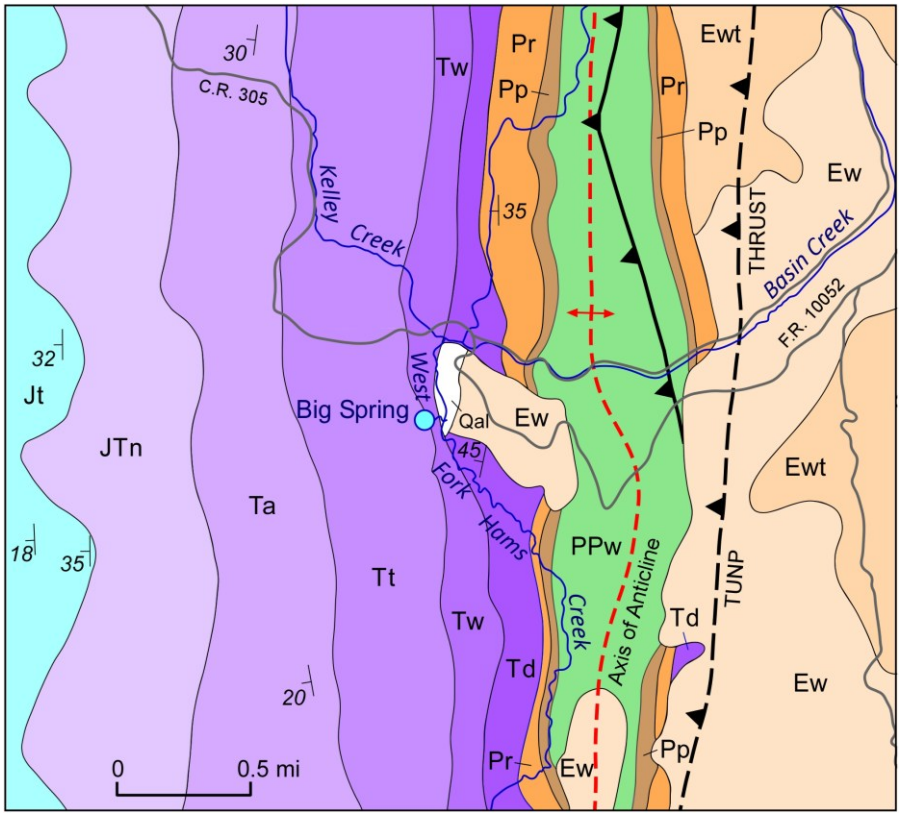


Figure 3.3.22 Geologic map of the Big Spring area. Modified from Rubey et al., 1980. Geologic units, from oldest to youngest: PPw—Permian and Pennsylvanian Wells Formation: quartzite and sandstone; limestone and siltstone. Pp—Permian Meade Peak Phosphatic Shale Member of Phosphoria Fm. Pr—Permian Rex Chert Member of Phosphoria Fm.: chert and limestone. Td—Lower Triassic Dinwoody Fm.: siltstone, shale, and limestone. Tw—Lower Triassic Woodside Red Beds: siltstone, shale, and sandy limestone. Tt—Lower Triassic Thaynes Limestone. Ta—Upper Triassic Ankareh Beds: mudstone, quartzite, sandstone, and siltstone. JTn—Nugget Sandstone (Jurassic? and Triassic?). Jt—Middle Jurassic Tween Creek Limestone. Ew—Eocene Wasatch Fm.: mudstone, sandstone, and conglomerate. Ewt—Eocene Wasatch Fm.: unsorted blocks of older rocks in mudstone matrix. Qal—Alluvium.

3.3.3 South Dakota

The largest karst springs in South Dakota are in The Black Hills, an extensive, elliptically domed area in northwestern South Dakota and northeastern Wyoming, about 125 miles long and 65 miles wide. Rapid City, South Dakota, is on the Missouri Plateau at the east edge of the Black Hills. Uplift caused erosion to remove the overlying cover of marine sedimentary rocks and expose the granite and metamorphic rocks that form the core of the dome. The peaks of the central part of the Black Hills are 3,000 to 4,000 feet above the surrounding plains. Harney Peak, with an altitude of 7,242 feet, is the highest point in South Dakota. These central spires and peaks all are carved from granite and other igneous and metamorphic rocks that form the core of the uplift.

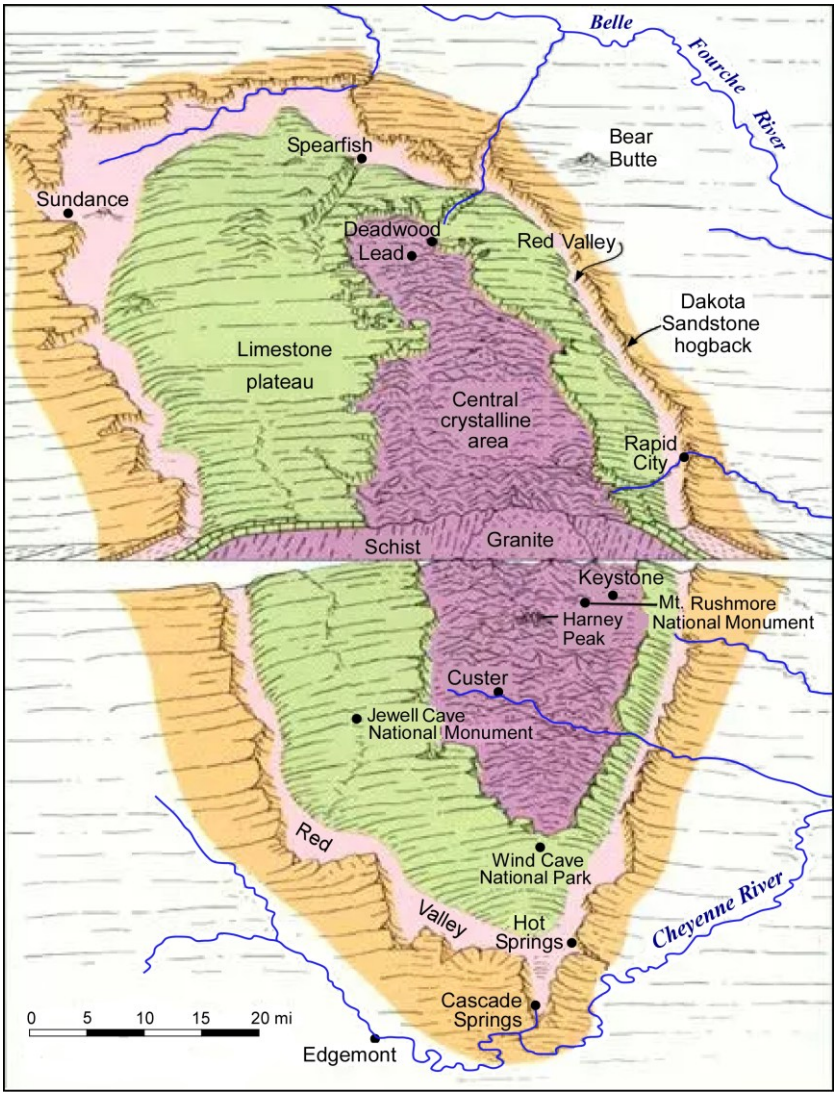
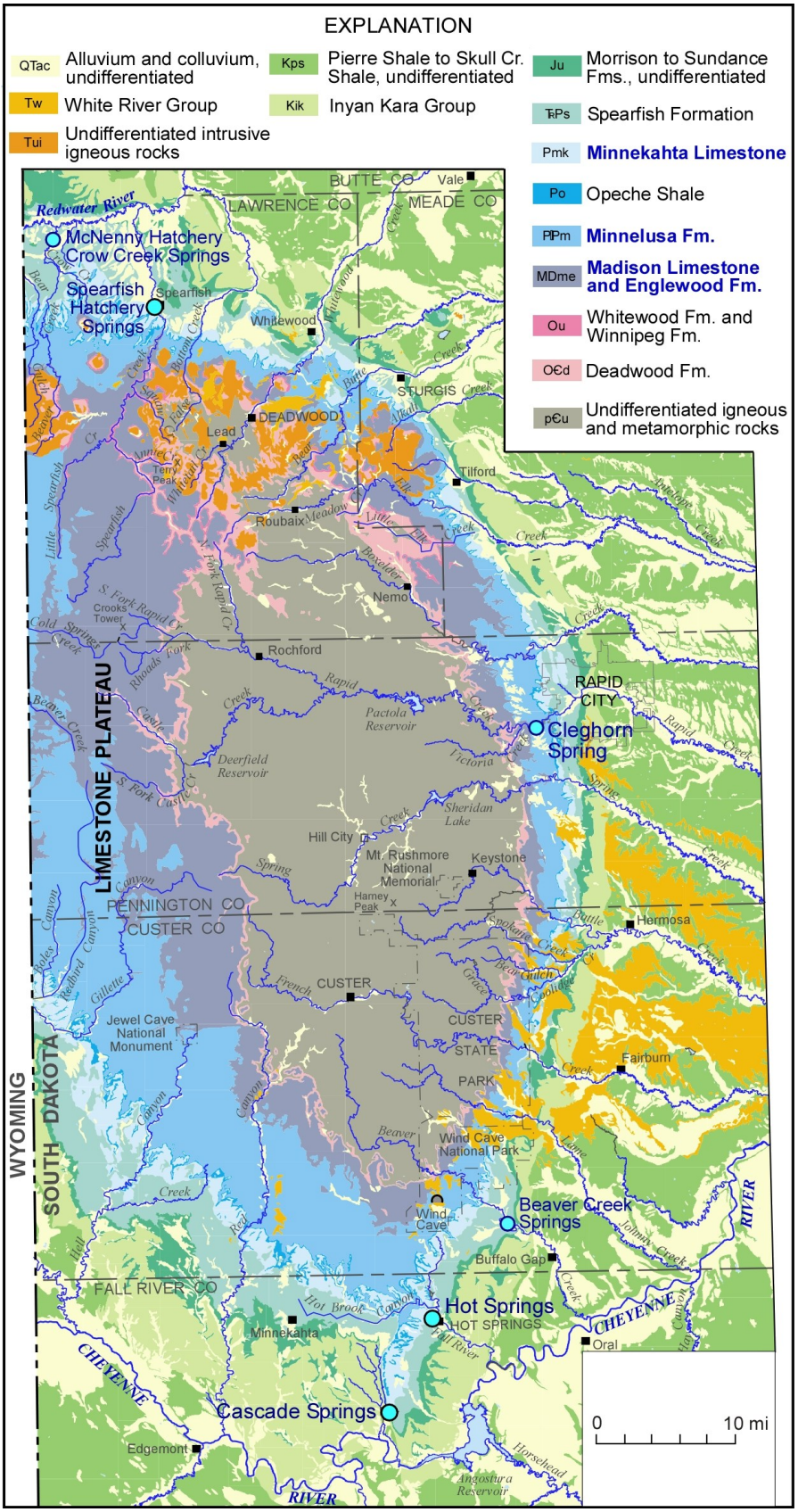


Figure 3.3.23 Schematic of the Black Hills uplift. Courtesy of NPS; available at . https://www.nps.gov/moru/learn/nature/geologicactivity.htm

Black Hills is a hydrogeologic setting characterized by streamflow losses that occur as streams cross outcrops of the karstified Madison (Pahasapa) Limestone of Mississippian age, and Minnelusa Formation of Pennsylvanian to Permian age (Driscoll et al., 2002). Within this setting, large artesian springs, originating primarily from the karstic Madison and Minnelusa aquifers, occur in many locations downgradient from the loss zone, most commonly within or near the outcrop of the confining Spearfish Formation of late Permian to Triassic age. These springs provide an important source of baseflow in many streams beyond the periphery of the Black Hills (Rahn and Gries, 1973; Miller and Driscoll, 1998).

As discussed by Rahn and Gries (1973), large amounts of water are recharged to the inner edge of the karstified carbonate aquifer encircling the impermeable Precambrian core of the Black Hills. When the streams originating at the impermeable core flow radially away from it across the karstified surface, they lose significant quantities of water to the underlying carbonate aquifers. Even larger quantities of water discharge from the outer edge of the



carbonate aquifers via numerous springs at topographic lows in the "groundwater dam" caused by the thick sequence of shales of Triassic and Jurassic age. As postulated by Rahn and Gries, the difference between the total streamflow loss (44 cfs) and the average spring discharge (190 cfs) must be due to precipitation falling directly on the carbonate aquifers.

Leakage between the Madison and Minnelusa aquifers is common, but highly variable (Peter, 1985; Greene, 1993). The general direction of groundwater flow in the southern Black Hills in both the Minnelusa and Madison aquifers is from north to south, but also may be from the northwest to southeast, or west to east (Whalen, 1994).

Contributions from the Madison and Minnelusa aquifers to individual artesian springs cannot necessarily be quantified precisely (Naus et al., 2001) because of geochemical similarities between the aquifers.

Figure 3.3.24 Geologic map of the Black Hills in South Dakota, including locations of the largest springs (large blue dots). Karst aquifers are shown with blue names in the Explanation. Modified from Driscoll et al., 2002.

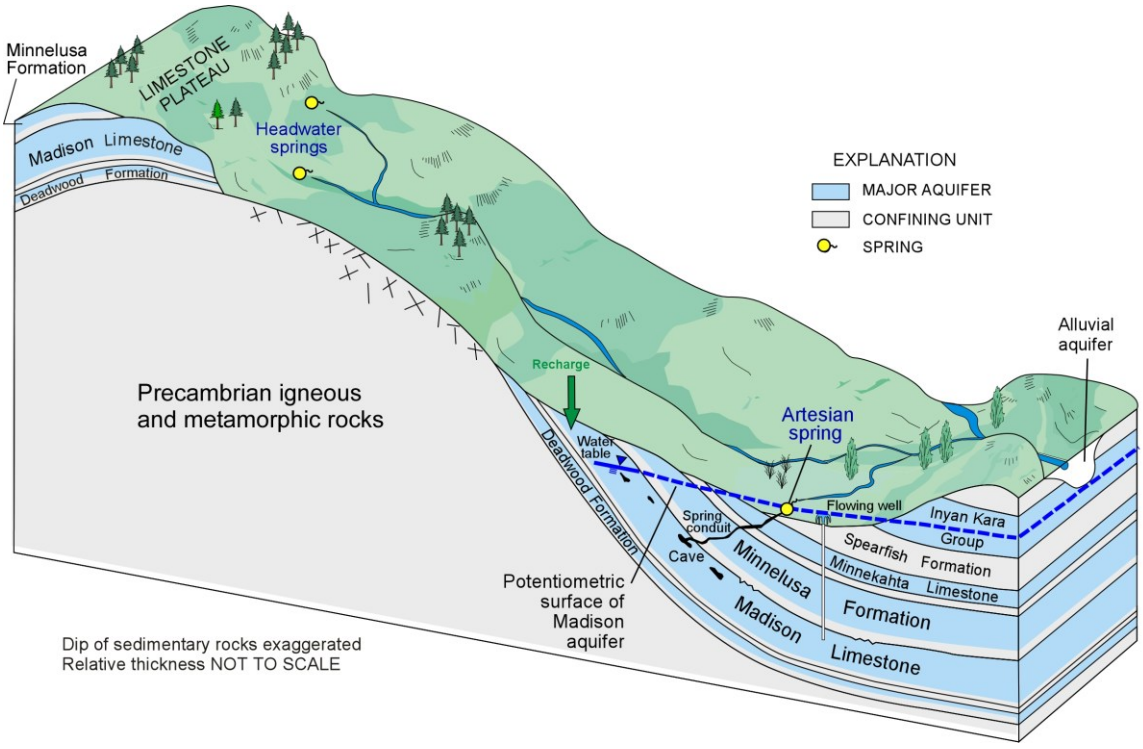


Figure 3.3.25 Schematic showing the hydrogeologic setting of Black Hills area. Modified from Driscoll et al., 2002.

For some springs, high sulfate concentrations could indicate Minnelusa influence but may also result from dissolution of Minnelusa minerals by water from the Madison aquifer. Previous investigators (Whalen, 1994; Klemp, 1995; and Hayes, 1999) used geochemical modeling to estimate contributions of the Madison and Minnelusa aquifers to selected springs. The Madison aquifer generally was identified as the primary source, with variable contributions from the Minnelusa aquifer, or chemical influences resulting from residence time within the Minnelusa Formation (Driscoll et al., 2002).



Figure 3.3.26 Boxelder Creek disappearing into the alluvium overlying sinkholes in the Madison (Pahasapa) Limestone. Note whirlpool. Originally published in Rahn and Gries, 1975, courtesy of South Dakota Geological Survey. The average discharge of Boxelder Creek where it exits from the Precambrian rocks is 13.23 cfs. It loses almost all its flow to several large caves and deposits of alluvium overlying the Pahasapa (Madison) Limestone. According to Gries et al. (1968), it is possible to crawl into these caves to a point under the stream bed where water still flows above. Farther downstream there are several large ephemeral springs; they flow when the water table is high. Boxelder Creek is the largest disappearing stream in Black Hills (Rahn and Gries, 1975.)

Cascade Springs, Fall River County

As described by Rahn and Gries (1973), "Cascade Spring is the largest single spring in the Black Hills of South Dakota. It issues forth dramatically at the contact of the Minnekahta and Spearfish Formations. It has a steady discharge and constant temperature of 67°F. One mile north of Cascade Spring itself is a smaller spring called "Cold Spring," which has a discharge of about 1.1 cfs and a temperature of 64°F. The discharge of Cascade Spring and Cold Spring together averages 23.65 cfs. Because of the large discharge rate of the spring, and because of the small surface drainage basin above the spring, it is obvious that Cascade Spring is recharged by water entering the limestone underlying other surface drainage basins; namely, the expanse of limestone to the northwest towards Jewel Cave."

There are six known discrete discharge points at Cascade Springs which occur within Quaternary-age alluvial deposits in a narrow valley along the contact between the Spearfish Formation and the underlying Minnekahta Limestone, near the crest of the strongly asymmetric Cascade Anticline (Post, 1967). These six individual springs form headwaters of Cascade Creek. The three most upstream spring vents are within a flat-bottomed upper spring basin. Farther downstream are the two "gazebo" springs, located within a middle area of hummocky relief where depressions may indicate other abandoned discharge points. The most downstream spring is located about 90 feet upstream from the USGS streamflow-gaging station (Hayes, 1999).

A description provided by Darton of USGS in 1901, reads "*Cascade Creek.*-This vigorous stream is the product of the springs at Cascade, where a large volume of water issues from the Minnekahta limestone. In Pl. CIII (Figure 3.3.27 in this book) are shown the two largest springs at this place. The water is slightly warm, and contains a large amount of mineral matter. The volume of the creek at its mouth was 25 second-feet on May 18, 1900."

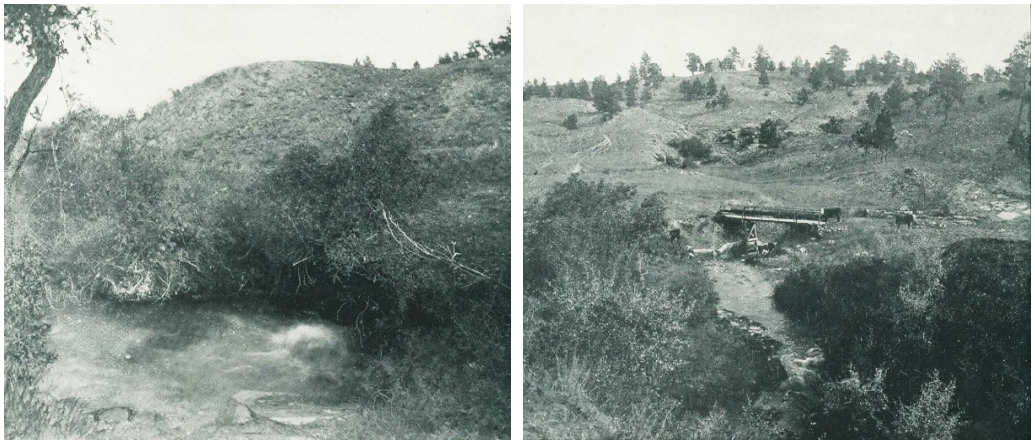


Figure 3.3.27 Two largest springs at Cascade Springs emerging from crevices in Minnekahta Limestone. Plate CIII from Darton, 1901.

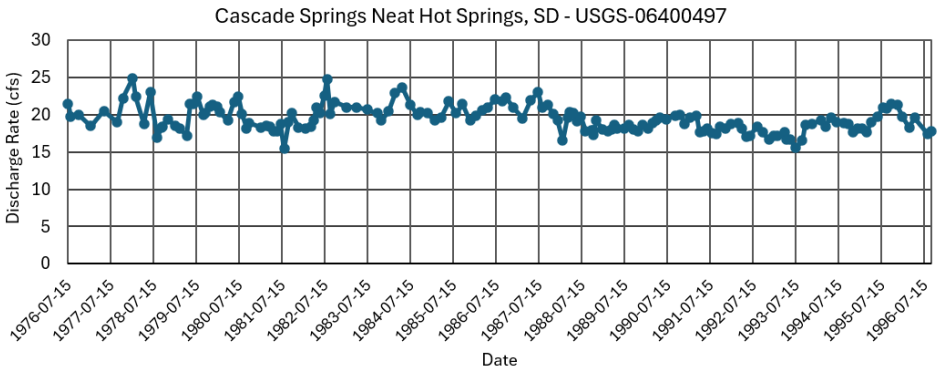


Figure 3.3.28 USGS maintained a gaging at Cascade Springs between 1976 and 1996. The average discharge rate for the period of record is 19.4 cfs.

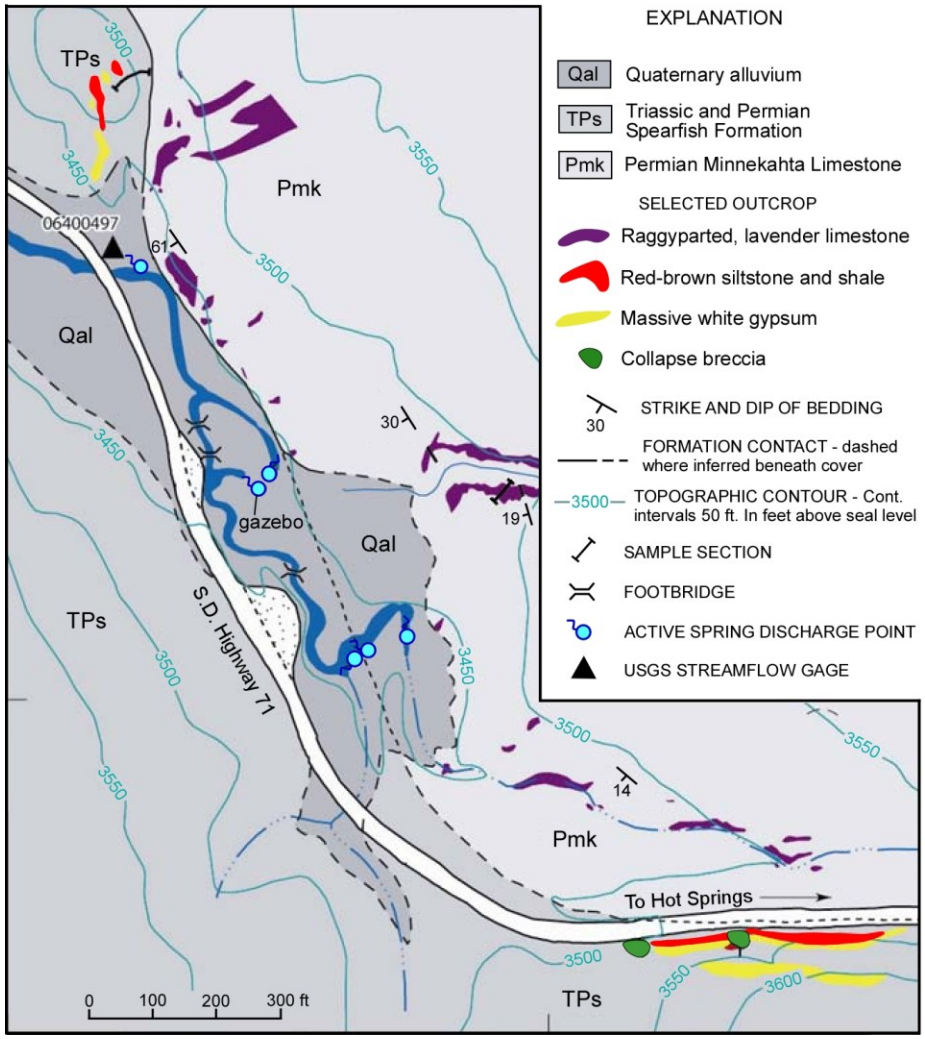


Figure 3.3.29 Geology in the vicinity of Cascade Springs. Modified from Hayes, 1999.

Groundwater discharging from the Madison aquifer at artesian springs was referred to as "rejected recharge" by Huntoon (1985). Huntoon hypothesized that recharge is rejected from the Madison aquifer as transmissivity decreases with distance from upgradient recharge areas. Interactions with the overlying Minnelusa Formation may be an important factor in this process.

Upwelling water from the Madison aquifer (rejected recharge) at locations like Cascade Springs is shown to be a likely process by which breccia pipes are formed through anhydrite dissolution in the Minnelusa Formation and other overlying bedrock units. Under this hypothesis, vertical hydraulic conductivity would be greatly enhanced within breccia pipes. Furthermore, horizontal hydraulic conductivity also may be enhanced in the Minnelusa Formation by dissolution processes related to upward leakage from the Madison aquifer. Potential processes could include dissolution resulting from leakage in the vicinity of breccia pipes that are abandoned spring throats, active spring discharge, and development of subsurface breccias with no visible surface expression or spring discharge, as well as general areal leakage from the Madison aquifer into the Minnelusa Formation.

Bowles and Braddock (1963) showed that the anhydrite dissolution was accompanied by conversion of dolostone beds to medium-crystalline limestone beds (dedolomitization).

Rahn and Gries (1973) described the contact between the Spearfish Formation and Minnekahta Limestone as a typical location for large springs that discharge from aquifers in the Madison Limestone and Minnelusa Formation. The chemical evolution of water within the Madison aquifer in the Black Hills area has been studied at the regional scale by Back et al. (1983), who noted that the water of Cascade Springs is likely a product of aquifer mixing, based on sulfur isotopes. Whalen (1994) concluded that the flow of Cascade Springs consists of a mix of approximately equal proportions of water from the Madison and Minnelusa aquifers.

Hayes (1999) hypothesized that upward leakage from the Madison aquifer was contributing to ongoing development of breccia pipes at Cascade Springs. He noted that breccia pipes commonly occur in the upper Minnelusa Formation, but very few have been observed in the lower part of the formation. Development of breccia pipes probably contributes to enhanced vertical hydraulic conductivity in the Minnelusa aquifer. Breccia pipes are a likely pathway for upward movement of large quantities of water through the Minnelusa aquifer at artesian spring locations.



Figure 3.3.30 Left: Collection basin for one of the Cascade Springs. Courtesy of Colorado Springs Mineralogical Society Blog, https://csmsgeologypost.blogspot.com/search?q=cascade+springs. *Right*: The larger of two "Gazebo Springs" of Cascade Springs. Photograph by J.S. Clark, courtesy of USGS, from Hayes, 1999.



Figure 3.3.31 *Left*: The Mississippian-age Madison Limestone, which consists primarily of buff-colored, massively bedded limestone, is exposed nearest to Cascade Springs at Onyx Cave in the upper half of Wildcat Canyon. Photo courtesy of Tom & Hope https://treadandsoul.weebly.com/wildcat-canyon.html. *Right*: The falls form where Cascade Creek pours over tufa deposit into a pool below. The pools vary in depth, with shallow wading areas near the banks and a deeper central pool for swimming. The site, about 2 miles downstream of the springs, includes covered picnic tables, restrooms, and a short path from the parking lot to the creek. The water of Cascade Springs is supersaturated with respect to calcium carbonate which is easily precipitated out when the water is aerated, such as at the falls. Photo courtesy of Hot Springs Chamber of Commerce.

Hot Springs, Fall River County

The occurrence of thermal springs in the town of Hot Springs and in the Fall River valley remains a puzzle that has been intriguing geologists and hydrogeologists for almost a hundred years. While the large discharge rate of the springs, the chemical composition of thermal water, and the general geology of Black Hills (including locally visible outcrops) indicate that the primary source of water discharging at the springs is most likely some combination of Madison, Minnelusa and Minnekahta carbonate aquifers, the source of the heat remains a mystery.

The entire region is underlain by pre-Cambrian metamorphics. The Deadwood (Cambrian) sandstone, and the Mississipian Englewood and Pahasapa (Madison) Limestone underlie, but do not outcrop, this area. The oldest rock which is exposed at the surface, and with which the thermal springs are associated, is the Pennsylvanian Minnelusa limestone and sandstone which is overlain in order by the Lower Permian Opeche Shale, the Middle to Late Permian Minnekahta Limestone, and the Permian to Triassic Spearfish red shale. Pleistocene conglomerates occur in the valley of Fall River.

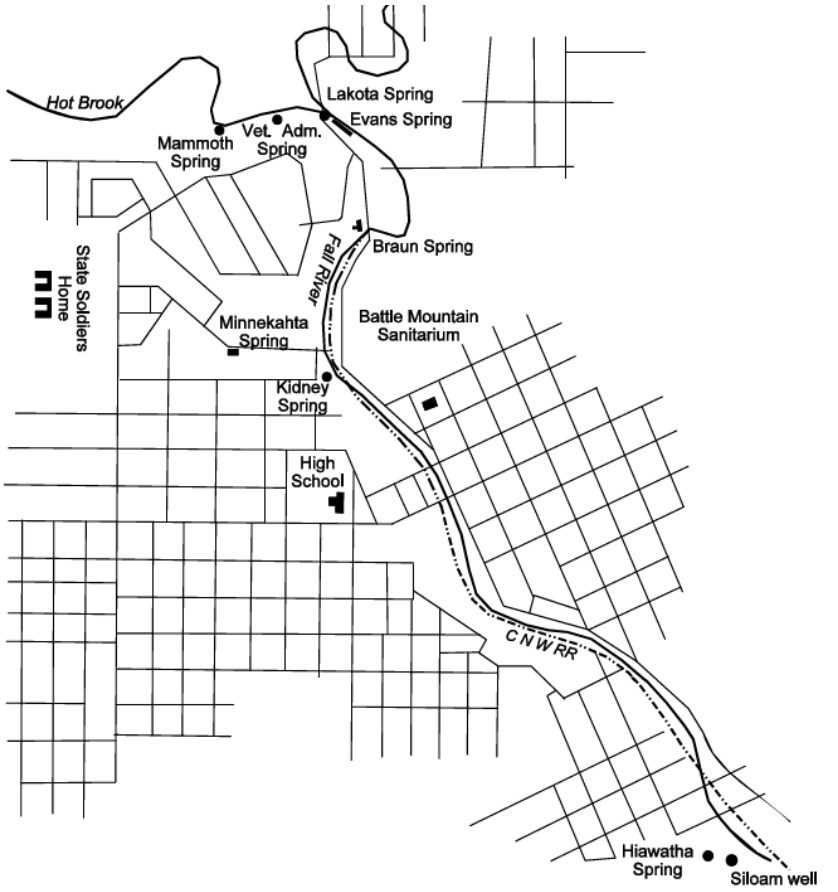
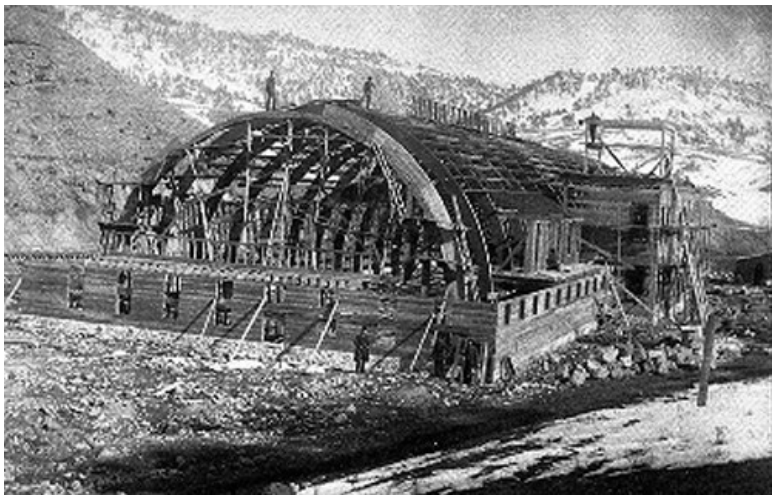


Figure 3.3.32 Map of Hot Springs showing the various hot springs. From Lund, 1997.

Although lying on the southeastern edge of the general Black Hills uplift, the dominant structural feature of the area is an asymmetric anticline just west of the city of Hot Springs. This anticline, which is plunging to the south, is first observed in the northern part of the area and extends in a southerly direction. The west flank of the anticline is more steeply dipping than the east. Having a length of more than 15 miles, the average width of 3 miles, and a closure of more than 800 feet, the Hot Springs anticline, as this structure is designated, is one of the largest secondary structures of the Black Hills anticlinorium (Work, 1934)

The following descriptions of thermal springs in the town of Hot Springs and the vicinity are from Lund (1997) and Work (1934).

The Evans Plunge was built in 1890 over numerous small sparkling springs and one mammoth spring of mineral water with a temperature of 87 degrees Fahrenheit and of medicinal qualities proclaimed on good authority, to be superior to that of the famous Warm Springs, Georgia. From the inflow of 5,000 gallons of water per minute



from the springs arising out of the pebble bottom there is a complete change of water 16 times daily, thus insuring clean, fresh, living water at all times. “The pool 50 x 200 ft., ranges in depth from 4 ft. to 6 ft. with two shallow enclosures for children.” (From a sign opposite the Evans Plunge.)

Figure 3.3.33 Construction of Evan’s Plunge in 1890. From Lund, 1997.

Springer Ranch Springs, from which the city of Hot Springs secures its water supply, is located in Hot Brook Canyon about 2.5 miles west of Hot Springs. There is no single spring, but for a distance of 200 feet the water issues along the bed of the canyon. Below this spring Hot Brook is a permanent stream, above it only intermittent. The springs emerge on the axis of the Hot Springs anticline, 0.5 miles north of the crest, and 100 feet above the base of the Minnelusa formation. They produce over 1,000 gpm of 81°F water. The water is low in dissolved solids (398 ppm - mg/L) of which the majority is calcium carbonate (based on analysis by the State Chemical Laboratory in 1938) and thus was felt to have limited medicinal uses (Natural Resources Commission, undated).

Mammoth Springs is located at the junction of Hot Brook and Cold Brook, and like the Springers Ranch Spring, the water issues through several orifices along the stream bed, from the base of a steeply dipping limestone cliff. Water temperature is 92°F and has been used to supplement the flow from Evan’s Springs to Evan’s Plunge. An analysis by Charles B. Gibson and reported by the South Dakota Geological Survey in 1900 gave a total residue (dissolved solids) of 1420 ppm (mg/L) of which the major constituents were calcium sulfate at 618 ppm sodium sulfate at 398 ppm, potassium sulfate at 96 ppm, calcium chloride at 96 ppm, magnesium chloride at 70 ppm, and magnesium carbonate at 60 ppm. According to Work (1934), Mammoth Spring has a larger flow than all the rest of the springs in the vicinity of Hot Springs put together. It issues about two-thirds of the way down the east limb of the Hot Springs anticline, from the top of the Minnekahta limestone formation.

The Minnekahta Springs near the center of town was used by Native Americans who had hollowed out a rock basin into which in the spring water flowed and was used as a bathtub. Most likely, the town received its name from this springs which means “Hot Water” in the Lakota Sioux language. In fact, the city was first called Minnekahta and later changed to the English translation of the Indian word. A sign opposite Evan’s Plunge states: “Long before the white man discovered the valley of healing waters, the Sioux and Cheyenne Indian tribes fought for possession of the natural warm water springs. Legend tells us that the battle was fought on the high peak above the springs and the Sioux emerged victorious. The mammoth spring at the north end of the interior of the plunge is known as the “Old Original Indian Springs. Here the Indians drank and bathed in its warm healing water.”

Minnekahta Springs was used by the Hot Springs Hotel which operated a bath house and plunge. The source appears to be the same as Mammoth Spring with a temperature of 90°F and a total dissolved solids of 1071 ppm (mg/L). These springs are primarily sulphate and sodium chloride waters. The hotel was built in 1889 and the bathhouse was the first in Hot Springs. The complex was torn down in 1963-64.

Braun Springs was used by the Braun Hotel for baths and drinking water. Two springs were used by the hotel, both located on the west side of Fall River. The water temperature was 90°F and had a total dissolved solids of 1496 ppm (mg/L) according to an analysis by Denton Dales (undated). The major constituents were calcium sulphate (428 ppm), calcium carbonate (351 ppm), sodium sulphate (296 ppm), sodium chloride (211 ppm) and magnesium sulphate (185 ppm) with a trace of silica, alumina and iron oxide. Since this water contained Epsom and Glauber salts and some common salt, they were useful for therapeutic purposes. The springs are no longer used by the hotel.



Figure 3.3.34 Evans Plunge pool with thermal water is owned by the City of Hot Springs since 2013. Historic (left) and current (right) photographs courtesy of Evans Plunge Mineral Springs, available at https://www.evansplunge.com/.

According to Rahn and Gries (1973), the total average discharge of all the thermal springs, exclusive of Hot Brook, is 22.92 cfs, and does not vary throughout the year. This value is about 2.2 cfs higher than the U.S. Geological Survey gage located in downtown Hot Springs during the period of October 1968 to January 1969, probably because this site where the discharge rate was measured includes sewage effluent discharge.

Work (1934) offers the following explanation about the origin of thermal Hot Springs: "Assuming that the thermal gradient for this region is one degree of Fahrenheit for every 60 feet of depth it is found that meteoric water would have to descend at least 2700 feet in order to reach a temperature of 90°F. The depth just given does not take into consideration any loss of heat by the water on its return to the surface. Unless Tertiary igneous rocks are present, this area is underlain at a depth of not more than 1100 feet by pre-Cambrian metamorphics. Therefore, water could not follow any artesian structure to a depth sufficient to raise its temperature to that of the water in the springs.

After considering the other possible sources for the heat we are led to the conception that the heat is probably derived from a cooling magma which was intruded in the Tertiary at the time of the general Black Hills uplift and forming the Hot Springs anticline. Fissures were formed along the northern part of the east limb, along the axis, and at a flexure near Cascade Springs of the anticline which extend down to this magma. Due to the gradual cooling and crystallization of the magma highly heated water arid steam is expelled which rises through the fissures, is cooled and condensed as it nears the surface and mingles with some meteoric water, loses any contained volcanic gasses by dissipation through the rock on its upward journey, and issues at the surface as warm springs."

Notably, the Work's hypothesis cannot explain the large discharge rates of the thermal springs which can only be generated by meteoric/surficial aquifer recharge mechanisms, not by "heated water arid steam" of deep magmatic origin, in whatever proportion it may be contributing.

The cause of anomalously warm water in springs and wells of the southern Black Hills remains a mystery. Several plausible alternatives are summarized by Rahn and Gries (1973):

"The unusually warm springs near the town of Hot Springs are too warm to be explained by a normal geothermal gradient. It is unlikely that ground water will circulate below the base of carbonate aquifer prior to flowing upward and discharging at the Minnekahta-Spearfish contact at Hot Springs. Thus groundwater heating by normal geothermal gradient should not be more than: (1,000-foot-thick carbonate section) (1.1°F/100 feet) = 11°F. Since the mean annual air temperature of Hot Springs is 48°F, this could heat the water from about 48°F to 59°F, at most. How then does the water get heated to 87°F?

There are at least four other possibilities to explain the high temperature. The first is that residual or partially cooled magma lies at shallow depth under the southeastern Black Hills, creating a geothermal anomaly. This could be an intrusive igneous body such as the Tertiary intrusives in the northern Black Hills, such as Bear Butte or Devit's Tower. Since the youngest age of the known intrusives is 30 million years, the probability of a new intrusion in the southern Hills seems remote. Additionally, there is no geological or geophysical evidence for such an intrusion.

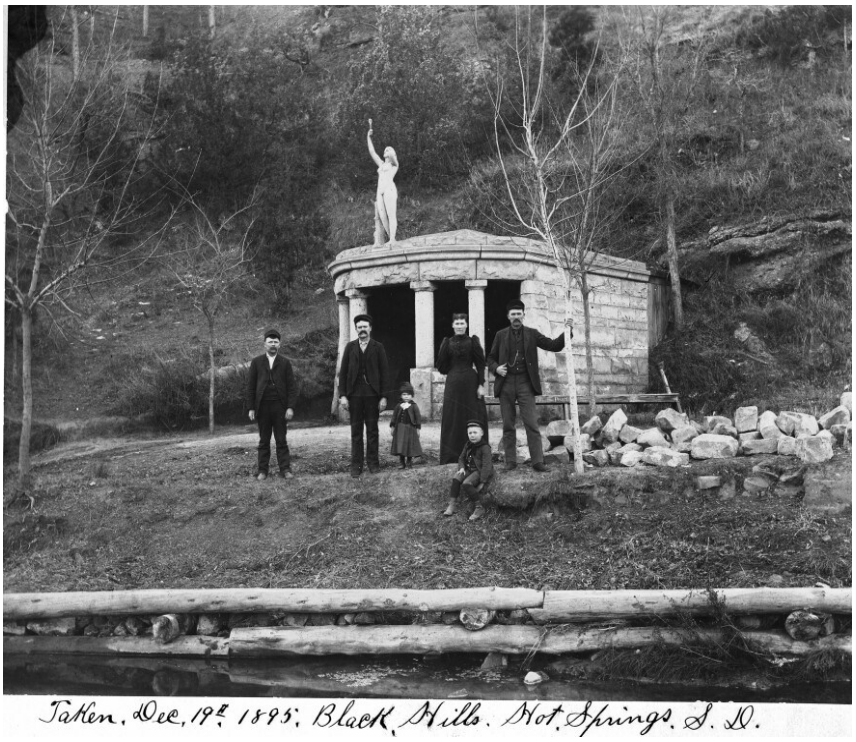


Figure 3.3.35 Historic photograph of Kidney Spring as it looked in 1895. The spring was initially named after Hygeia, the Greek goddess of health, cleanliness, and hygiene. A gazebo built over the spring in 1893 even featured a nude sculpture of the goddess. The statue's nudity scandalized early Victorian-era tourists. The controversial statue was washed away by a flood in 1897. The name eventually shifted to "Kidney Spring" because the waters were bottled, sold, and drank specifically as a remedy for chronic kidney ailments. Photo courtesy of South Dakota Historical Society. Fascinating history of the Hot Springs is provided by Michael Zimny at History of the Hot Springs bath houses and holistic health retreats | SDPB

Another possibility is that ground waters become warmed by the chemical weathering reactions that take place as they flow through the rocks. All oxidation and hydration weathering reactions are exothermic. The chemical weathering of anhydrite to gypsum, at a temperature of 20°C, releases 142 calories/mole. Subsurface drilling in the prairie shows that as much as 300 feet of anhydrite is present in the Minnelusa Formation. None is present in the outcrop except near Sundance, Wyoming. The weathering of this anhydrite and other minerals may account for the thermal springs. Support for this hypothesis is gained by the observation that there is a general correlation between spring temperature, the distance the ground water has traveled prior to discharge, and the total dissolved solids in the spring water. For instance, Cascade Spring is quite warm (67°F). The groundwaters producing this spring have traveled tens of miles, and the spring water has such a high mineral content that it precipitates calc-tufa over its

streambed. Support for this theory is the fact that breccia pipes are found in outcrops of the Minnelusa Formation (Brobst and Epstein, 1963; Post, 1967). These breccia zones apparently formed where anhydrite was dissolved away. Another chemical reaction that could account for high temperatures is the fact that heat is evolved when gas goes into solution with water. Such a mechanism could account for hot water in the Dakota Formation (Schoon, 1971), but is considered unlikely at Hot Springs.

The third possibility may be due to heat generated by radioactive decay. There are naturally radioactive deposits in the Black Hills, especially in the Cretaceous sandstones near Edgemont. Keene (in preparation) noted that the water from the flowing artesian well at Provo is 139°F. It is difficult to explain the origin of all thermal springs in the Black Hills by radioactive decay, however, because the thermal spring waters do not flow through known radioactive rocks, and the thermal spring waters are not particularly radioactive.

A fourth possibility is that the rocks underlying Hot Springs could have a higher conductivity. Thus, perhaps a highly heat-conductive rock within the Precambrian conducts more heat up from the depths of the earth's crust, creating a high geothermal gradient under the town."

Cleghorn and Jackson Springs, Pennington County

Artesian Cleghorn and Jackson Springs are on the grounds of the Cleghorn Springs State Fish Hatchery and originate primarily from the Madison Limestone (Peter, 1985), rise up through the overlying Minnelusa Formation and discharge through the alluvial sediments of Rapid Creek near the contact between the Minnekahta Limestone, which is a minor aquifer in the Black Hills area, and the Spearfish Formation (see Fig. 3.3.25).

The Jackson Springs Gallery is located immediately adjacent to Rapid Creek about 0.5 mi upstream of Canyon Lake. This site was selected for installation of a pumping gallery for the water supply of Rapid City because of the large flows of Jackson Springs. When constructed in 1942, the gallery was positioned over the spring, with one of the gallery laterals extending partly under Rapid Creek. The Jackson Springs Gallery has a sustained production capacity of about 8 cfs, which is equivalent to about 5.2 Mgal/d. It is the largest individual source of municipal water for Rapid City (Anderson et al., 1999).

Cleghorn Springs, which is located immediately adjacent to Jackson Springs, supplies water for the Cleghorn Springs Fish Hatchery and for the Cleghorn Springs Water Users Association. A gravity-flow collection system captures the flow of Cleghorn Springs for delivery to the hatchery raceways. Cleghorn Springs was the site where trout were introduced to the Black Hills in 1886 or earlier by local residents, using stock from Colorado (Henris, 2015). As trout is not native to South Dakota, high fishing pressure, low natural reproduction, and limited food supplies create the need for stocking. The hatchery began operations in 1928 and produces over 100,000 pounds of fish each year.



Figure 3.3.36 Raceways of the Cleghorn Springs State Fish Hatchery. Cleghorn Spring, which supplies the hatchery, is enclosed and the only visible flow from it is part of the combined outfall from the hatchery. Photo courtesy of Cleghorn Springs State Fish Hatchery.

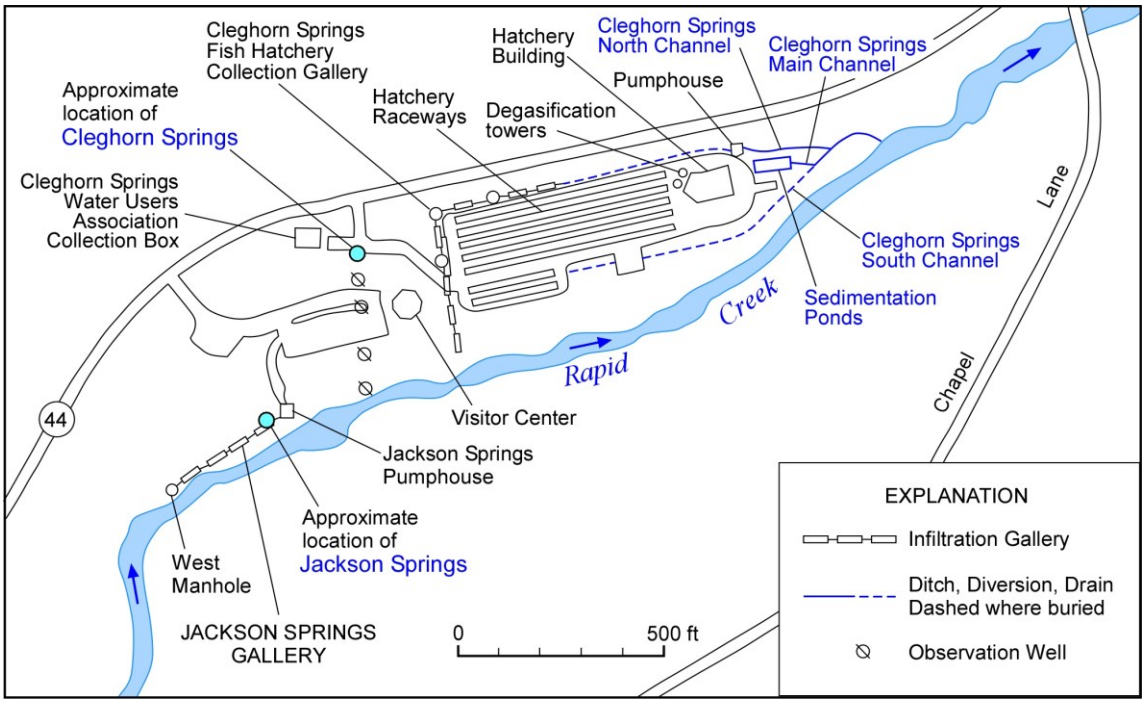


Figure 3.3.37 Map of Cleghorn Springs State Fish Hatchery complex. Modified from Anderson et al., 1999.

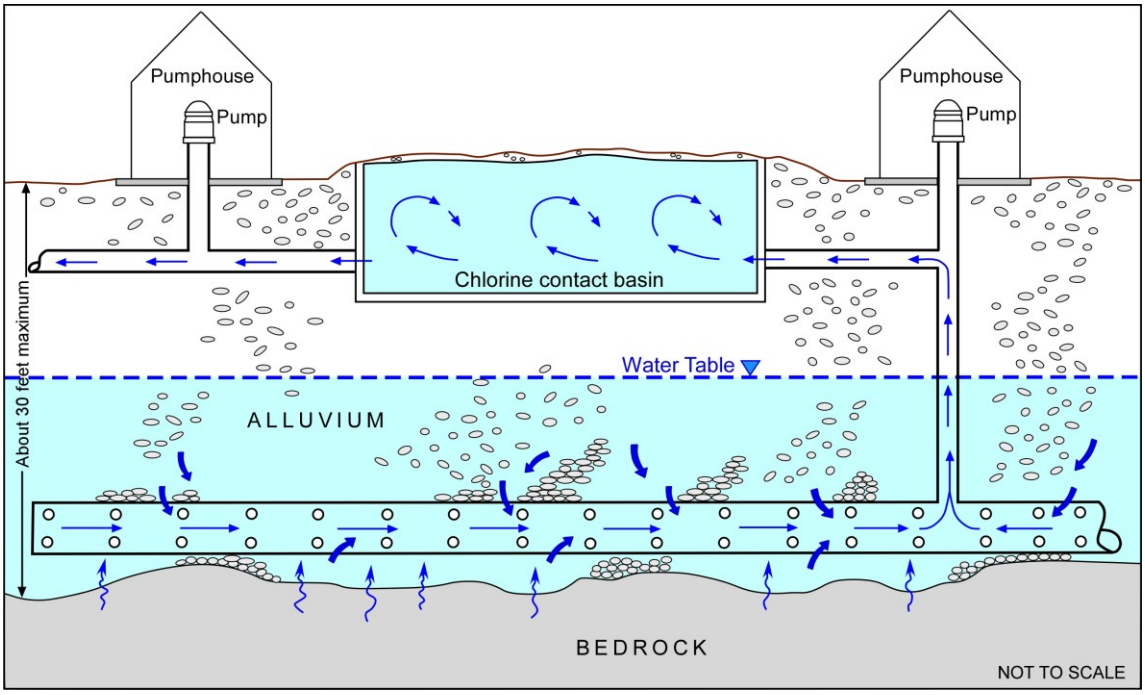


Figure 3.3.38 General schematic of galleries installed along Rapid Creek for the water supply of Rapid City. The Jackson Springs Gallery is built directly over the ascending Jackson Springs, which issue from the karstified bedrock, and has one pumphouse. Modified from Anderson et al., 1999.

Jackson and Cleghorn Springs may be separate springs that issue from a single source. Because of site development, the only visible flows from the collective springs are the outfall from the hatchery and seepage into Rapid Creek adjacent to the springs. The springs are collectively referred to as Jackson and Cleghorn Springs and

the site is referred to as the Jackson-Cleghorn Complex. When referred to individually, Jackson Springs refers to the springflow near the Jackson Springs Gallery and Cleghorn Springs refers to the outfall from the Cleghorn Springs Fish Hatchery.

The Jackson Springs Gallery production during water year 1988-89 averaged 6.0 cfs, and the measured outfall from the adjacent Cleghorn Springs Fish Hatchery was 11.5 cfs. The control-volume analysis indicates that combined bedrock springflow from Jackson and Cleghorn Springs averaged about 21.6 cfs. Gallery pumping caused a drawdown of about 5 ft below the creek level in the west manhole of the gallery. This indicates that at least part of the gallery production results from induced infiltration from the creek. Drawdown was only about 2 ft in observation wells located about 200 ft east of the gallery, and the hydraulic gradient remained toward Rapid Creek. Water levels in the alluvial aquifer recovered completely within 2 days after pumping was discontinued, and flow in Rapid Creek responded almost immediately, with an increase similar to the pumping rate. Most of the increase consisted of additional seepage from the alluvium to the creek; however, the outfall from the hatchery also increased measurably. The flow of the hatchery outfall was determined to be related to flow (or stage) in Rapid Creek as well as to the gallery pumping rate (Anderson et al., 1999).

The hydrochemical characteristics of Jackson and Cleghorn Springs are very similar. Water from both springs is supersaturated with respect to dissolved nitrogen and undersaturated with respect to dissolved oxygen, which is characteristic of bedrock aquifers in the area. It is concluded from interpretation of data for 18O and D that Jackson and Cleghorn Springs represent a regional discharge point from the Madison aquifer. The flow of these springs consists of isotopically heavier water from the Madison aquifer south of Rapid Creek and a slightly smaller component of isotopically lighter water, which consists of a mix (of unknown proportions) of water from the north and water recharged in the loss zone of Rapid Creek in Dark Canyon. It is estimated that about 90 percent of the Jackson Springs Gallery production is springflow from the Madison aquifer and about 10 percent is induced infiltration from Rapid Creek (Anderson et al., 1999).

Spearfish Hatchery Springs, Lawrence County

Spearfish Hatchery Springs were utilized by the Spearfish National Fish Hatchery, established in 1896 as one of the oldest, and still operating, trout hatcheries in the country. In the early 1890s, a bill introduced in the United States Senate by South Dakota Senator R.F. Pettigrew, appropriated $500 for investigating the possibility of placing a fish culture station in Nebraska, Wyoming, Montana, or the Dakotas.



Figure 3.3.39 This photo, taken in the early 1900's, shows several of the Spearfish National Fish Hatchery ponds lined by stone walls, the historic stone bridge between ponds, and the 1899 hatchery building in the background. The hatchery was later renamed D.C. Booth Historic National Fish Hatchery in honor of its first superintendent, DeWitt Clinton ("D.C.") Booth. Courtesy of U.S. Fish and Wildlife Service.

Dr. B.W. Evermann lead an investigation of water resources in the area. Several localities in South Dakota were examined for their suitability. Spearfish was finally chosen because of the abundance of pure, cold spring water available for a gravity system and its accessibility to the town and regions railroad system. Evermann's favorable report led to an initial appropriation of $5,000 by Congress on June 11, 1896, to establish a fish culture station in the Black Hills at Spearfish, South Dakota.

Land was purchased June 10, 1898, at the entrance to Ames Canyon on the southern boundary of the City of Spearfish. Between 1899 and 1904, workers under the direction of DeWitt Clinton ("D.C.") Booth, first superintendent of the hatchery, built a series of spring-fed trout runs to feed, store, and protect young fry until they were large enough to be planted in Black Hills streams. Hatchery employees also created extensive limestone retaining walls to channel Ames Canyon floodwaters away from the facility and to protect the hatchery grounds from potential high water on neighboring Spearfish Creek. These intensive efforts to manipulate the hydrology of Ames Canyon and the surrounding springs eventually paid off. By 1909, the Spearfish National Fish Hatchery produced two million trout annually, and its system of spring-fed catchment basins brought sustainability to fingerling production while the stone channels and retaining walls protected the hatchery from the devastating floods that often visited the Spearfish Valley (Henris, 2015).

The hatchery was built because of access to powerful springs. Those springs were partially replenished by Spearfish Creek, but that was before Homestake Mining Company built its hydroelectric plant. After the plant was completed in 1912, about 7 miles of Spearfish Creek dried up as water was diverted from the creek bed through a tunnel feeding the power-generating turbines. This massive diversion lowered the local water table, causing the original Ames Canyon springs to diminish to a trickle. Because the springs failed, the facility was forced to pivot. The hydroelectric plant is still operating today, and the hatchery primarily utilizes surface water pumped out of Spearfish Creek. In recent years, the hatchery has frequently requested to drill deep wells to escape river-borne invasive species. (Sierra Ferguson, Black Hills Pioneer, February 14, 2026; To secure its future D.C. Booth needs a new well and some guarantees from the city | Local News | bhpioneer.com.)



Figure 3.3.40 Spearfish Hydroelectric Plant, built in 1912 by Homestake Mining Company, is now owned by the city of Spearfish. Photo courtesy of John Mitchell, from article by Sierra Ferguson, Black Hills Pioneer, November 25, 2023, "Preserving the past, planning for the future: Spearfish Hydroelectric Plant to undergo full inspection, upgrades"; Preserving the past, planning for the future | Local News | bhpioneer.com

The most devastating effect on the pristine, spring-fed streams in the Black Hills was the introduction of the cyanide process in extracting gold. Cyanide refining democratized mining in the region. It made refractory ore districts that were previously not economically viable to mine profitable again, as gold could now be extracted for

as little as two dollars per ton of ore. The cyanide process allowed large-scale mining to advance into new watersheds, most notably along Battle Creek near Keystone and along the small streams that drained the southeastern limestone rim of Spearfish Canyon. Cyanide, unfortunately, proved lethal both to trout and to the plants that supported the biota they fed upon. For example, in 1901 an accidental release of wastewater from the Spearfish Company cyanide mill killed thousands of trout in Spearfish Creek (Henris, 2015).

Franklin R. Carpenter, dean of the South Dakota School of Mines in Rapid City and an early pioneer in pyritic smelting, best articulated the mining industry's attitude toward piscatorial endeavors in the Black Hills. "The objection to the defiling of certain streams also by reason of fish culture," wrote Carpenter in 1905, "is, in the opinion of the writer, absurd in [the] extreme, and may prevent the opening of a million dollar mine for the preservation of a hundred dollars worth of fish." (Henris, 2015).

As explained by Rahn and Gries (1973), springs occur over a considerable reach of Spearfish, Little Spearfish, and East Fork of Spearfish Creeks, and there are many diversions. The total flow of Spearfish Creek at the U.S. Geological Survey gaging site in Spearfish is 52.43 cfs, of which about 39.83 cfs is spring discharge by base flow. Of this spring flow, about 11.60 cfs comes from Little Spearfish Creek, 9.12 cfs from upper Spearfish Creek, and 9.51 cfs from East Fork Spearfish Creek. The total of these three sites is 30.23 cfs. The difference between the 39.83 cfs base flow at Spearfish and 30.23 cfs is 9.60 cfs, which comes from springs along Spearfish Canyon below Cheyenne Crossing.

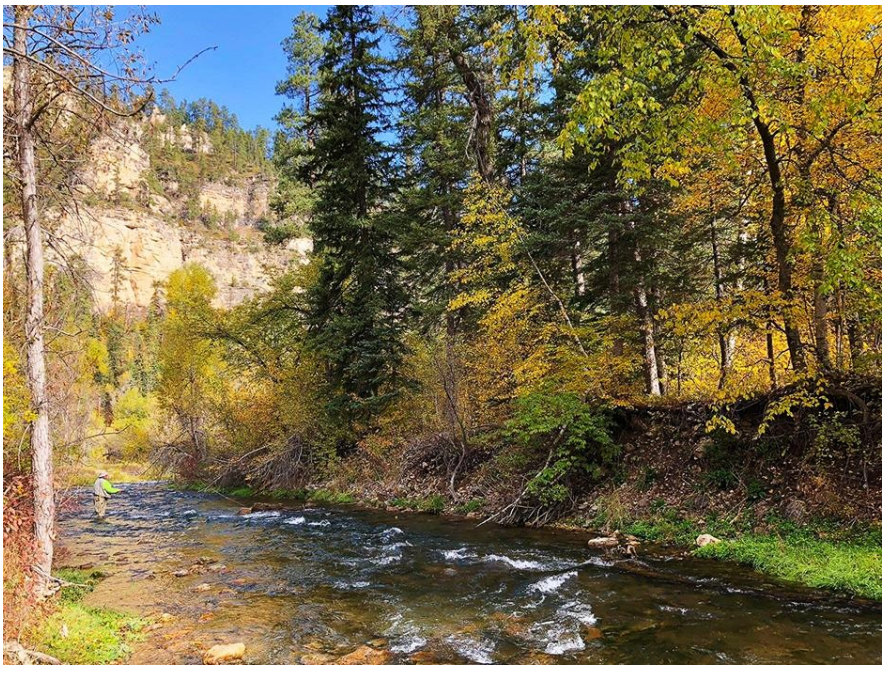


Figure 3.3.41 Spearfish Canyon, Black Hills National Fores. Photograph by Beth Doten, courtesy of U.S. Forest Service

Excerpts below are from an article by Cory Heidelberg, published in South Dakota Magazine on December 14, 2011; There's Gold in Spearfish Canyon: Leave It There! - SouthDakotaMagazine.com

Last month, the state Board of Minerals and Environment approved Wharf Resources' proposal to mine more gold near Terry Peak. Our state officials are now talking with Valentine Mining about gouging out another chunk of the Black Hills. Gold miners would like to make the Ragged Top Mountain area, just over the rim of Spearfish Canyon near Savoy, a little more ragged.

Ragged Top was first mined back in 1886. Gold mining there petered out after 1914, since which time Spearfish Canyon has become known as one of the most scenic drives in South Dakota. Tourists flocking to the canyon for the fine fall colors make September one of Spearfish's best months for sales tax revenue.

The Ragged Top site is on a list of what the state Department of Environment and Natural Resources calls "Special, Exceptional, Critical or Unique lands." State statute defines such lands as territory of great ecological sensitivity and influence and/or "scenic, historic, archaeologic, topographic, geologic, ethnologic, scientific, cultural, or recreational significance."

Spearfish Canyon region is clearly special, exceptional, critical, and unique. Frank Lloyd Wright recognized this fact during a trip through the canyon in 1935:

"But how is it that I've heard so little of this miracle and we, toward the Atlantic, have heard so much of the Grand Canyon when this is even more miraculous. All the better eventually ... that the Dakota are not on the through line to the Coast ... My hat is off to South Dakota treasures."

Expanding the non-sustainable gold industry makes little sense. We could rev up another gold mine in the Northern Hills, drive more industrial traffic up Highway 14A, and hack up some more trees and mountaintops, just for the sake of increasing the supply of a metal used mostly for non-essential purposes. But we can only mine gold once, and such industry threatens the other natural assets of the region. The autumn rush for golden leaves enriches Spearfish Canyon and its visitors time after time, for as long as we preserve the area's natural beauty.



Figure 3.3.42 Spearfish Canyon, Black Hills National Forest. Photograph courtesy of U.S. Forest Service.

Booth Society Land Management (https://dcboothfishhatchery.org/land-acquisition/)

The Booth Society capital campaign was a success!

We proudly made the final payment on our loan in September 2020. Our goals and objectives, as presented to the community during our fundraising efforts, of our 64 acre land acquisition continue to be as follows:

- *First and foremost is the protection of the hatchery ground water supply which is important for fish culture purposes and the viability of the hatchery's existence into the future.*
- *Conservation of open space for the preservation of native habitat and the species that utilize it.*
- *Prevent the construction of a housing development overlooking and ruining the aesthetics of a historic property and popular tourist attraction.*
- *Access for educational groups.*
- *Public access in the form of a pedestrian walking trail which may include interpretive signage.*

McNenny Fish Hatchery, and Crow Creek Springs, Lawrence County

Numerous springs occur at the McNenny State Fish Hatchery along Crow Creek. The hydrogeology of Crow Creek in the northern Black Hills is defined by its placement over a major artesian discharge zone. According to the U.S. Geological Survey Black Hills Hydrology Study (Carter et al., 2002), Crow Creek forms part of the single largest artesian spring discharge area in the region, alongside Cox Lake. While the springs themselves issue from the Spearfish Formation, their high discharge rate originates primarily from two major underlying regional bedrock units: the Madison (Pahasapa) Limestone and the Minnelusa Formation. This mechanism is confirmed by the presence of artesian wells at the fish hatchery. The springs cause unusual sand boils along Crow Creek and have eroded out large areas such as Cox Lake and Mirror Lakes.

According to the U.S. Geological Survey Black Hills Hydrology Study (Carter et al., 2002), Crow Creek forms part of the single largest artesian spring discharge area in the region, alongside Cox Lake. The USGS notes that flow variability for the Crow Creek spring complex is small, with a constant water temperature of 52°F (Rahn and Gries, 1973). This unique thermal signature creates a localized microclimate where the water never freezes in the winter, supporting the McNenny State Fish Hatchery. Three artesian wells and springs provide nearly 4,000,000 gallons of water daily. After being used at the hatchery, the water flows into Crow Creek (https://gfp.sd.gov/mcnenny/).



Figure 3.3.43 Upper Mirror Lake, pictured here, and Lower Mirror Lake are on the grounds of the McNenny State Fish Hatchery and are formed by direct discharge of artesian springs. Note exceptional water clarity, which is the same year-around. Photo courtesy of Black Hills Hiking, Biking and More https://www.blackhillshikingbikingandmore.com/upper-mirror-lake.

3.4 Utah, and Nevada

3.4.1 Utah

In 1971, J.C. Mundorff of the USGS, in cooperation with the Utah Geological and Mineralogical Survey (now Utah Geological Survey; https://geology.utah.gov/), published a comprehensive report on the State's nonthermal springs. For this report a nonthermal spring is defined as one having a temperature that does not exceed the mean annual temperature of the surrounding area by more than about 10°F (5°C). Currently, the official database maintained by the Utah Geospatial Center contains 16,719 springs (https://gis.utah.gov/products/sgid/water/nhd-springs).

Most major springs which have a discharge rate of several cubic feet per second (cfs) or more are in or near carbonate mountain ranges or plateaus where precipitation is much greater than in other parts of the State. These springs are concentrated in several main regions:

1. Bear River Range (Northeastern Utah)

The largest and most well-studied karst springs in the state discharge from Paleozoic-age carbonate rocks in the Bear River Range ("Lower Carbonate Aquifer"). Source aquifers are developed primarily in the Garden City Formation (Ordovician-age) and Mississippian-age limestones. Key springs are Ricks Spring, Dewitt Spring, Spring Hollow, and Logan Cave Spring which all discharge into the Logan River. Dewitt Spring is a significant portion of the municipal water for the City of Logan.

2. Located on the southern slope of the Uinta Mountains is the Ashley Karst National Recreation and Geologic Area which hosts several major springs discharging from Mississippian limestones. ("Upper Carbonate Aquifer"). The best known is Ashley Spring, the principal municipal water source for the Ashley Valley and Vernal area in Uintah County, in the Green River Basin in northeastern Utah near the Colorado and Wyoming state borders. It is fed by a complex karst system that captures surface flow from losing streams like Brush Creek.

3. Basin and Range (western Utah)

In the remote West Desert, carbonate (limestone and dolomite) aquifers within the mountain ranges and beneath basin fill aquifers feed several large spring complexes, some of which are thermal, reflecting deep regional flow in Paleozoic rocks. The best known are springs in Snake Valley, including Gandy Warm Springs just north of Garrison, and roughly an hour's drive east of Baker, Nevada.

4. Markagunt Plateau in southwestern Utah

The largest instantaneous discharge observed at any spring in Utah was at Mammoth Spring in southwestern Utah (314 cfs; Wilson and Thomas, 1964, p. 24). Discharges exceeding 100 cfs have been observed at Swan Creek Spring in northern Utah, and Big Brush Creek Spring in northeastern Utah. Maximum discharges of several other springs range from 25 to 90 cfs. Maximum discharges generally are during or within a few weeks after the main period of snowmelt, which is usually from late April to the middle of June.

The lack of continuous records of discharge for nearly all springs in Utah prevents any statewide classification of springs by magnitude. Mammoth Spring near the town of Mammoth Creek, and Swan Creek Spring near Garden City are recognized as the largest springs in Utah. If the average discharge rate is the basis for classification of spring magnitude, then no first magnitude springs exist in Utah. The number of second magnitude springs in Utah

is not known but is probably less than a hundred; most of these springs are near the minimum average discharge required for designation as second magnitude (Mundorff, 1971).

Mammoth Spring, Garfield County

Numerous springs discharge from the volcanic rocks and underlying limestone on the Markagunt Plateau in southwestern Utah including Mammoth Spring, the largest and one of the most variable springs in Utah, with discharge that ranges from less than 5 to more than 300 cfs. The Plateau lies at an altitude of about 9,500 feet, largely within Dixie National Forest. It is capped primarily by Tertiary- and Quaternary-age volcanic rocks that overlie Paleocene- to Eocene-age limestones of the Wasatch (Claron) Formations, which form escarpments on the west and south sides of the plateau.

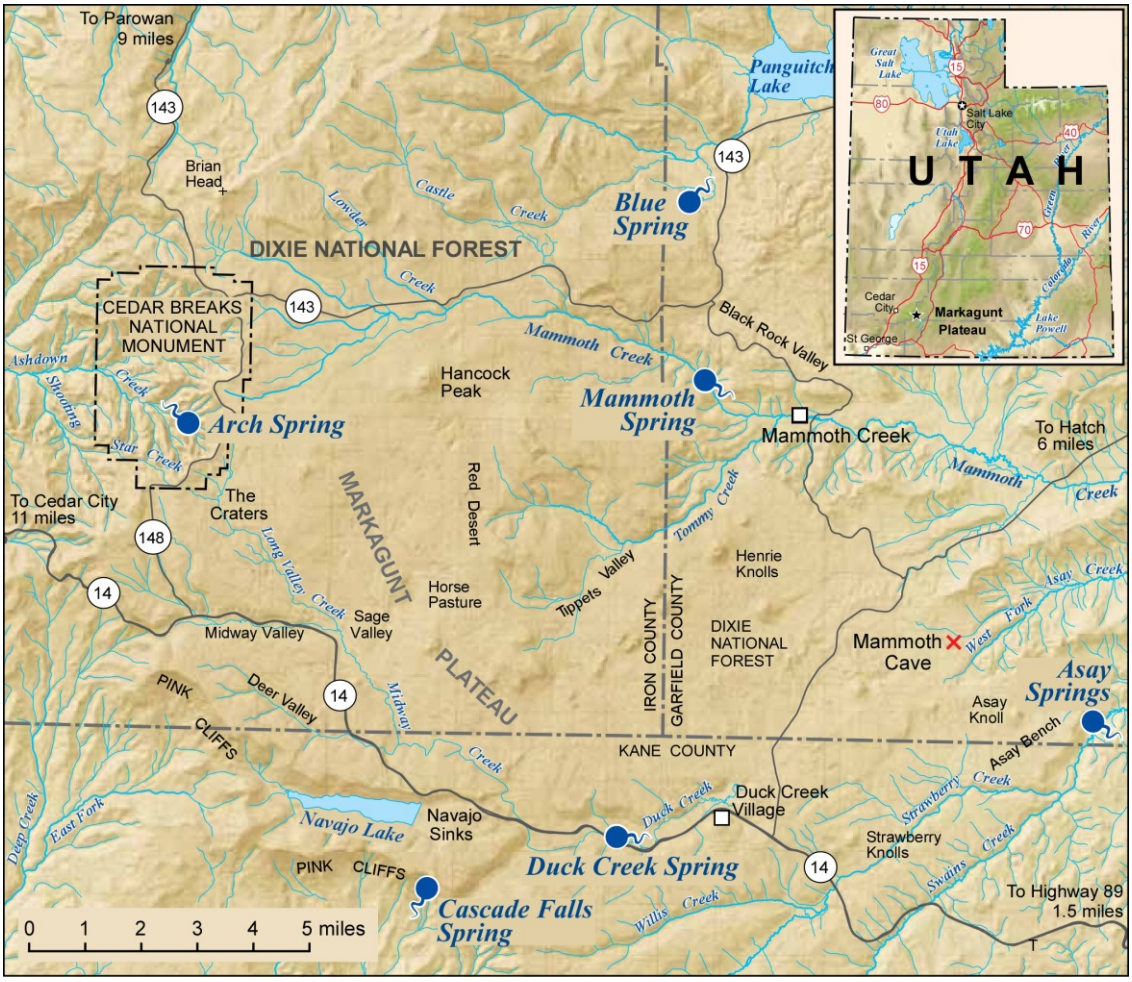


Figure 3.4.1 Locations of Mammoth Spring, the largest spring in Utah, and other large springs on the Markagunt Plateau in southwestern Utah. Modified from Spangler, 2012.

As noted by Bowers (1972), certain confusion exists when naming the sedimentary formations in Utah such as those on the Markagunt Plateau and exposed in the Cedar Breaks National Monument. The rocks of latest Cretaceous to early Tertiary age in the southern High Plateaus of Utah have commonly been referred to as the Wasatch Formation, whereas in the eastern Basin and Range province the name Claron Formation has been used

for similar strata of Eocene (?) age. Use of the name Wasatch for lower Tertiary rocks in southern Utah has been questioned by some geologists because of differences in age and lithology from the Wasatch Formation (Eocene) in northeastern Utah. Some prefer the name Claron Formation, which has been used for similar strata west of Cedar Breaks National Monument. Other names have been proposed as well.

Results of dye-tracer tests performed by USGS indicate that recharge to Mammoth Spring largely originates from southwest of the spring and outside of the watershed for Mammoth Creek, particularly in areas where large sinkholes and losing streams are present. This includes Midway, Sage, and Long Valleys, and the Horse Pasture, Hancock Peak, and Red Desert areas.



Figure 3.4.2 Mammoth Spring at average high flow during the snowmelt runoff period on the Markagunt Plateau, southwestern Utah. Spring flow diminishes to less than 10 cfs during periods of baseflow. From Spangler, 2012.

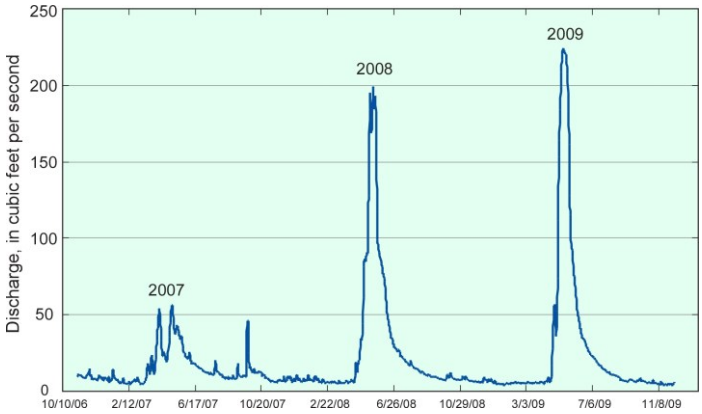


Figure 3.4.3 Daily mean discharge for Mammoth Spring, Markagunt Plateau, southwestern Utah, November 2006 to December 2009. From Spangler, 2012.

A significant component of recharge to the spring takes place by both focused and diffuse infiltration through the basalt and into the underlying Claron Formation. Losing reaches along Mammoth Creek are also a source of rapid recharge to the spring.

Maximum groundwater travel time during the snowmelt runoff period from focused points of recharge as far as 9 miles away and 1,900 ft higher than the spring, was about 7 days, indicating a velocity of more than a mile per day. Samples collected from Mammoth Spring during baseflow conditions and analyzed for tritium and sulfur-35 showed that the groundwater in storage is relatively young, with apparent ages ranging from less than 1 year to possibly a few decades. Ratios of oxygen-18 and deuterium also showed that water from the spring represents a mixture of waters from different sources and altitudes. Based on evaluation of dye-tracer tests results and relations to adjacent basins, the recharge area for Mammoth Spring probably includes about 40 square miles within the

Mammoth Creek watershed as well as at least 25 square miles outside and to the south of the watershed (Spangler, 2012).

The land surface in some areas of the Markagunt Plateau can be characterized as a *pseudokarst* terrain (see Chapter 2.10 for explanation). In these terrains, karst-like features such as sinkholes and caves, are produced by non-solutional processes, such as surface collapses into lava tubes, which previously served as conduits for molten lava (Figure 3.4.4-*Left*). In the southwestern part of the plateau, particularly between the Red Desert and an area known as The Craters (see map in Fig. 3.4.1), a unique terrain is present that is characterized by large sinkholes or dolines as much as 1,000 ft across and 100 ft deep (Fig. 3.4.4-*Right*). Most of these sinkholes are related to dissolution of limestone in the underlying Claron Formation and subsequent collapse and (or) subsidence of the basalt, rather than collapse into lava tubes. Sinkholes also are developed in the Claron Formation where it is not covered by basalt, particularly in the areas north and southeast of Navajo Lake (Spangler, 2012).



Figure 3.4.4 Typical pseudokarst features on the Markagunt Plateau, southwestern Utah. *Left*: Entrance passage in Mammoth Cave, formed by collapse of the roof of the lava tube allowing access to more than 2,000 feet of passage. *Right*: Sinkhole in the Red Desert area of the Markagunt Plateau. Dissolution of the underlying Claron Formation has resulted in subsidence and (or) collapse of the basalt to depths up to 100 feet. Photography and text from Spangler, 2012.

Many of the outcrops of basalt, particularly the flows that have broken into jagged blocks, are permeable enough that they can absorb the water of maximum storms or maximum snowmelt without runoff, and the water percolates downward rapidly from the bare outcrops. In other places, vegetation has obtained a foothold, even developed a forest cover, but the soil is capable of absorbing precipitation at high rates with negligible runoff (Wilson and Thomas, 1964).

Cascade Falls Spring, Kane County

Cascade Falls Spring discharges from the limestone of the Claron Formation which forms Pink Cliffs, the southern escarpment of the Markagunt Plateau (see map in Fig. 3.4.1). In the 1950s and 1960s, USGS performed detailed hydrologic and hydrogeologic investigations of the area because several proposals for additional development and use of the water of Navajo Lake led to controversies and raised questions regarding the total water supply in arid regions downstream where competition for municipal, industrial, and irrigation water supplies was very intense.

The entire outflow of Navajo Lake, located in a closed basin on the plateau less than two miles northwest of Cascade Falls Spring, disappears underground. During parts of low-water years, this outflow was great enough to

completely drain the lake until it was artificially impeded by a north-south dike about 17 feet high constructed across the lake; the dike separates the western three-fourths from the sink area and creates a permanent lake for fish propagation and human recreation (Wilson and Thomas, 1964).

The lake receives subterranean inflow from an area appreciably larger than that of the topographic basin in which it lies. It is unique in that large quantities of surface water escape from its eastern end through a sink area called Navajo Sinks. Hydraulic tests and dye tracing showed that the sinks are connected by underground channels developed in the limestone of Claron Formation with springs in two adjoining river basins—Duck Spring in Duck Creek in the east, a tributary of the Sevier River in the Great Basin, and Cascade Falls Spring in the south which flows to Provo Deer Creek, headwaters of the Virgin River in the Colorado River basin (Wilson and Thomas, 1964).

The discharge of Cascade Falls Spring increased 1 hour after water was released to the sink area and the discharge of Duck Creek Spring increased 12 hours later. Fluorescein dye placed in the sinks appeared at Cascade Falls Spring within 8.5 hours and at Duck Creek Spring in 53 hours. All the water entering Navajo Sinks eventually discharged from Cascade Falls and Duck Creek Springs when sufficient time was allowed to drain the added storage from the groundwater reservoir. The apportionment was 60 percent to Duck Creek Spring and 40 percent to Cascade Falls Spring.

Water issuing from Duck Creek Spring flows about 2.5 miles eastward and enters Duck Creek Sinks. During August 1954, water released to Duck Creek Sinks caused an increased flow from Lower Asay Spring in 9 hours, and fluorescein dye showed a travel time of 68 hours. Lower Asay Spring is a major contributor to the base flow of the Sevier River. The observed rates of travel of fluorescein dye from Navajo Sinks to Cascade Spring ranged from 5.6 to 12.5 fpm (feet per minute); the velocity to Duck Creek Spring was 5.8 fpm.



Figure 3.4.5 Discharge of Cascade Falls Spring from the Claron Formation along the Pink Cliffs, Markagunt Plateau. The cave from which the spring discharges is developed along north, east, and northwest-trending joints and extends for more than 1,000 feet into the plateau. Photographs courtesy of Utah Geological Survey. Available at https://geology.utah.gov/map-pub/survey-notes/geosights/cascade-falls/

Renewed volcanic activity during the Pleistocene Epoch produced numerous cones and flows of basalt in the Navajo Lake region. Some of the lava flowed down the preexisting valleys, and some flowed across and formed barricades in former drainage channels. The broad valley whose lower part is now occupied by Asay Creek has been barricaded in two places by basalt flows. Duck Creek occupies the part of the valley between the two barricades, and Navajo Lake occupies the part above the upper basalt barricade. The original eastward drainage is maintained solely by underground channels between Navajo Sinks and Duck Creek Spring, and between Duck Creek Sinks and Lower Asay Spring (Wilson and Thomas, 1964).

Studies in the vicinity of Navajo Lake show that subsurface movement of water is primarily in solution channels rather than in a continuous body of saturated sediments. The numerous sinks and springs in the Navajo Lake region also indicate the existence of various compartments or channels that may be isolated or very indirectly connected with each other. One example of such isolation is provided by the two Asay Springs: dye tracing tests show that one is connected with Duck Creek Sinks, the other completely independent (Wilson and Thomas, 1964).

Cascade Falls Spring is located just south of State Highway 14 which can be accessed from I-15 in Cedar City on the west or from U.S. Highway 89 at Long Valley Junction on the east. To get there from I-15 and Cedar City, head 27 miles east on Highway 14. After passing the Navajo Lake scenic pull-out with its descriptive signs, turn right (south) on the road to the lake. After 0.4 miles, the road splits. The right fork goes on to the Navajo Lake boat docks and lodges, and the left fork goes approximately three miles to the Cascade Falls overlook and trail parking lot. The trail to the falls is approximately 1/2-mile one way.

Cascade Falls Spring is one of three springs featured in an outstanding series of GeoSights articles produced by the Utah Geological Survey. The articles, which highlight lesser-known geologic wonders in Utah with photographs, are available at https://geology.utah.gov/apps/geosights/?page=GeoSights

Bear River Range Springs

The alpine Bear River Range, which stretches north-south between the City of Logan in the west and Bear Lake in the east, consists in large part of a thick sequence of carbonate rocks (more than 3,000 feet of limestone and dolomite) that range in age from Cambrian to Mississippian (Dover, 1987). The sequence is composed of eight principal geologic units, all of which are capable of transmitting water along dissolution-enhanced fractures, faults, and bedding planes. The formations make up the upper part of the Logan Peak syncline in the Logan Canyon area, a large regional structure that influences the movement of groundwater in much of the western portion of the range (USGS, 2004).

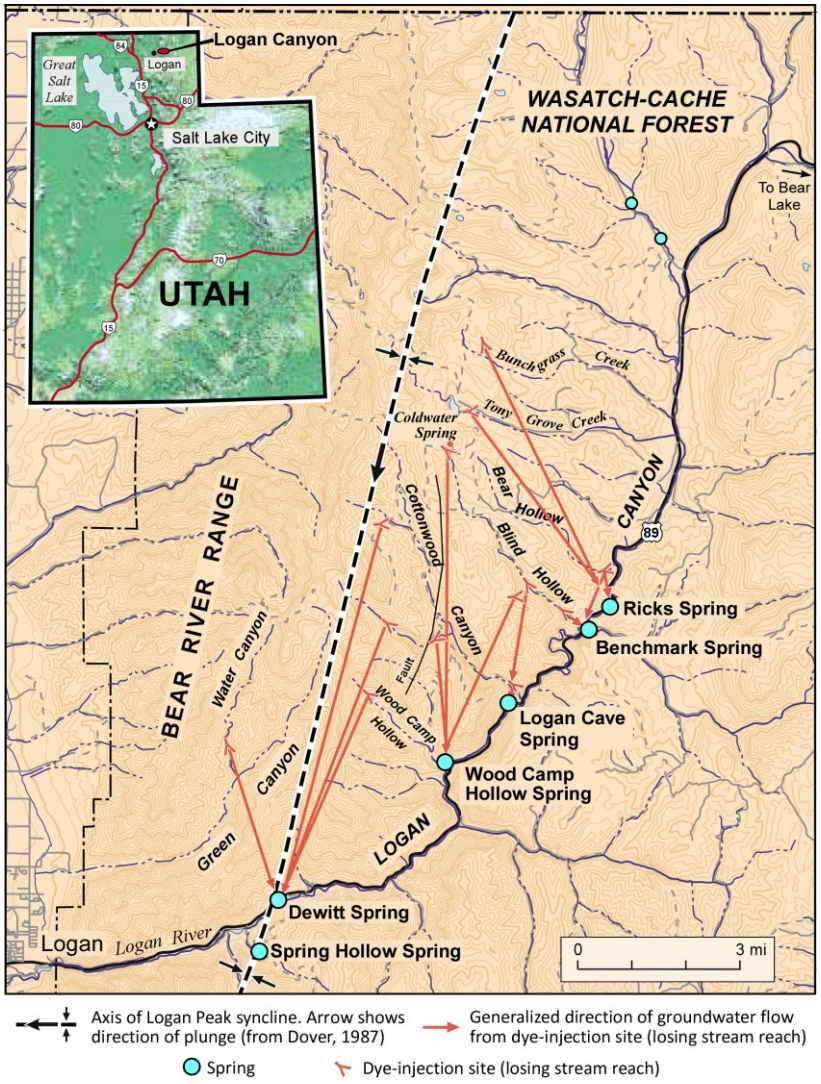
Annual average precipitation in the area is 61 inches as reported by the Tony Grove Lake SNOTEL station (Natural Resources Conservation Service (NRCS), 2024) with an average of 71% falling as snow.

In all of the Bear River Range, karst development is extensive and includes large springs that discharge along major streams, losing streams in tributary drainages, caves and pits, blind valleys, sinkholes, dolomite pavement, and surficial karst forms such as karren (Neilson et al., 2018; Spangler, 2001, 2019). Glaciation occurred above 8,000 ft during the Pleistocene, resulting in destruction of karst landforms that developed during interglacial periods (Wilson, 1979).

Speleothem age-dating, fluvioglacial deposits in caves, and deranged topography indicate that existing karst features, particularly caves, are largely remnants of former karst landscapes. Karst systems in alpine terrains are substantially different from those in relatively flat-lying strata in more temperate regions. Characteristics of alpine

karst systems include a large component of vertical solution development and a thick unsaturated (vadose) zone, steep hydraulic gradients, spring discharge that responds primarily to snowmelt runoff, pit development in high-altitude meadows, and cold-temperature dissolution of carbonate rocks (USGS, 2024).

Discharge from the carbonate karst aquifer in the western portion of the Bear River Range is primarily from large springs along the Logan River which is the principal base level stream for groundwater discharge (Fig. 3.4.6). Three second magnitude (average discharge between 10 and 100 cubic feet per second (ft^3/s)) and two third magnitude (average discharge between 1 and 10 ft^3/s) springs, along with several smaller springs, discharge along the north and west sides of the river. These include Dewitt, Wood Camp Hollow, Logan Cave, and Ricks Springs. Only one large (second magnitude) spring, Spring Hollow Spring, is known to discharge along the south side of the river. Collective discharge of the springs provides a substantial component of streamflow in the Logan River.



Spring discharge responds primarily to snowmelt runoff, with peak flow from late spring to early summer and base flow during the winter months (Figure 7.80). Recharge to the carbonate aquifer occurs through point sources (sinkholes and sinks), as seepage losses through fluvioglacial deposits, and as diffuse infiltration. Based on dye tracing, recharge areas for Dewitt, Wood Camp Hollow, and Ricks Springs are estimated to be between 7.5 and 15 square miles and as much as 3,200 feet higher than the altitude of the springs. Results of dye tracing indicate maximum groundwater travel times of 8 to 31 days from losing streams as far as 7.2 miles. Dye tracing also indicates that surface water drainage basins generally do not coincide with groundwater basins (Spangler, 2001).

Figure 3.4.6 Generalized groundwater flow paths to selected springs, based on dye tracing results, Logan Canyon, Utah. Modified from Spangler, 2001.

According to Kolesar et al. (2006), chemical and flow data suggest that aquifer characteristics are different north and south of the Logan River. Spangler's (2001) data show that the northern part of the aquifer is a very well-developed karst drainage system, with estimated flow velocities on the order of a thousand feet per day or greater. Mineral saturation indices (log(IAP/Ksp)) are indicators of the degree to which the water is saturated with respect

to particular minerals. A saturation index of 0 indicates saturation or equilibrium, a negative number indicates undersaturation, and a positive number indicates supersaturation. The mean calcite saturation index in water from DeWitt Spring is 0.2 and for quartz 0.1, slightly supersaturated, consistent with a relatively short period of water-rock interaction. However, the mean calcite saturation index in water from the spring at Spring Hollow, south of the Logan River, is 1.2 and for quartz 2.0, an order of magnitude greater. Such a difference suggests a much lower flow velocity, more time for water-rock interaction, and thus implies a less-well-developed system.

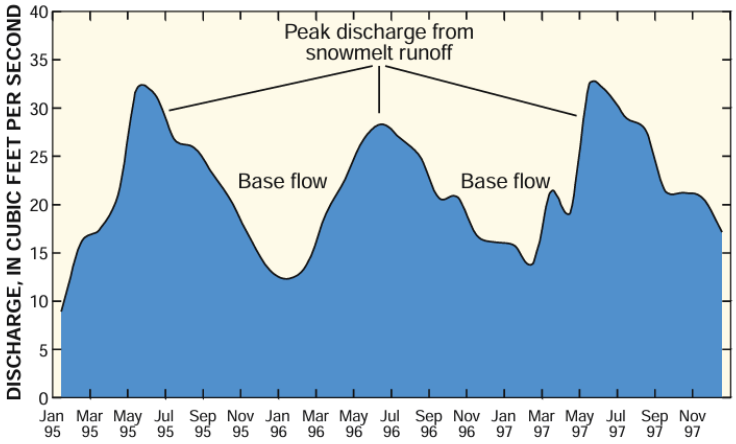


Figure 3.4.7 Hydrograph showing typical seasonal response of an alpine karst spring to snowmelt runoff, Dewitt Spring, Logan Canyon, January 1995 to January 1998 (Data from City of Logan Water Department, written communication., 1998). From Spangler, 2001.

Dewitt Spring, Cache County



Dewitt Spring discharges from the Water Canyon Formation along the flood plain of the Logan River. Water discharges from an unknown number of capped outlets and is diverted into a collection system. Water from the spring serves as a public supply for the city of Logan, about 7 miles to the west. Discharge of the spring is metered and generally ranges from about 10 to 35 cfs (Dennis Corbridge, City of Logan, written communication, 1998).

Figure 3.4.8 Dewitt Spring. Photo courtesy of City of Logan.

As discussed by Spangler (2001), four dye traces to Dewitt Spring indicate a recharge area northwest to northeast of the spring that largely coincides with the extent of the Logan Peak syncline. Groundwater movement is probably downdip along the west and east limbs of the syncline toward the axis, and subsequently southwest to the spring, which is located along the axis of the syncline where the Logan River breaches the structure.

Results of dye tracing during moderate flow indicate a maximum groundwater travel time of 22 days for losing streams in drainages 7.2 mi from, and 2,900 ft higher than Dewitt Spring. The substantial base flow of this spring relative to peak flow may indicate a large storage component that is recharged primarily from infiltration. Groundwater travel times from this diffuse component of flow are likely to be considerably longer than those from losing streams. Based on dye tracing, geology, and discharge, the groundwater basin for Dewitt Spring is estimated to be about 15 square miles (Spangler, 2001).

The spring and a secluded, riverside picnic area next to it got their name from Aaron Dewitt, one of Logan's first pioneers. The city of Logan acquired the water rights, built a reservoir, and started providing drinking water to the city's residents.

Ricks Spring, Cache County

Ricks Spring is located along U.S. Highway 89 in Logan Canyon about 17 miles northeast of (up-canyon from) the city of Logan, at mile marker 477. It lies within Uinta-Wasatch-Cache National Forest at an elevation of 5880 feet. Situated at the base of a hillside, the spring is one of the largest and most scenic along the Logan River. Water from the spring flows out of a large alcove, under Highway 89, and into the Logan River, about 150 feet from the spring. Pullouts on both sides of the highway provide parking for visitors to the spring, and a boardwalk crosses the spring run, which allows access to the rise pool in the alcove. Several signs at the spring provide information about its history and hydrology. Ricks Spring typically flows during the spring, summer, and fall months, but can have periods of no flow during the winter months, particularly during extended periods of cold weather (Spangler, 2019; Tennant and Neilson, 2024).

Since 2007 when it was first accessed, the submerged conduit feeding Ricks Spring has been explored by cave divers for about 2300 feet into the mountainside. The initial discovery and exploration of Ricks Spring Cave are summarized in an article published in the National Speleological Society (NSS) Cave Diving Section's Underwater Speleology newsletter in May/June 2008.



Figure 3.4.9 Ricks Spring rises along a nearly vertical normal fault (dashed red line) in the Ordovician Garden City Formation. Bedding dips to the southeast at about 36° near the fault. The small cave on left side of photo could have been a former spring outlet. Photograph by Larry Spangler, August 2011.

As described by Stephanie Carney in the January 2024 Survey Notes of the Utah Geological Survey featuring Ricks Spring GeoSight (available at https://geology.utah.gov/map-pub/survey-notes/geosights/geosights-ricks-spring-cache-county/), Ricks Spring has been a roadside attraction for more than 100 years. It was discovered by and named after a well-known Utah pioneer Thomas E. Ricks, who lived in the Cache Valley area in the mid- to late 1800s. At the recommendation of Mormon leader Brigham Young, he and several others began constructing a road eastward along the Logan River. The goal was to connect Cache and Bear Lake Valleys. The first "leg" of the road ended at Ricks Spring. While in Cache Valley, Ricks served as sheriff and county assessor and devoted much time and effort to building railroads.

During the winter of 1972, severe cold weather froze the river, resulting in ice jams in some locations, which caused the water to back upstream. Subsequently, Ricks Spring began to flow and then subsided when the river level dropped again, suggesting a link between the river and the spring. Later that summer, dye was added to the Logan River and was subsequently observed in Ricks Spring, confirming the connection between the river and the spring (U.S. Forest Service, Logan District Ranger, verbal communication, 2010, From Spangler, 2019).

Beginning in 1991, the USGS (Spangler, 2001) started further investigations into the source of water for Ricks Spring. Results of these investigations, based on dye-tracer tests, showed that in addition to the river, water also originated from Bear Hollow, northeast of the spring, Tony Grove Creek near Tony Grove Lake, and from Bunchgrass Creek, north of the spring. Results of the dye tracing indicated that the source area or groundwater basin for Ricks Spring extends more than 5 miles to the northwest in the Tony Grove Lake area and 2600 feet higher than the spring. Maximum groundwater travel time from the dye-injection points to the spring was about 4 weeks during low-flow conditions.

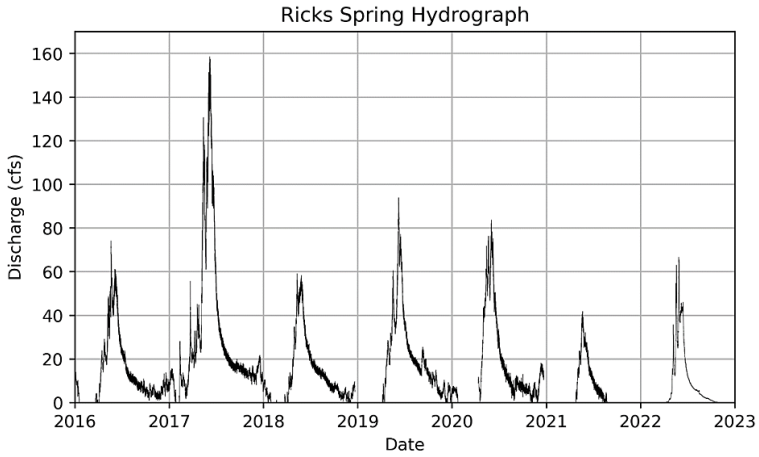


Figure 3.4.10 Ricks Spring hydrograph from 2016 through 2023 (*Logan River Observatory: Ricks Spring above Confluence with Logan River Aquatic Site (RS_CONF_A) Raw Data | CUAHSI HydroShare*, 2024). Winter gaps represent no-flow conditions. From Tennant and Neilson, 2024, Civil and Environmental Engineering, Utah Water Research Laboratory.

Swan Creek Spring, Rich County

Springs emanating from carbonate karst terrain in the eastern Bear River Range are numerous (Fig. 3.4.11) and their discharges differ by order of magnitude. Swan Creek Spring, one of the two largest in Utah, has an average discharge of about 60 cfs (Mundorff, 1971) and a maximum discharge of about 321 cfs (based on periodic measurements between 1979 and 2004 conducted by U.S. Environmental Protection Agency; Bright, 2009).

According to Kaliser (1972), the only available daily record of Swan Creek's flow was for 1944 and 1945 when Swan Creek was gaged. Estimates for instantaneous discharge of about 300 cfs have been made, however, for the spring. Rain and snowfall readily recharge the aquifer as far as 6.25 miles away and 1,500 ft above the spring outlet including through large sinkholes on the Bear River Range (Tennant and Neilson, 2024).

Recharge west of the divide also contributes to the flow of Swan Creek Spring as indicated by dyes tracing results. Dye injections have appeared at the spring outlet within 3 days of injection (Spangler, verbal comm., 2024).

Swan Creek Spring provides by far the greatest part of Swan Creek's total discharge. Another spring of significant magnitude is also discharging to Swan Creek. Figure 3.4.12 shows the Swan Creek hydrograph for part of 1945 in relation to the daily precipitation at Lifton Pump Station. Although the precipitation regime at Lifton is not exactly that of the Bear River Range, it was the closest station at which daily precipitation rates were recorded. The discharge measurements reflect the rapid response of the carbonate springs feeding Swan Creek to snowmelt during April and May and to large rainfall later.

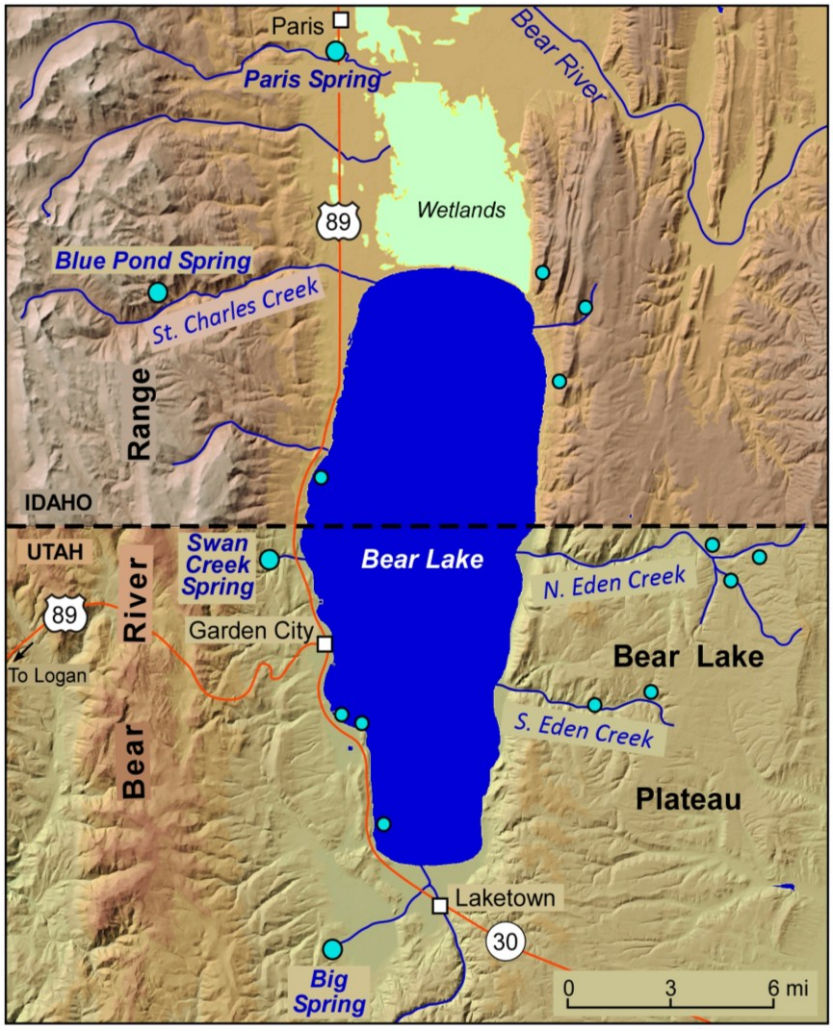


Figure 3.4.11 Large (labeled) and other springs (blue circles) in the Bear Lake area. Shaded relief map courtesy of USGS. Spring locations from Bright, 2009.

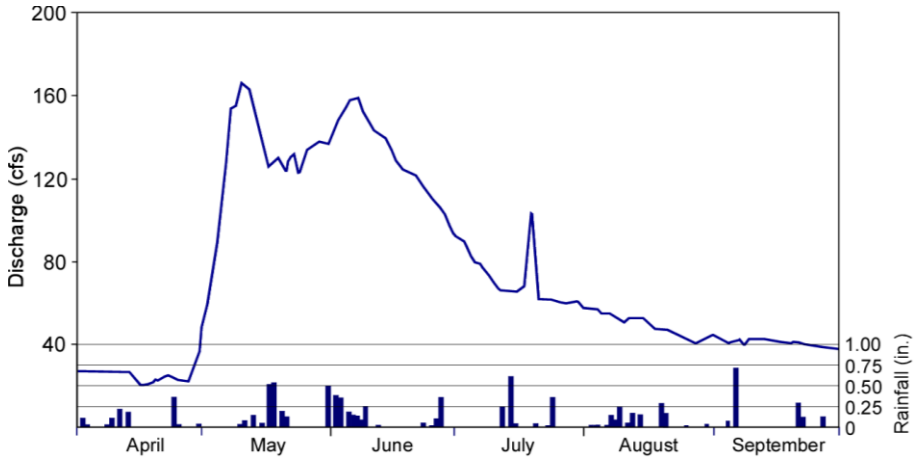


Figure 3.4.12 Relation of Swan Creek discharge to daily precipitation in 1945 recorded at Lifton Pump Station. From Kaliser, 1972.



Figure 3.4.13 Swan Creek control structure where approximately 50 cfs is allowed to naturally flow from the discharge point in the cliff band to Bear Lake while approximately 10 cfs is being diverted for agricultural irrigation. Photo by Hyrum Tennant, August 2024, Civil and Environmental Engineering, Utah Water Research Laboratory.

The impressive discharge of Swan Creek Spring is indicative of a large and well-developed karst conduit system within the Bear River Range. The sensitivity of Swan Creek Spring to rainfall events (Kaliser, 1972) indicates a strong linkage to the surface. Infiltration into the Swan Creek Spring aquifer is likely fast and, given the spring's cold temperature (42 to 50°F), passes quickly through the mountain range along shallow, and possibly short, flow paths (Bright, 2009). The spring is fed by a large solution channel (Kaliser, 1972) and is located along one of a series of north-south–trending faults in the Bear River Range, west of Bear Lake (Dover, 1995).

The solute chemistry of Swan Creek Spring varies in response to its discharge. Solutes derived primarily from dolomite dissolution dominate baseflow and solutes derived from increased limestone dissolution dominate peak discharge (Bright, 2009).

The large volume of water issuing from Big Spring (average is 26 cfs, max is 67 cfs, based on periodic measurements by U.S. EPA from 1979 to 2004; Bright, 2009) also suggests a strong connection to the Paleozoic carbonate aquifer, but the calcium and magnesium concentrations at Big Spring during assumed peak discharge (April) and base flow (September) conditions are not discharge dependent. The groundwater basin and conduit-fracture network that feeds Big Spring may be significantly different from the system that feeds Swan Creek Spring. Additional data are needed to test this hypothesis however (Bright, 2009).



Figure 3.4.14 Bear Lake contains abundant suspended microscopic particles of white-colored calcium carbonate (lime) that reflect the water's natural blue color back to the surface, giving the lake its intense turquoise-blue color. This color is why Bear Lake is known as the "Caribbean of the Rockies." These particles of calcium carbonate are mainly derived from the abundant limestone of the Bear River Range and are brought in by spring-fed streams, submerged springs, and diffused discharge of groundwater (Davis and Milligan, 2011). Photo courtesy of Bear Lake Valley Convention & Visitors Bureau.

Snake Valley Springs

The groundwater, springs, and streams formed by springs in Snake Valley are predominantly used locally for growing alfalfa and hay and enhancing habitat for livestock and wildlife. Increasing concerns about additional groundwater-development pressures for water supply of Las Vegas in Nevada have been the impetus for numerous hydrogeologic, groundwater, and wildlife resource studies in the region undertaken by the Utah Geological Survey (Menuz et al., 2014). Adequate water to supply the valley's wetlands is important to provide habitat for snail species found nowhere else on earth and wildlife species having limited habitat range (least chub and Columbia spotted frog), as well as to allow the ranching way of life to continue as it has for over 100 years.

As explained by Goodwin et al. (2021) in a report summarizing 10 years of hydrologic monitoring in the Snake Valley, the aquifers in the Snake Valley drainage basin include unconsolidated Quaternary-Tertiary sedimentary deposits, semi-consolidated to consolidated Tertiary sedimentary deposits, Quaternary and Tertiary volcanic rocks, and Paleozoic carbonate rocks (Welch et al., 2007). Unconsolidated Quaternary-Tertiary deposits make up the valley floor and mountain fronts, and mantle bedrock in the mountains. Tertiary sedimentary and volcanic rocks crop out along the mountains and underlie Quaternary-Tertiary aquifers below the valley floor. Carbonate-rock units crop out in the mountains, as well as along the mountain fronts and underlie the unconsolidated Quaternary-Tertiary aquifers in much of the region. Aquitards include Proterozoic Paleozoic quartzite, schist, and metamorphic and igneous rocks that crop out in the northern part of the southern Snake Range and in the southern part of the northern Snake Range. The mountain fronts are defined by steeply dipping, valley-side-down normal faults that created the Basin-Range topography and accommodated deposition of thick (>5000 feet in places) deposits of Quaternary-Tertiary basin fill below the valley floor (Goodwin et al., 2021).

In situ groundwater recharge to the Snake Valley aquifers occurs where mean annual precipitation is greater than 12 inches per year (Prudic et al., 2015), i.e., along the upper mountain fronts and in the mountain ranges (Fig. 3.4.15). In the mountains, recharge of snowmelt occurs into the Paleozoic carbonate-rock aquifers (Heilweil and Brooks, 2011); this groundwater moves through solution-widened fractures in the carbonate-rock aquifer into the basin-fill aquifer, or continues in the carbonate-rock aquifer below the basin fill. Groundwater recharged in the mountain block must cross the range-bounding normal faults, which localize springs along the mountain front and likely slow or divert cross-fault subsurface flow (e.g., Wilson and Guan, 2004). Surface runoff is more prevalent in mountain areas underlain by Proterozoic-Paleozoic aquitards (Heilweil and Brooks, 2011). In those areas, streams are more likely to run onto the mountain fronts and valley floors, where they provide direct recharge to the basin-fill aquifer, bypassing the range-bounding normal faults (Goodwin et al., 2021).

Springs in the Snake Valley occur along the mountain fronts and on the lower valley floors (Fig. 3.4.15; Table 3.4.1). The mountain-front springs are fault-controlled whereas those on the valley floors are contact springs, with the exception of Miller Spring in Snake Valley which is interpreted as fault controlled (Hurlow, 2014, p. 117). Discharge rates of the evaluated springs vary from about 0.3 cubic feet per second (cfs) from the upper Clay Spring and Miller Springs to 6 cfs from Dearden Springs (Table 3.4.1). Discharge rates at most springs were at their lowest values around 2012 through 2014 and increased to higher values by 2018 to 2020. Exceptions are Clay and Foote Springs, where flow has been steadily declining since 2012 (Goodwin et al., 2021).

Spring and groundwater chemistry evolves systematically from mountain front to valley floor (Hurlow, 2014). Water type along the mountain fronts is Ca-HCO_3 having low total dissolved solids (TDS) concentrations. Major-solute chemistry acquires greater Na, Cl, and TDS toward the valley floors due to ion exchange with the basin-fill aquifer, and groundwater age also increases from mountain front to valley floor (Hurlow, 2014).

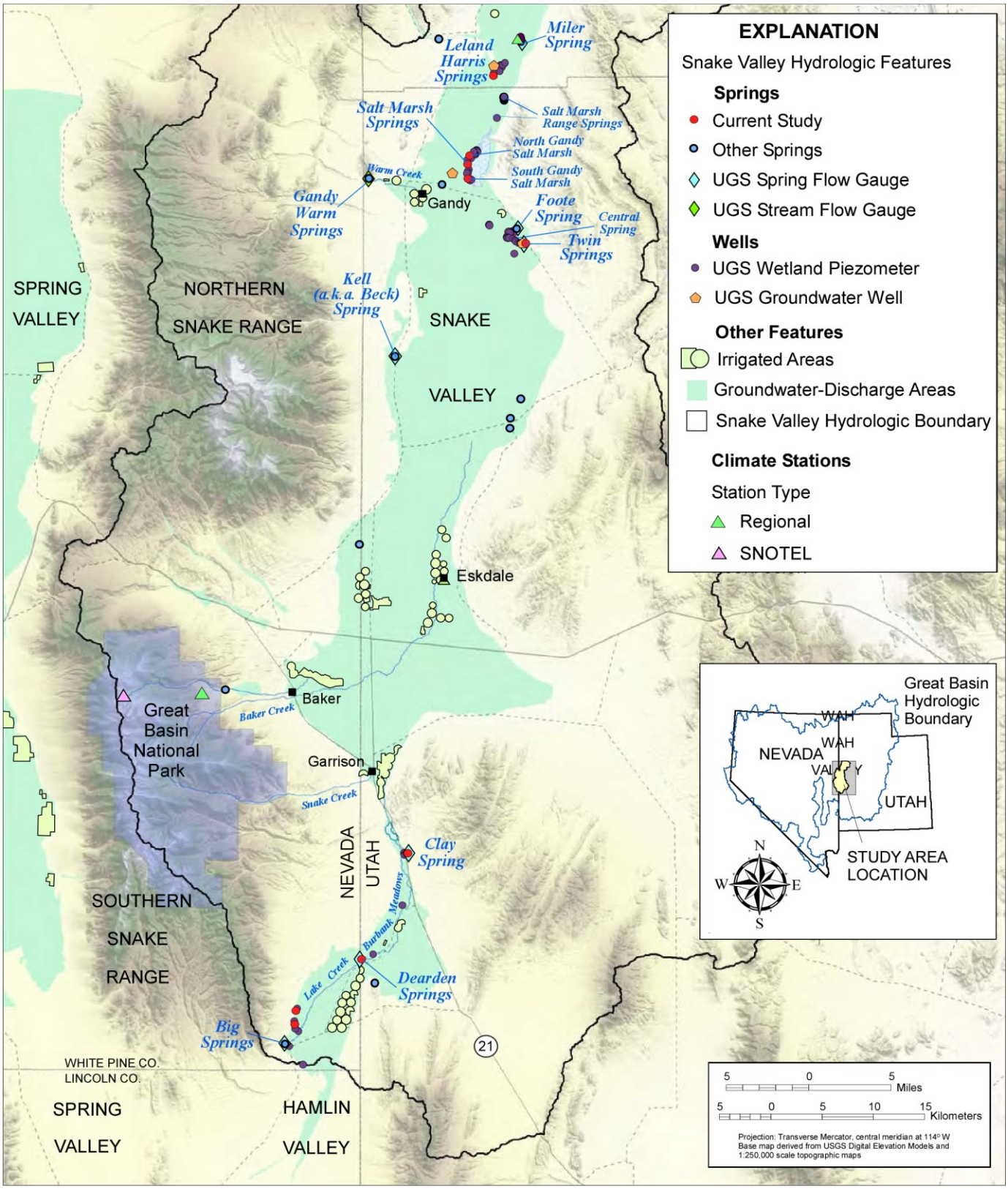


Figure 3.4.15 Map of springs and monitored wetlands in Snake Valley. Modified from Goodwin et al., 2021.

Table 3.4.1 Summary of springs that support wetlands in Snake Valley. Modified from Goodwin et al., 2021.

Spring	Wetlands System	Type*	Aquifer(s)**	Flow (cfs)
Clay Spring***	Clay Spring	Contact (?)	Basin Fill (BF)	~ 0.3
Big Springs	Big Springs	Fault	BF & Carbonate Rock	~ 3.5
Dearden (Stateline)	Burbank Meadows	Fault	Carbonate Rock	~ 6
Twin Springs	Twin Spring	Fault	BF & Carbonate Rock	~ 2.7
Foote Reservoir Spring	Foote, Foote Rd.	Fault	BF & Carbonate Rock	~ 3.5
Miller Spring	Miller Spring	Fault	BF & Carbonate Rock	~ 0.3
North Gandy Salt Marsh	North Gandy Salt Marsh	Contact (?)	Basin Fill	--
Leland Harris	Leland Harris	Contact (?)	Basin Fill	--

* Based on Kistinger et al. (2009) and Hurlow (2014).

** Dearden Springs discharge directly from bedrock. Big, Twin, Foote, and Miller Springs discharge from basin fill but are interpreted to be sourced primarily from the Paleozoic carbonate-rock aquifer.

***Smaller (upper) Clay Spring. The flow of larger, lower Clay Spring is not monitored. This spring likely discharges from the Paleozoic karst aquifer.

Groundwater pumping for irrigation in Snake Valley ranged from about 4000 to 7500 acre-feet per year between 1974 and 2005, then increased significantly during the following decade, averaging about 22,000 acre-feet per year from 2010 through 2014. Groundwater levels in monitoring wells within 5 miles of the main areas of agricultural pumping have been declining by 0.2 to 0.6 feet per year since the mid-1990s (Hurlow, 2014; Hurlow and Inkenbrandt, 2016). Rates of decline in some of these monitoring wells increased markedly during the time of increasing pumping. Long-term water level records show declines since the late 1980s (Hurlow, 2014; Hurlow and Inkenbrandt, 2016).

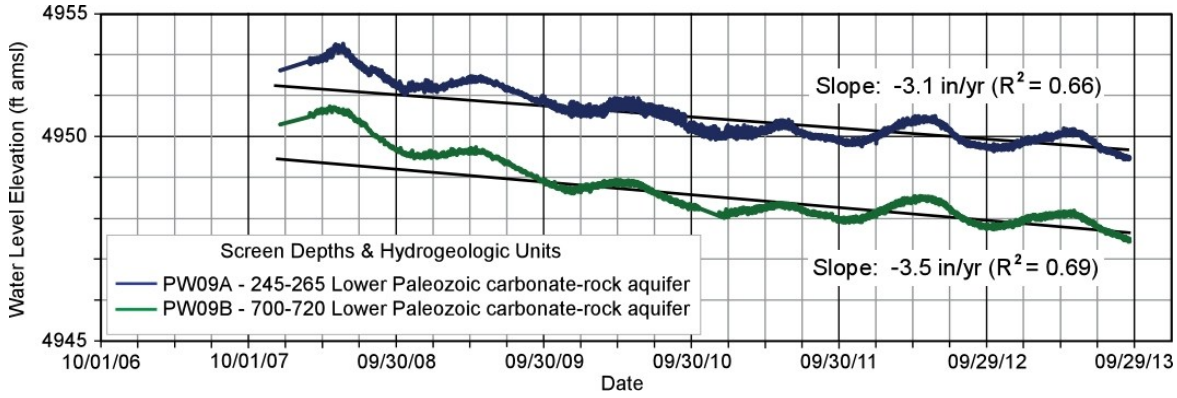


Figure 3.4.16 Groundwater levels east of Eskdale in Snake Valley from UGS data. Site 9 shows seasonal variations in groundwater levels in the Paleozoic carbonate-rock aquifer resulting from groundwater pumping, superposed on a gradual decline of about 3.1 to 3.5 inches per year which indicates a likely mining of the aquifer. Graph from Hurlow and Jordan, 2014.

Collectively the data show that in Snake Valley agricultural areas, groundwater is currently being withdrawn at rates greater than it can flow to the discharge areas. Proposed large-scale groundwater development projects would greatly expand the area and magnitude of groundwater withdrawal and discharge capture, resulting in lower groundwater levels and reduced spring flow throughout southern and central Snake Valley (Hurlow and Jordan, 2014).

Clay Spring, Millard County

Clay Spring consists of two distinct discharge areas: an upper smaller spring, the flow of which has been monitored by the UGS since 2009 (Site SF29, Snake Valley Surface Water Data - Utah Geological Survey) and a lower, larger spring pictured on Fig. 3.4.17-*Top*. The lower spring pool is remarkably deep—so much so that its exact depth remains unknown to locals and researchers (https://protectsnakevalley.org/watertour.html). The spring originally drew the Clay family, some of the very first settlers, to Snake Valley. In the 1930s, the Civilian Conservation Corps (CCC) utilized the spring to supply water for their regional camp. The Utah Geological Survey actively monitors Clay Spring because its wetlands support sensitive species. These include the least chub, the Columbia spotted frog, and over 20 species of desert snails found nowhere else on Earth. It also provides essential water for wild and domestic horses (Menuz et al., 2014).



Figure 3.4.17 *Top*: Clay Spring, basin of the lower spring. Photo courtesy of Gretchen Baker. *Bottom*: Utah Geological Survey researcher Ryhan Sempler collects water quality data at the upper Clay Spring while curious horses look on. In addition to supporting domestic and wild horses, wetlands in Snake Valley are home to two habitat-restricted species of concern, least chub and Columbia spotted frog, and over 20 species of snail found nowhere else in the world. From Menuz et al., 2014.

An example of one of the many threats facing springs and groundwater resources in the Great Basin is the fight by locals in Spring and Snake Valleys in Nevada and Utah to protect their water from the relentless thirst of Las Vegas. Various environmental and citizens groups challenged a proposed northeastern expansion of Southern Nevada Water Authority (SNWA) water supply wells and a 300-miles long pipeline.

The Utah Association of Counties petitioned the federal government, requesting that four agencies of the Department of Interior, including the Bureau of Land Management and the Bureau of Indian Affairs, file protests against the plan. In 2013, the Utah Governor Gary Herbert decided not to sign a controversial water-sharing agreement with Nevada, despite the pressure from Utah's neighbor and the threat of a lawsuit. The states negotiated for years over the division of water from the Snake Valley aquifer, which straddles the border and is home to ranching and farming communities.

The conclusions of the Utah Geological Survey assessments of likely impacts of the proposed water diversion from Snake Valley are (Kirby and Hurlow, 2005):

- Wells proposed by the Southern Nevada Water Authority will likely adversely affect groundwater conditions in nearby Utah.
- The total drawdown of groundwater near Garrison in western Millard County could be greater than 100 ft.
- The proposed pumping may change or reverse groundwater flow patterns for much of the east-central Great Basin in Utah and Nevada. The effects may eventually propagate eastward and impact discharge at important regional springs in Wah Wah Valley and Tule Valley.
- Discharge of agriculturally and ecologically important springs will decrease.

The excerpts below are from the article "*Water authority gives up 30-year proposal to pump groundwater to Las Vegas"* by Miranda Wilson published on May 21, 2020, in Las Vegas Sun. (https://lasvegassun.com/news/2020/may/21/water-authority-gives-up-30-year-proposal-to-pump/)

The Southern Nevada Water Authority is retiring a controversial, decades-old proposal to pump groundwater from rural Nevada to Las Vegas and will instead focus on water conservation measures and the pursuit of possible partnerships with other Colorado River water users.

The water authority's Board of Directors unanimously voted today to withdraw all pending groundwater importation applications, return a right-of-way associated with groundwater importation plans to the Bureau of Land Management and take other actions to move the multibillion-dollar groundwater development project — sometimes referred to as the water pipeline project— into "indefinite deferred status."

The decision is a win for environmentalists, ranchers, Native American tribes and other stakeholders mostly in Eastern Nevada who have opposed the water authority's intention to pump groundwater from White Pine, Lincoln and rural Clark counties to Las Vegas through buried pipelines. Opponents said the pipelines would harm the ecology of the region, adversely impact sites considered sacred to some Native American tribes, hurt ranching and worsen dust issues in Utah.

It appears that, for now, the Snake Valley springs in Utah are spared the destiny that many other beautiful springs in the country could not escape.

Gandy Warm Springs, Millard County

Gandy Warm Springs in Millard County is a natural desert oasis located on the western edge of Snake Valley in Utah, half a mile from the Nevada border. The springs are at the base of the southern tip of Spring Mountain (also called Gandy Mountain) and are known for crystal-clear geothermal water that remains a constant 81°F (27°C) year-round. The springs discharge approximately 9,000 gallons of water per minute. Unlike typical "hot" springs, these are "lukewarm" and best enjoyed during the warmer months of spring, summer, and fall. The Middle Pool (Fig. 3.4.18-*Right*) is the most popular with a small waterfall and a small cave to explore.



Figure 3.4.18 Upper (*left*) and Middle (*right*) Pools of Gandy Warm Springs. Photographs courtesy of Utah Geological Survey.

The source of water for Gandy Warm Springs is the northern Snake Range and perhaps locales farther west. Possible conduits to the surface for the warm water are faults located in Snake Valley. Alternatively, the springs could flow from sediments containing groundwater that is forced to the surface by the bedrock barrier of Spring Mountain. Passageways through limestone bedrock could also channel groundwater to Gandy. Chemical analysis of the spring water's isotopes suggests some older regional groundwater thousands of years old mixing with more recently recharged groundwater (https://geology.utah.gov/map-pub/survey-notes/geosights/gandy-warm-springs/).



The spring water that cascades down the slope of Spring Mountain joins a larger spring that emerges from a cave initiating the eastward-flowing Warm Creek (also called Gandy Creek). Gandy is a popular spot for locals who use the area for soaking, swimming, and baptisms.

Figure 3.4.19 Cave at the Middle Pool of Gandy Warm Springs. Photo by Corrine, Outdoor Fam Fun, available at https://www.outdoorfamfun.com/gandy-warm-springs-utah/

The rock of Spring Mountain is largely Ordovician to Cambrian-age Notch Peak Formation, a dark brownish-gray dolomite and gray limestone originally deposited on the seafloor about 500 million years ago. The springs flow from another limestone—the Cambrian-age Orr Formation. The limestones create typical karst topography characterized by cavities, alcoves, and caves; some of the spring-derived streams disappear into caves and reemerge down slope. A mile north of the springs is Crystal Ball Cave, which contains large, roundish, sparkly calcite crystals, called "nailhead spar" (thus the name Crystal Ball Cave) up to a foot thick on the cave walls, as well as stalactites and stalagmites, columns, and platforms that mark ancient groundwater levels. Many Ice Age mammal bones were collected from the cave (for example, saber-toothed cat, camel, "large-headed llama," and musk ox). Discovered in 1956, the cave is privately owned through a mineral claim on Bureau of Land Management land.

Ashley Spring Complex, Uintah County

Ashley Spring Complex is in Ashley Karst National Recreation and Geologic Area (often abbreviated Ashley Karst NRGA). Created by Congress in 2019 and administered by the United States Forest Service as part of the Ashley National Forest, the designation aims to conserve the area's karst aquifer and headwater streams, as well as geological, recreational, ecological scenic, and cultural resources. It is approximately 173,475 acres in size and lies northwest of Vernal, Utah.

The area is underlain by soluble Mississippian limestones that form a classic karst terrain of sinkholes, underground drainage, and caves. Notable features include Big Brush Creek Cave and Little Brush Creek Cave, high on the forest's south slope. Both are known for seasonal ice formations and complex subterranean passages that capture surface flow from Brush Creek.

The Ashley Spring complex is a series of springs and seeps flowing into Ashley Creek along its banks and into its channel (Godfrey, 1985). The largest spring (with greatest discharge) of the Ashley Spring complex is a municipal and irrigation water source on the east side of Ashley Creek, usually referred to as Ashley Spring (singular). The main spring has had a spring-collection box since about 1880 (Utah Division of Water Rights, 2013). The City of Vernal, Ashley Upper Irrigation Company, Island Ditch Company, Rock Point Canal and Irrigation Company, and Ashley Central Irrigation Company have an 1874 water right priority on Ashley Spring (Utah Division of Water Rights, 2013). Water is collected from the Ashley Spring diversion box (Fig. 3.4.20) by the Central Utah Water Conservancy District (CUWCD) and the Ashley Valley Water and Sewer Improvement District.



Figure 3.4.20 Upstream (north) view into Ashley Gorge; the white building in the foreground contains Ashley Spring.
From Inkenbrandt et al. 2015.

The CUWCD distributes water to Vernal City, and the Ashley Valley Water and Sewer Improvement District provides water to other smaller communities and rural domestic locations throughout the Ashley Valley area, as well as diversion to irrigation companies having priority rights to the water (Inkenbrandt et al., 2025).

Discharge from Ashley Spring (both the diverted spring water and non-diverted west side springs) varies seasonally and correlates with snowmelt. While periodic precipitation does influence spring discharge, melting snowpack is dominant. Based on flow measurements from 1944–1945 and 1954–1955 of Ashley Spring (USGS station 9266000) and Ashley Creek (USGS station 9266500), the maximum flow of Ashley Spring, occurring from May to June during snowmelt typically does not exceed 50 cfs, and minimum flow, from January to March, generally does not drop below 12 cfs. Maximum flow of Ashley Spring (including west-side springs and neglecting diverted water) based on the difference between the upstream (USGS station 9265500) and downstream (9266500) gauges, was 220 cfs; the average maximum discharge is about 70 cfs (Inkenbrandt et al., 2025). Mundorf (1971) reported a range of discharge of 15 to 90 cfs.

Horseshoe Springs, Tooele County

Horseshoe Springs are in Horseshoe Springs Wildlife Habitat Management Area, in a remote north-south trending Skull Valley in northeastern Tooele County, about 45 miles west of Salt Lake City, and 25 miles southwest of Grantsville.

The springs are entirely fault-controlled. They discharge directly along the Skull Valley faults located at the western foot of the Stansbury Mountains, the main recharge zone for the deeply seated karst aquifer developed in very thick Paleozoic carbonates. These fault zones act as highly permeable structural conduits, allowing deep, pressurized groundwater to migrate upward through tight valley-fill sediments to the surface.

Structurally, the backbone of the Stansbury Mountains is a massive, north-south-trending A-shaped fold called the Deseret anticline. The older Cambrian Tintic Quartzite forms the eroded center axis of this fold, while younger Paleozoic carbonate layers dip steeply away down the mountain flanks (Teichert, 1958; Davis, 1959; Tooker, 1983).



Figure 3.4.21 Horseshoe Springs. Photo courtesy of Alan Cressler. Available at https://www.flickr.com/photos/alan_cressler/

Horseshoe Springs is classified as a thermal spring system, maintaining a constant temperature of approximately 70°F (21°C) year-round. This stable temperature reflects the deep geothermal gradient the water encounters during its underground journey.

Horseshoe Springs is both a Wildlife Management Area (WMA) and a historic site on the California National Historic Trail located in Skull Valley, Tooele County, Utah. Two large springs come together to form a horseshoe shape, giving the site it's name. Horseshoe Springs was utilized extensively by local Goshute tribes and first visited by a white man in 1827: by legendary trapper and explorer Jedediah Smith while returning to Utah from California. In 1846, Horseshoe Springs became a stop on the Hastings Cutoff emigrant trail and was visited by the infamous Donner Reed party. Today, Horseshoe Springs is managed for its important riparian, fisheries, and widlife habitat. There is a visitor kiosk and boardwalk providing access for fishing and other recreation. There also is an interpretive display for the California National Historic Trail. (https://www.blm.gov/visit/horseshoe-springs-wildlife-management-area-wma)

3.4.2 Nevada

The results of a regional groundwater flow study in southern Nevada are shown on Fig. 3.4.22. The total rate of flow in carbonate-rock aquifers is estimated to be about 77 thousand acre-feet per year (approximately 108 cfs; note that one acre-foot-year equals 0.0014 cfs). The main discharge is at two large groups of springs, Ash Meadows and Muddy River Springs, or through the basin-fill sedimentary aquifers that partly fill valleys, while the rest flows out of Nevada into adjacent states, mostly California (Dettinger, 1989). It is interesting that, due to very low recharge (large portions of Nevada and Utah are desert terrains), the entire groundwater flow in this vast area is less than the average flow of many individual first magnitude karst springs in Florida or the Ozarks.

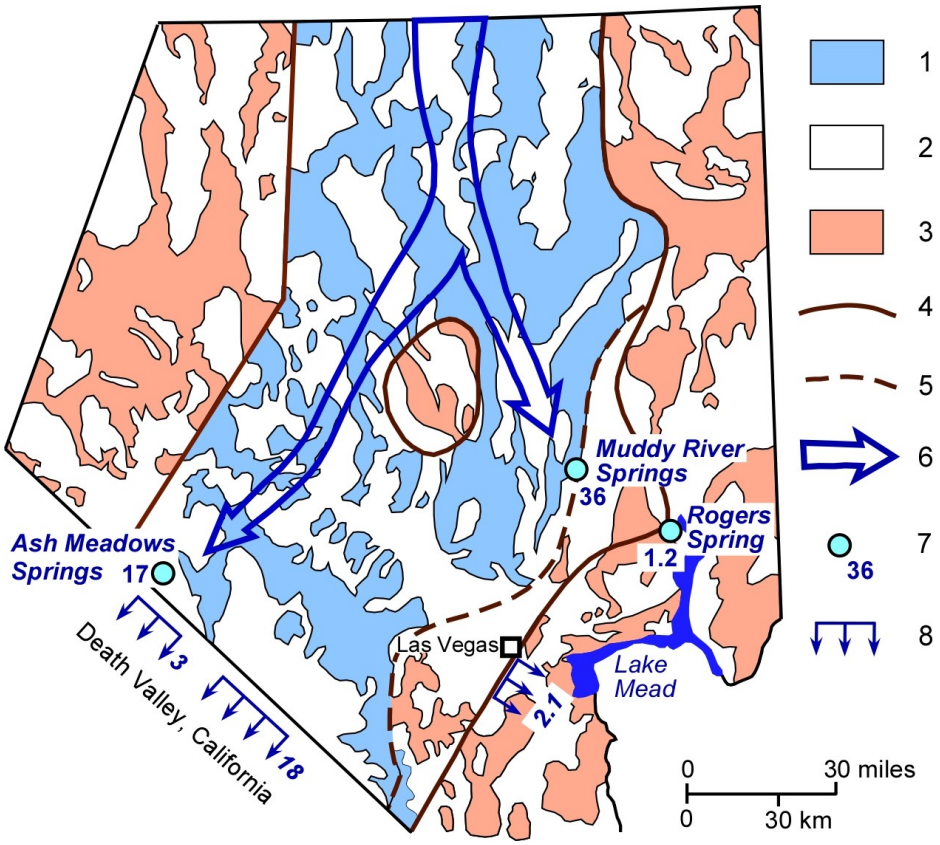


Figure 3.4.22 Components of regional groundwater budget within Central Corridor in southern Nevada. (1) Carbonate rock outcrops within Central Corridor; (2) Basin fill; (3) Areas underlain by thin or isolated carbonate rocks, or carbonate rocks that may contain poor-quality water; (4) Boundary of Central Corridor; (5) Western limit of rock sequences containing salt-bearing minerals and overlying carbonate rocks. Groundwater east of this line is generally poor quality; (6) General direction of regional groundwater flow; (7) Spring with the annual discharge in acre-feet per year or AFY (one AFY equals 0.0014 cfs); (8) Regional groundwater outflow to Death Valley and California, in acre-feet per year. Modified from Dettinger, 1989.

Meadows Springs, Nye County

Ash Meadows is drained by Carson Slough, a through-flowing tributary of the Amargosa River, which terminates in Death Valley. The boundaries of Ash Meadows are poorly defined, but the name is generally applied to the gently sloping terrain, watered by numerous springs, within the southeastern part of the Amargosa Desert, part of the greater Mojave Desert. The western boundary of Ash Meadows corresponds to the western limit of the low salt meadows in the Carson Slough bottomland. Fairbanks Spring defines the northern limit, and Bole Spring is approximately the southern limit. As thus defined, Ash Meadows lies wholly in Nevada (Dudley and Larson, 1976).



Figure 3.4.23 Ash Meadows area from above. Courtesy of Ecoflight, 2023.

Groundwater discharges from the Ash Meadows groundwater flow system at dozens of springs and seeps in the Ash Meadows National Wildlife Refuge. The refuge includes about 21,850 acres of desert uplands and spring-fed oases. The springs and seeps in the Refuge are an area of major discharge of groundwater that flows beneath the eastern part of the Nevada Test Site toward Death Valley in California (Nichols et al., 1997; USGS).

The highly transmissive lower carbonate aquifer is widely distributed beneath the ranges and basins lying to the northeast of Ash Meadows. Recharge to the aquifer moves through fractures and fault zones, which form a conduit system so permeable that it acts as a gigantic drain for about 4,500 mi^2 (Winograd and Thordarson, 1975). The flow system encompassed by this area discharges at the numerous springs in Ash Meadows and is therefore known as the Ash Meadows groundwater system. Walker and Eakin (1963) estimated the natural spring discharge of the system to be about 17,000 acre-ft per year (about 23.5 cfs).

Between the altitudes of 2,325 ft and 2,175 ft, in the 1970s, dozens of springs and small seeps discharged a total of about 11,000 gal/min (24.5 cfs). About 3,000 gal/min (6.7 cfs) of that was from Crystal Spring alone (Dudley and Larson, 1976).

Ash Meadows National Wildlife Refuge can be reached from Las Vegas by taking US-95 North to NV-373 South at Lathrop Wells/Amargosa Valley and continuing on NV-373 South to Death Valley Junction and turning on State Line/Belle Vista Road to Spring Meadows Road (dirt road) to the South Entrance.

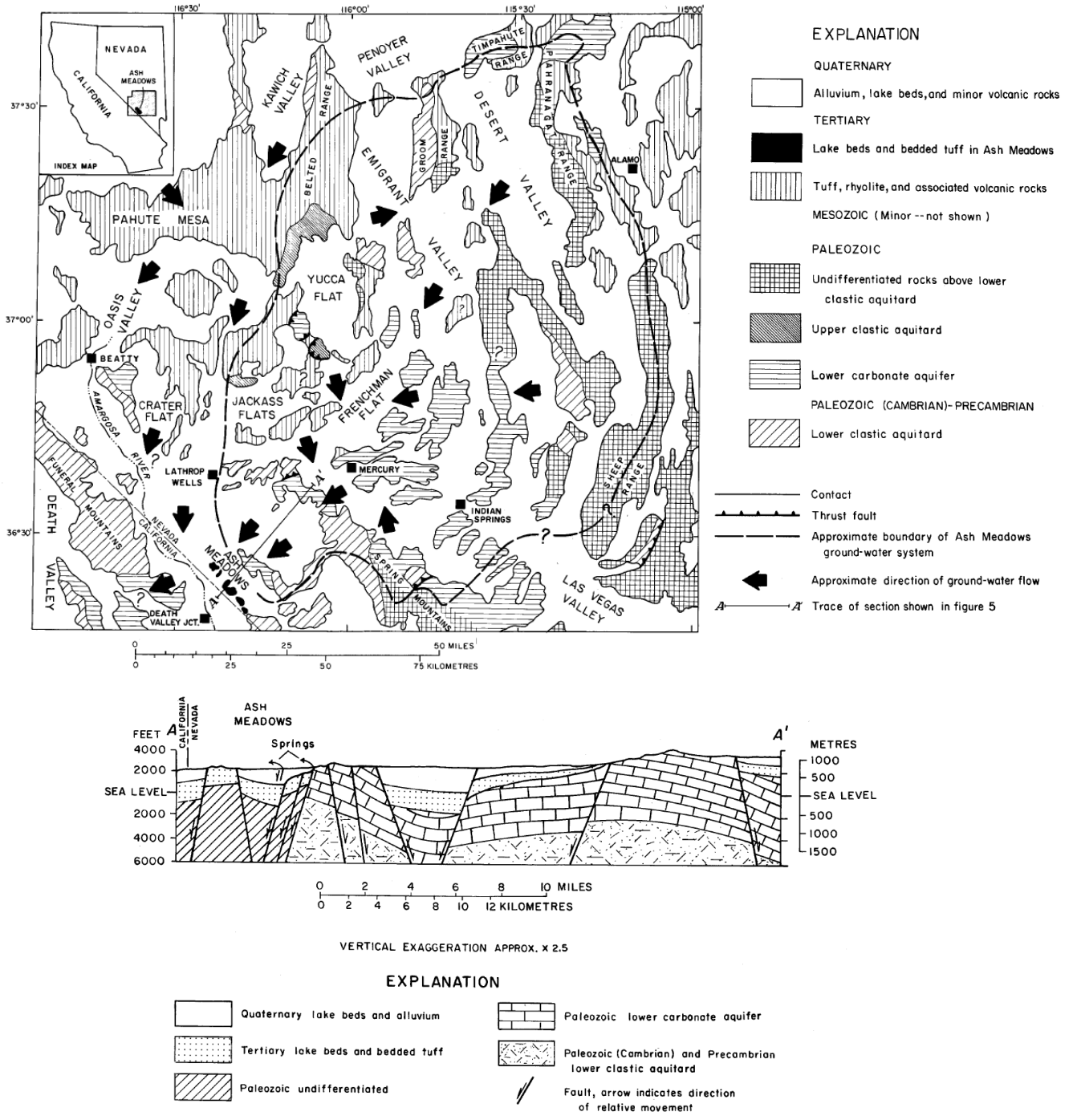


Figure 3.4.24 *Top*: Generalized hydrogeology and boundary of the Ash Meadows groundwater system. Adapted from Carlson and Willden, 1968; Denny and Drewes, 1965; and Winograd and Thordarson, 1975. *Bottom*: Generalized geologic section along flow path approaching Ash Meadows. Modified from Winograd and Thordarson, 1975. From Dudley and Larson, 1976.



Figure 3.4.25 Ash Meadows NWR Crystal Spring Boardwalk leads to a beautiful Caribbean-blue spring pool. This spring produces 2,800 gallons of water per minute (6.25 cfs), is approximately 15 feet deep, and the water stays a constant 87°F 30°C. The length of the boardwalk is approximately 0.9 miles round-trip and there are benches, viewing area complete with scopes and colorful informational panels along the way. The boardwalk is wheelchair accessible, but no bicycles are allowed. Photo by Rod Colvin, U.S. Fish & Wildlife Service.



Figure 3.4.26 Ash Meadows wetland. Photo by Alan Nyiri, U.S. Fish & Wildlife Service (USFWS).

As written by Adamas Weitzenfeld and Jennifer Heroux in a story for the USFWS entitled “Galapagos in Mojave Desert”: *Despite being located in the hottest and driest corner of the United States, Ash Meadows National Wildlife Refuge is home to the second greatest concentration of endemic species in North America. In just under 24,000 acres live 26 species found nowhere else on Earth. That's because approximately 10,000 years ago, as the climate warmed, aquatic species survived by seeking refuge in receding and fragmenting wetlands. Isolated for millennia, the survivors evolved into new species in separated "islands of water" scattered across a sea of desert. Nearly all of Ash Meadows' endemic species live in or around its 50 springs and seeps.*

The story of this wetland’s conservation begins thousands of years before the United States existed. Countless generations of the Nuwu and Newe’s ancestors seasonally cared for the land known as Koieechee and Qwaiuhwuzzah. In the 19th century, Anglo-American settlers arrived and re-named and re-imagined the land as Ash Meadows. Nonetheless, the Nuwu and Newe continued to live and gather at Ash Meadows until the mid-1900s when large agricultural endeavors displaced them. Driven more by profit than stewardship, the landowners drained and mined the marshes, bulldozed and plowed mesquite meadows, straightened and piped spring outflows, and introduced bullfrogs and bass, thus imperiling endemic species and the original caretakers’ relationship to the land.



Figure 3.4.27 Endanegerd and Federally protected Devil's Hole pupfish, one of the world's rarest fish. Photo by O. Feuebacher, U.S. Fish & Wildlife Service.

*Though a 1976 U.S. Supreme Court ruled in favor of federal water rights to protect the Devils Hole pupfish from dropping water levels (*caused by groundwater pumping for irrigation; author's note*), this landmark decision did not fully protect Ash Meadows' other endemic species. Soon after, a land development company purchased the land from the farmers to build a 50,000-person resort community. The proposed community would have paved over rare habitats and consumed three times the annual discharge of all 50 springs. In 1984, the efforts of conservationists paid off again, and the Nature Conservancy successfully acquired and transferred the land to the care of the U.S. Fish and Wildlife Service. On June 18 it became Ash Meadows NWR.*
https://www.fws.gov/story/ash-meadows-galapagos-mojave-desert





Figure 3.4.28 *Left*: Kings Pool is one of many freshwater springs at Ash Meadows National Wildlife Refuge (NWR). Visitors can view the spring by walking along the boardwalk at the Point of Rocks trail. Photo by USFWS. *Right*: Big Spring at Ash Meadows NWR. Photo by Kate Bloomfield, USFWS.

Ash Meadows National Wildlife Refuge was named after the galleries of ash trees described in expedition notes from 1893. This desert oasis, a very rare and unique ecosystem, is recovering and playing an important role in global conservation efforts. The refuge strives to promote conservation management and awareness through environmental education, outreach programs, volunteerism, and visitor services programs (https://www.fws.gov/refuge/ash-meadows)



Figure 3.4.29 Crystal Spring, Ash Meadows NWR. Courtesy of Amargosa Conservancy.

AMARGOSA CONSERVANCY
THE FIGHT TO SAVE ASH MEADOWS, DEATH VALLEY, AND THE AMARGOSA VALLEY
https://www.amargosaconservancy.org/news-blog/the-fight-to-save-ash-meadows/

Included here are several excerpts from a desperate call for action (summer of 2023):

Canadian mining company Rover Critical Minerals (then named Rover Metals) in summer of 2023 submitted an application to the Bureau of Land Management (BLM) seeking to conduct a lithium exploration project that included drilling up to 30 boreholes to a depth of 300 feet on the northern boundary of the Refuge.

The looming prospects of exploratory drilling and open-pit mining illuminated the reality that despite 40 years of protected status and management under the U.S. Fish and Wildlife Service, Ash Meadows remains highly vulnerable to impacts from beyond its designated borders. And beyond Ash Meadows, the rural communities of this corner of the Mojave desert also came to the realization that their lives and livelihoods revolved around the sustainable conservation of groundwater. Consensus quickly formed around the notion that *left unchecked, mineral exploration and industrial mining on unprotected public lands posed a potentially existential threat to communities whose lifeblood is the flow of groundwater ferried by the Amargosa River.*

A coalition including the Timbisha Shoshone Tribe, 24 nonprofit organizations, local governments, and concerned citizens led the fight against this project over the summer and raised the alarm. As a result of public pressure and litigation filed by the Amargosa Conservancy and the Center for Biological Diversity, the BLM rescinded their approval of Rover Metals' initial project application on July 19th, 2023. The BLM also required the company to conduct a full National Environmental Policy Act review and to submit a full Plan of Operations.

The petition championed by the Amargosa Conservancy successfully prompted the U.S. Department of the Interior to initiate a 20-year administrative mineral withdrawal process, temporarily freezing new mining claims across 308,890 acres of public land surrounding Ash Meadows. This grassroots momentum also led the Amargosa Conservancy and local communities to pursue permanent legislative protections, such as designating an Ash Meadows National Conservation Area.

Muddy River Springs, Clark County

The Muddy River Springs, also commonly referred to as Warm Springs, are at the northwestern end of Moapa Valley, and include a network of springs and seeps that form the headwaters of the Muddy River (Fig. 3.4.30). Over time, this area has been home to the Anasazi and the Southern Paiute Indian tribes, outlaws, prospectors, Mormon settlers, and others (Beck et al., 2006; Baxter and Haworth, 1996; U.S. Fish and Wildlife, 1991). The availability of water provided past and current inhabitants with the resources necessary to sustain life in the desert. In addition, nearby entities, such as the Moapa River Indian Reservation, the Moapa Valley Water District (MVWD), and the Nevada Power Company (NPC), use water obtained from surface streams fed by the springs (Beck et al., 2006).

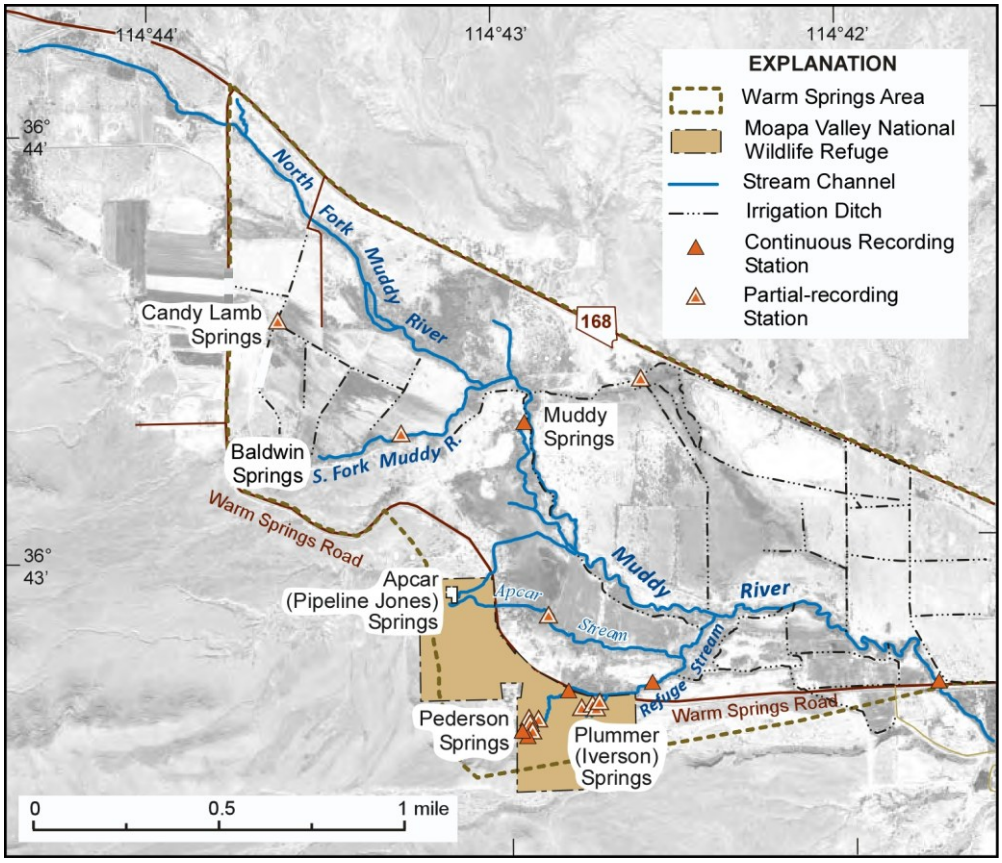


Figure 3.4.30 Location of major spring groups, and U.S. Geological Survey monitoring stations on August 17, 2004, in the Warm Springs area near Moapa, Nevada. Modified from Beck et al., 2006.

The Warm Springs area is home to the Moapa Valley National Wildlife Refuge. The refuge was established in 1979 solely to protect and secure the riparian habitat of an endangered native minnow, the Moapa dace (Moapa coriacea; Fig. 3.4.31). It was the first National Wildlife Refuge created for the purpose of protecting an endangered fish species.

The Moapa Valley NWR is in the Warm Springs area of the upper Moapa Valley in northeastern Clark County in southern Nevada. It lies just south of State Highway 168 and the Muddy River, between Interstate 15 on the east and U.S. Highway 93 on the west. The refuge is bounded on the North by Warm Springs Road, on the south by Battleship Wash, and on the east and west by private property. It is approximately 60 miles northeast of the city of Las Vegas and 9 miles west of the town of Glendale.

Dace habitat on the refuge consists of stream channels supported by six springs, three of which are within refuge boundaries (Apcar, Pederson, and Plummer Springs; see Fig. 3.4.30).



Figure 3.4.31 Moapa Dace (front center) and Springfish. Photo by Mark Hereford, USGS. The Moapa dace is one of six endemic species at Warm Springs. These endemic species are native and can only be found in this region of Nevada and nowhere else on earth. Of all the sensitive species at Warm Springs, the Moapa dace is the most imperiled. Listed as an endangered species in 1967 under the Federal Endangered Species Preservation Act, the preservation and population recovery of the Moapa dace has received the highest priority in the management of Warm Springs, and the U.S. Fish and Wildlife Service considers the recovery of the Moapa dace to be essential in this region.

Stable flows from the refuge's springs (see Fig. 3.4.32) fill meandering channels downstream that provide ideal habitat for dace, Virgin River chub and other species of endemic fish and invertebrates. The spring bank and riparian plant communities provide habitat for southwestern willow flycatcher as well as a rich diversity of migratory and resident songbirds, colonial nesting species, and other native wildlife. The springs are thermal in nature and have an average annual water temperature of 90 degrees Fahrenheit at their points of discharge.

Before the refuge was established, dace populations were in peril due to habitat destruction and modification. Competition with introduced species such as mosquitofish, shortfin molly, and tilapia also contributed to the dace's decline. However, as more habitats are restored (Fig. 3.4.33) and non-native species are removed, the fish has begun to rebound. Population surveys show an increase in numbers in recent years. As of February 2024, there were 1,935 dace in the Muddy River system. Biologists say that there needs to be a stable population of at least 6000 to ensure the security of the species (https://www.fws.gov/refuge/moapa-valley).

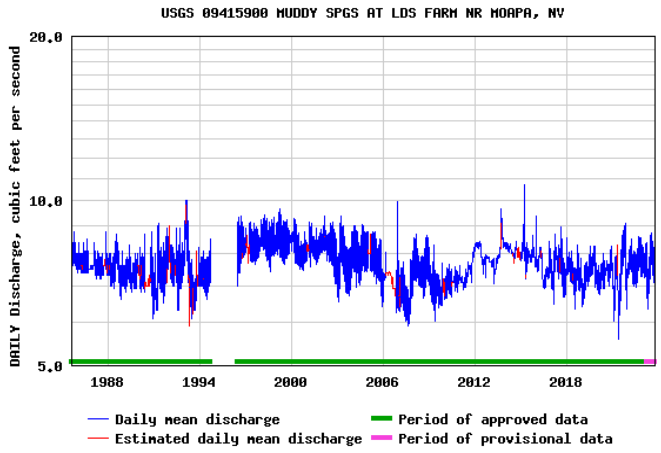


Figure 3.4.32 Discharge rate, in cfs, of Muddy Springs at gaging station Church of the Latter-Day Saints (LDS) farm near Moapa, Nevada.

Warm Springs are near the southern boundary of the White River groundwater flow system which developed in carbonate rocks of Paleozoic age and sedimentary rocks of Tertiary age. Carbonate rocks form complex aquifers whose extents and thicknesses are largely unknown (Prudic et al., 1995). Discharge from the springs and seeps of the Warm Springs area is believed to be the largest and most southerly outflow from this groundwater system. Recharge is primarily from precipitation in the high mountain ranges in east-central Nevada (Eakin, 1966).



Figure 3.4.33 *Left*: Restored Pederson Spring. *Right*: Confluence of three spring runs coming from Plummer Springs. Photographed June 9-10, 2004, by D. Beck. From Beck et al., 2006.

The fascinating history of the Muddy River Springs water resources development is described by Beck et al. (2006, Appendix A; see also Figs. 3.4.34 and 3.4.35). Below are a few interesting points.

The first known dwellers in the area were the Anasazi and the Southern Paiute Indians (Baxter and Haworth, 1996). Little is known about the Anasazi in southern Nevada. By the early 1800s, the Paiutes had sizable populations along the Muddy and Virgin Rivers. Although most of these riverine and desert groups were primarily foragers and hunters, the Paiutes were known to have irrigated crops, such as corn, squash, melons, and wheat, along the banks of the Muddy River (Mozejko, 1981). In addition to using the river as a source of irrigation and drinking supply, the Paiutes also used it for ceremonial rites, which are still practiced today within the Moapa Paiute Indian Reservation (Phil Swain, Moapa Band of Paiutes, oral communication, 2004).

The first known ranch in the Warm Springs area was started in 1871 with several cattle and horses rustled by a fugitive bank robber from Texas. He built a rock house just north of the current Church of the Latter-Day Saints (L.D.S.) Recreation Area (Baxter and Haworth, 1996). He called the ranch Stone Cabin Springs and worked it until his death in 1882.

Past and present owners of large tracts of land and various springs include Francis Taylor, who established Warm Springs Ranch. Billionaire Howard Hughes, after seeing the Warm Springs area during a test flight, purchased the entire Warm Springs Ranch. Although he owned the Ranch for nearly 8 years, Hughes reportedly never set foot on it; (Baxter and Haworth, J., 1996). The Church of the Latter-Day Saints (L.D.S.), Desert Oasis Warm Springs Resort, Frederick Apcar, the Federal Government (the land is managed by the U.S. Fish & Wildlife Service), Clarvid Lewis, were all owners along with other private ranchers and farmers.

Farming and ranching continue today within the Warm Springs area, but land usage has dropped significantly since its peak in the mid-1970s.

In December 1997, the Nature Conservancy and U.S. Fish and Wildlife Service signed an agreement for cooperative management of the Moapa Valley National Wildlife Refuge. The endangered Moapa Dace have increased in Refuge streams as UFWS and Conservancy staff and volunteers have restored stream habitat by removing excess vegetation, digging out years of accumulated pea gravel, and restoring natural pools, riffles, and the natural substrate. The Conservancy has played a key role in developing cooperation between public agencies and private landowners to control the invasion of tilapia, an exotic fish, which has appeared as an emergency threat to dace and other native fish in the Muddy River.



Figure 3.4.34 View of the Francis Taylor Mansion built in the 1950s in the Warm Springs area. The pool is fed by a spring. Photographed in July 2004 by D. Beck. From Beck et al., 2006.



Figure 3.4.35 Swimming pool fed by a spring in the recreation area of the Latter-day Saints Church property in Warm Springs near Moapa. Photographed in June 2004 by D. Beck. From Beck et al., 2006.

For years, Warm Springs , together with the precious Moapa dace and all other endemic species that live nowhere else on Earth, were threatened by the proposed Coyote Springs development upgradient of the springs. The development is envisioned as a man-made oasis, entirely relying on pumping of groundwater for nearly 160,000 homes, a golf course, and all of the other excesses that go with it, all on more than 40,000 acres in the middle of the desert. Below are excerpts from a press release of the Center for Biological Diversity, illustrating the (temporary?) victory for those that were opposed to the development.

Center for Biological Diversity
For Immediate Release, January 25, 2024
Nevada Supreme Court Affirms Protecting Water for Wildlife

CARSON CITY, *Nev.*— The Nevada Supreme Court ruled today that the state has a right to manage groundwater for the preservation of senior water rights and the public interest, including wildlife.

Today's decision in the Lower White River Flow System case will help determine the future of water management in the driest state in the union. The Center for Biological Diversity was a co-respondent in the case.

"This is a monumental victory for the conservation of water resources in a time of escalating climate change and drought," said Patrick Donnelly, Great Basin director at the Center. "Nevada's animals are already under stress from the lack of water resources, and our victory will help ensure that industry and developers don't leave wildlife high and dry."

The case centers around an aquifer that sustains the Muddy River in Clark County, Nevada. This spring-fed oasis provides habitat for an endangered fish called the Moapa dace. The Muddy River is also a source of drinking water for Las Vegas.

Coyote Springs, a proposed city of a quarter-million people in the desert 50 miles northeast of Las Vegas, applied for groundwater rights to pump water that scientists say would deplete the springs the Moapa dace relies on for survival.

The state engineer, Nevada's water czar, ordered a pump test and extensive hydrologic investigations. He then ruled that there was a finite supply of water available in the aquifer and that excessive pumping would impair senior water rights and harm the Moapa dace.

In its ruling today, the court overturned a district court ruling in *Sullivan et al. v. Lincoln County et al.*

The court also remanded the case back to the District Court Judge Bita Yeager for a ruling on whether the state's order was based on substantial evidence. This is a distinct set of legal criteria from today's ruling, which evaluated whether the state had the power to issue the order.

"We've been fighting for this little fish for more than 15 years, and we're not backing down an inch," said Donnelly. "We'll continue to press our case and we won't rest until we can pound the nails into Coyote Springs' coffin."

Breaking down $2B Coyote Springs water fight in Nevada
By Latoya Silmon, Fri, July 10, 2026 at 1:39 PM
https://news3lv.com/features/breaking-down-the-law/breaking-down-2b-coyote-springs-water-fight-in-nevada

LAS VEGAS (KSNV) — A trial that could cost Nevada taxpayers up to $2 billion begins this week in Las Vegas.

The plaintiffs are the developers of Coyote Springs — a wouldbe masterplanned city 50 miles north of Las Vegas, meant to rival Summerlin, that today has one golf course and zero homes.

The developers say Nevada's water regulators effectively took their property rights without paying for them.

The state says nobody "took" anything — the water simply is not there.

Attorney Matthew Hoffman with Battle Born Injury Lawyers joined us to break down the legal issues at play and why taxpayers could be on the hook.

And the fight to save one of the most unique desert wildlife habitats in the world continues…

Rogers Spring, Clark County

Rogers Spring is in Lake Mead National Recreation Area and is operated by the National Park Service. It is situated right beside Northshore Road (Hwy 167), 51 miles north of the visitor center and about one hour drive from Las Vegas.

As described by NPS (2024; https://www.nps.gov/lake/learn/nature/rogers-spring.htm), Rogers Spring and other smaller springs such as Corral Spring and Blue Point Spring in the "North Shore Complex" comprise one of the terminal discharge areas for the regional carbonate-rock aquifer system of eastern Nevada and western Utah.

Groundwater discharge from these springs along the Rogers Spring Fault totals approximately 1,000 gallons per minute, with temperature ranging between 82 and 90°F. Due to the importance of the springs from both an ecological and recreational standpoint, the National Park Service continues to monitor them including potential adverse effects associated with nearby, large-scale ground water pumping from the regional carbonate-rock aquifer (see Fig. 3.4.36).

The prevailing theory suggests that much of the recharge water that enters the carbonate-rock aquifer occurs in the high mountain ranges around Ely, Nevada, located 250 miles north of Lake Mead. As this groundwater flows south through the carbonate rocks, it encounters several faults along the way, including the Rogers Spring Fault, which has caused the older carbonate rocks (primarily limestone and dolomite) to be displaced against younger evaporite deposits of the Muddy Creek and Horse Spring formations. Here, the lower permeability of these evaporite deposits, along with high subsurface water pressure, forces the ground water in the carbonate rocks to flow upward along the fault and emerge at the surface as Rogers Spring.

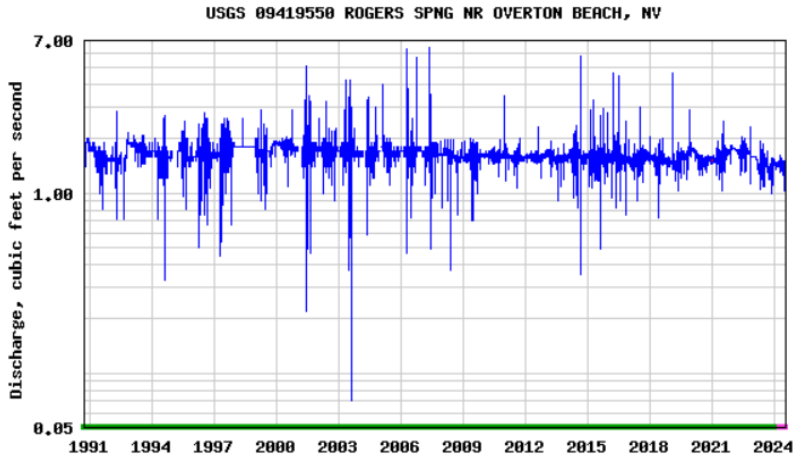


Figure 3.4.36 Discharge rate hydrograph of Rogers Springs. Average is 1.8-1.9 cfs, with a slightly declining trend since late 1990s, likely due to large-scale pumping in the area.



Figure 3.4.37 View of Rogers Springs with Lake Mead in the background. NPS Photo by Andrew Cattoir.

3.5 Arizona, and Colorado

Grand Canyon Springs, Coconino County, Arizona

The Grand Canyon measures over 270 miles long, up to 18 miles wide and a mile deep making it one of the largest and deepest canyons in the world. The Grand Canyon tells one of the world's greatest geologic stories. Its distinctive features allow researchers to piece together the history of this unique location, one of America's treasures and a UNESCO World Heritage Site. Thinking of the geologic record as a book is helpful to understand each page of Earth's history, the story of which is provided by the USGS on its webpage https://www.usgs.gov/geology-and-ecology-of-national-parks/geology-grand-canyon-national-park.

The beginning of the story starts at the bottom of the canyon and moves forward in time as one gets closer to the rim. The three main rock layer sets in the Grand Canyon are grouped based on position and common composition into (1) Metamorphic basement rocks, (2) The Precambrian Grand Canyon Supergroup, and (3) Paleozoic strata. These three main sets of rocks were first described by the explorer and geologist John Wesley Powell during his expeditions of the Grand Canyon in the late 1860s and early 1870s.

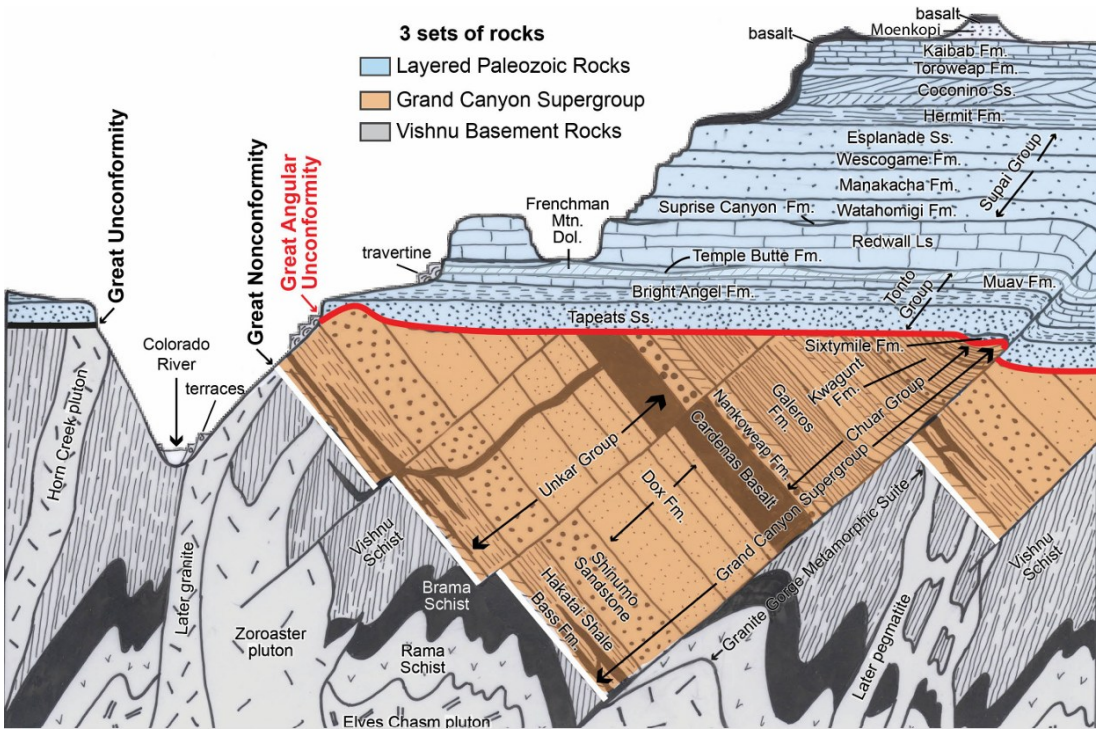


Figure 3.5.1 Stratigraphic column of rocks of the Grand Canyon region showing the three sets of rocks and major unconformities: Great Nonconformity (white line), Great Angular Unconformity (red line) and Great Unconformity (black line). Fm = Formation; Ss = Sandstone; Ls = Limestone. From Karlstrom et al., 2021.

The oldest rocks in the Grand Canyon, found at the bottom of the canyon, are primarily metamorphic, with igneous intrusions collectively called Zoroaster granite. The name given to this combination of metamorphic and igneous rocks is Vishnu Basement Rocks. Primarily schist (metamorphic) with granite (igneous), these rocks have visible crystals and are about 1.7 billion years old (Proterozoic era). On Powell's expedition to explore and map the Grand Canyon, he named this part of the exposed rock "The Granite Gorge."

The middle rock set, the Grand Canyon Supergroup, is primarily sandstone and mudstone, both sedimentary rocks, with some areas of igneous rock. They are from the late Proterozoic, only slightly younger than the metamorphic basement rocks. These rocks do not contain many fossils, because they formed before complex life on Earth was common. The few fossils that are present include stromatolites, columns of sediment formed by cyanobacteria. The composition (sandstone) and presence of stromatolites indicate that this area was previously a very shallow sea. The rock layers in the Grand Canyon Supergroup have been tilted, whereas the other rocks above this set are horizontal. This is known as angular unconformity. The top of these sediment layers was then eroded away, forming the Great Unconformity.

Above the Great Unconformity are Paleozoic sedimentary layers, primarily sandstone and limestone. This set makes up most of the typical reddish and grayish-whiteish layers one often sees in images, and which come to mind when thinking of the Grand Canyon. The most common fossils are small sea creatures, such as brachiopods, bryozoans, coral, and crinoids, which indicates that the region was a warm, shallow sea when these sediments were deposited.



Figure 3.5.2 *Upper Left*: Kaibab limestone (white rock in the forefront) tops the Paleozoic sedimentary sequence at the Grand Canyon's South Rim. *Upper Right*: Aerial view south toward a sinkhole in Kaibab Formation, west side of upper Mohawk Canyon drainage. Photo from Grand Canyon collection of Billingsley et al., 2019; USGS. *Lower Left*: Fractured Coconino sandstone, with typical crossbedding, as seen from Bright Angel trail below South Rim. *Lower Right*: Entrance to a cave in the Redwall Limestone. Note human figures for scale. NPS photograph by Dale Pate. From Graham, 2020.

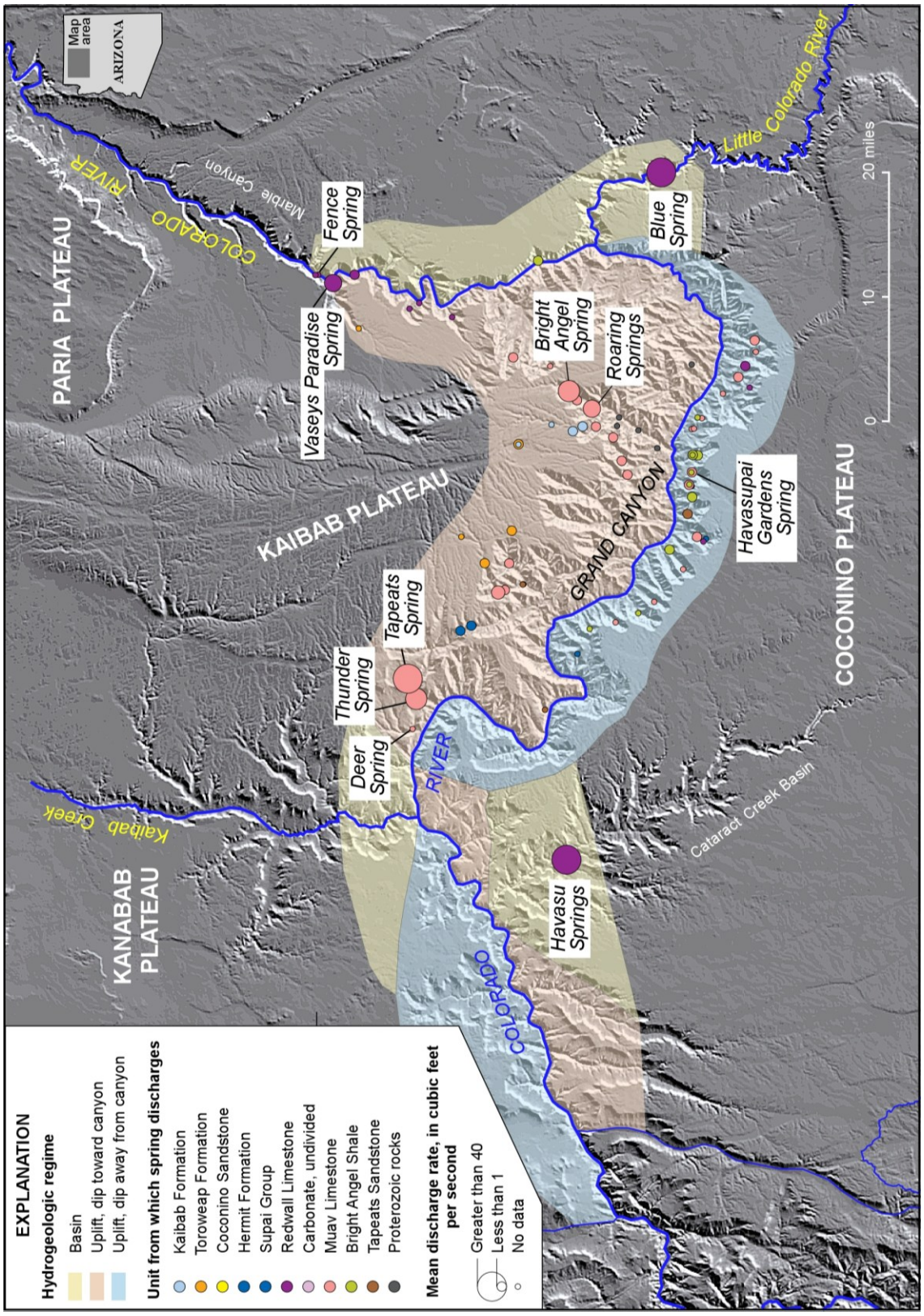


Figure 3.5.3 Map of the hydrogeologic regimes of springs discharging to the Grand Canyon and their designated geologic units. Symbol size indicates relative magnitude of mean discharge rate. Modified from Knight and Huntoon, 2022.

As explained by Bills et al., 2007; and Graham, 2020, the springs in Grand Canyon result from the complex relationship between caves and karst in the canyon. Smaller springs emerge from C-aquifer which is defined as the saturated parts of the Kaibab Formation, Toroweap Formation, and the Coconino Sandstone, all of Permian age (Cooley et al., 1969). The C-aquifer is a water-table aquifer for most of its occurrence with depths to water that range from a few hundred feet to more than 1,500 feet.

Generally, the inter-crystalline permeabilities associated with the carbonate rocks that form most of the Kaibab and Toroweap Formations are negligible, but secondary permeability in the form of dissolution-widened joints and fractures coupled with a well-developed epikarst allows for recharge. The interstitial permeability found in the underlying Coconino Sandstone, although modest, is the largest found in any of the clastic units in the Grand Canyon Paleozoic section (McKee and Resser, 1945) and accounts for most of the storage that is present in the C-aquifer. The result is that perched saturated zones develop above the confining strata of the underlying aquitard composed of Hermit Formation and Supai Group. Groundwater then flows laterally to seeps at the base of the Coconino Sandstone along the canyon or to extensional fault zones and fractures in the aquitard that allow the water to descend vertically and recharge the underlying R-aquifer (Knight and Huntoon, 2022).

The larger springs emerge from the R-aquifer, which is composed of Redwall and Muav limestones. The Redwall-Muav aquifer is a confined aquifer for most of its occurrence with hydraulic heads of several hundred to more than 500 feet above the top of the aquifer in the western part of the Coconino Plateau and more than 2,000 feet above the top of the aquifer in the area near Flagstaff south of the Grand Canyon.



Figure 3.5.4 The Royal Arches are a seep and large alcove in Redwall Limestone on River Mile 41.5 in Marble Canyon, Grand Canyon National Park. Large alcoves of this kind are often formed in arid climates by salt wedging. Salt wedging is an erosional process where saline groundwater emerges, (in this case on a cliff face), and undergoes evaporation, leaving salt crystals that grow in cracks and pores and spall out pieces of rock. Photo and text courtesy of David Kreamer.

Blue Spring

Blue Spring is on Navajo Tribal Land and hiking or camping is not allowed without prior permission from the Navajo Nation. The Blue Spring groundwater system is a basin-type karst system. It is the largest groundwater system that drains to the Grand Canyon, encompassing 27,000 square miles underlying the Little Colorado River drainage basin in Arizona and western New Mexico (Knight and Huntoon, 2022).

Blue Spring is a collective description of at least 36 individual springs on the floor of the Little Colorado River canyon within about 13 river miles of its confluence with the Colorado River. According to Cooley (1976), base flow upstream from Blue Spring is only about 5–7 ft^3/s. Blue Spring discharges at a constant rate of about 90 ft^3/s, which is almost half of the perennial flow at the mouth of the Little Colorado River. Cooley (1976) also recorded 60 ft^3/s contribution to streamflow from two unnamed springs just downstream from Blue Spring, 40–45 ft^3/s from other springs before river mile 10, and another 20 ft^3/s before river mile 3.1. In total, the springs of the Blue Spring groundwater system contribute 220 ft^3/s to the Little Colorado River (Cooley, 1976).

The large and steady discharge rate from the springs is consistent with the size of its catchment area. The water emerges at an average temperature of 70 °F and is turquoise in color owing to the spontaneous precipitation of calcium carbonate (Johnson and Sanderson, 1968). Deposits of calcium carbonate create a continuous series of travertine dams and rimstone pools between the springs and the Colorado River.

The Blue Spring is situated within an extensional rift zone riven by a dense set of generally north-south trending small displacements owing to normal faults. The regional dip of the Paleozoic strata is very gentle toward the southeast and a structural low near or upstream from Cameron, Arizona. The result is that the floor of the Little Colorado River cuts ever deeper into the Paleozoic section in the downstream direction along this reach. Most of the water emerges from a few major springs in the Mooney Falls Member of the Redwall Limestone, which discharge from saturated caves and dissolution-widened fractures associated with faulting (Knight and Huntoon, 2022).



Figure 3.5.5 View southwest toward three main springs of Blue Spring on southwest side of Little Colorado River; springs issue from Mooney Falls Member of Redwall Limestone. Photo dated January 26, 1970; from the Grand Canyon collection of Billingsley et al., 2019; USGS.

Cooley (1976) argues that most of the Blue Spring water is derived from the C-aquifer underlying the Little Colorado River Basin. In Cooley's conceptual model, the water circulates laterally through the C-aquifer downstream within the basin until it is intercepted by the north-trending extensional faults oriented subparallel to the East Kaibab Monocline west of Cameron. It circulates downward within the fault zone to the R-aquifer where it then follows dissolution-widened fractures and caves in the Redwall Limestone that developed along the faults, emerging at the downstream springs.

Groundwater circulation rates through the fault zone that supplies the Blue Spring groundwater system are likely to be exceptionally fast given the expected karstic permeabilities. However, those rates have not been quantified.



Figure 3.5.6 View southeast toward Blue Spring jetting out of Moony Falls Member of Redwall Limestone on southwest side of Little Colorado River. Photo dated May 2, 1970; from the Grand Canyon collection of Billingsley et al., 2019; USGS.



Figure 3.5.7 Confluence of Little Colorado River (turquoise color) and Colorado River. Screenshot from a video *Virtual Tour—Grand Canyon, AZ Dams Threaten the Little Colorado River*, produced by EcoFlight in cooperation with the Grand Canyon Trust.

https://www.grandcanyontrust.org/little-colorado-river-dam-proposals

The Threat

The Little Colorado River, which joins the mainstem Colorado River inside Grand Canyon National Park, is the wrong place to build dams (https://www.grandcanyontrust.org/little-colorado-river-dam-proposals)

This is the title of an article on the web site of the Grand Canyon Trust in which one can read:

After years of advocacy by Native communities, tribes, the Grand Canyon Trust, and others, in 2022, the federal government cancelled preliminary permits it had issued to a Phoenix-based company for two proposed hydroelectric projects on Navajo Nation lands mere miles from the national park boundary. Both would have flooded a pristine section of the Little Colorado River Gorge, muddied its distinctive turquoise-blue waters, destroyed areas sacred to Native peoples, and threatened the habitat of the already endangered humpback chub.

However, a third application to dam Big Canyon, a nearby tributary canyon adjacent to the Little Colorado River, is still pending.

Public outcry from Grand Canyon advocates defeated the proposed Marble Canyon Dam in the 1960s. It's time we raise our voices again to stop dams from altering the Little Colorado River forever.

What is at stake is flooding and destruction of the most sacred places of the Navajo Nation and many other Native American Tribes, including the largest spring in the Grand Canyon Area, Blue Spring, which contributes most of the flow of the Little Colorado River and creates the most amazing sight one can imagine (photographs in Figs. 3.5.7 and 3.5.8 can hardly do the justice). One of the proposed projects would have flooded a Hopi sacred site, a place where the Hopi people believe they emerged into this world.

In April 2024, Federal Energy Regulatory Commission (FERC) denied the proposed Big Canyon Dam, citing the Navajo Nation's opposition.



Figure 3.5.8 Confluence of Little and Colorado Rivers. Courtesy of Grand Canyon Trust. As groundwater travels through deep karst aquifer, the highly acidic, carbon dioxide-rich water dissolves the limestone, filling the water with high concentrations of calcium, magnesium, and bicarbonate ions. When this pressurized groundwater emerges at the springs, it experiences a drop in pressure as it meets the open air. The water rapidly loses its carbon dioxide gas to the atmosphere (off-gassing). This shifts the chemical equilibrium and causes the dissolved minerals to transform into tiny, solid crystals of calcium and magnesium carbonate that remain suspended in the water and reflect the blue light. The Little Colorado River's blue waters have been of deep cultural significance to Native peoples since time immemorial. They have traditional connections to what Zuni people refer to as the umbilical cord of the world: "The Little Colorado River is like an umbilical cord that connects us back to our emergence place, the Grand Canyon."

The Little Colorado River, an Umbilical Cord to the Grand Canyon | Grand Canyon Trust

Havasu Spring

The Havasu Spring is on the Havasupai tribe reservation, surrounded by the Grand Canyon National Park. In order to maintain the pristine beauty of this isolated desert paradise, the Havasupai tribe limits the number of visitors allowed to visit the reservation. There is no day hiking permitted in the canyon. Visitors must have a reservation, and entrance fees are now paid in advance (see https://theofficialhavasupaitribe.com/).

As explained by Knight and Huntoon (2022), Havasu Spring is the primary point of discharge for groundwater from the Cataract groundwater system which underlies the Coconino Plateau (part of the greater Colorado Plateau) south of the Colorado River.

The R-aquifer within most of the Cataract groundwater system is a classic basin-type karst system with a main discharge point at Havasu Springs on the floor of Cataract Canyon. The rocks that compose the R-aquifer, the Redwall Limestone and the Muav Limestone, are fully saturated under the lower reaches of Havasu Canyon. In contrast, the rocks are mostly dewatered where they are exposed along the South Rim of the Grand Canyon high above the Colorado River and dipping outward, away from the river and canyon wall. The narrow band along the South Rim to the east and west of Cataract Canyon is better described as an uplifted, outward-dipping karst system.

Although not exposed, the karstic character of the flow system supplying Havasu Spring can be readily inferred. The volume of water flowing from the spring reveals that it is fed by a cavern network dissolved in the Redwall Limestone that is analogous to the dewatered two-dimensional planimetric cavern networks exposed within the interior of the Eminence graben in the walls of Marble Canyon (Knight and Huntoon, 2022).

A major flood in 1993 coursed down the otherwise ephemeral Cataract Creek with losses of water into sinkholes that opened along fissures in the fault system 40 miles upstream from Havasu Springs (Melis et al., 1996). Within 2 years, the water quality at the spring freshened (concentration of total dissolved solids decreased), noticeably revealing that the flood waters had diluted the water in the R-aquifer and arrived. It is reasonable to infer that preferential dissolution enhancement of permeability within the R-aquifer aligned with the regional hydraulic gradient, and accounts for the hydraulic connection between the upstream fault zone and Havasu Springs.



Figure 3.5.9 Aerial view toward Havasu Springs area south of Supai village, Havasu Creek, Cataract Canyon. Esplanade Sandstone forms the rim and platform above Pennsylvanian group of strata. Photograph from the Grand Canyon collection of Billingsley et al., 2019; USGS.

The interior of the Cataract Canyon drainage basin is riven by a network of closely spaced, orthogonally intersecting, small-displacement normal faults that are well exposed in the Kaibab Formation at the surface. Approximately 20 percent of the surface area of the basin is internally drained by sinks along these faults, which may conduct available surface waters down through the C-aquifer directly to the R-aquifer (Melis et al., 1996).

A seepage investigation conducted from Havasu Spring to the mouth of Havasu Creek for the Havasupai Tribe showed that the creek loses about 14 percent of its flow from the spring to Havasu Falls and then gains 29 percent of its flow from Havasu Falls to the mouth (U.S. Geological Survey, unpublished data, 1995; from Bills et al., 2007).

The water from Havasu Spring is rich in magnesium, calcium, and high concentrations of bicarbonate. As it makes its way through Havasu Creek, these ions combine to form tufa and world-famous dams and falls between the spring and the Colorado River. Taken together, the dissolved ions and tufa/travertine, reflect sunlight resulting in a more intense blue-green color (https://azgs.arizona.edu/photo/havasupai-falls). Havasupai literally means people of the blue-green waters.

The temperature of the water at the spring is about 70 °F and the discharge is fairly uniform at about 60 ft^3/s (Johnson and Sanderson, 1968). The water of the springs sustains a pristine natural environment and biota, some of which are endemic.

As discussed by Knight and Huntoon (2022), it is important to determine flow vectors to the R-aquifer, either in the subsurface or overland, that can be expected from contaminant sources including mined ore and waste rock stored on the surface and mine production water. Of particular concern are migration pathways to the swallow holes in the floor of Cataract Creek upstream from Havasu Springs. No quantitative work has been carried out in the form of dye traces to determine flow rates between Havasu Springs and the sinks in the floor of Cataract Creek in the fault zone 40 miles upstream. The transmissive character of the R-aquifer, particularly where it is fractured by highly conductive extensional faults, remains unquantified.



Figure 3.5.10 View southeast and upstream toward Havasu Creek spring flow, lower Cataract Canyon, just upstream of river mile 157.2. Photograph from the Grand Canyon collection of Billingsley et al., 2019; USGS.

The lifeline of the Havasu Tribe is Havasu Springs and the Havasu Creek which they form, including its spectacular waterfalls. Uranium mining in the watershed of the Springs, which are sacred to the Havasu Tribe and other Native Americans, may impair the quality of groundwater, Springs, and surface water, to the point of no return.



Figure 3.5.11 *Left*: Havasu Falls with travertine pools in the foreground, Havasu Canyon, part of the Grand Canyon area. Photo courtesy of David Price. *Right*: The confluence of Havasu Creek with the Colorado River (river mile 157) is a popular place for boaters to stop and admire the striking blue-green water of Havasu Creek. The turquoise color is caused by water with a high mineral content. At the point where the blue creek meets the turbid Colorado river there often appears a definite break. NPS photo by Erin Whittaker, August 15, 2011 (under https://creativecommons.org/licenses/by/2.0/#).

Congress Must Act to Save the Havasupai Tribe from Extinction

This is the title of the December 3, 2021 opinion piece for Newsweek written by the Havasu Tribe Chairwoman Kissoon in which one can read (https://theofficialhavasupaitribe.com/):

We have seen the irreparable damage uranium mining can do. For generations we have been at the forefront, working to permanently protect our homelands from uranium mining, which has disproportionately harmed and sickened Indigenous people across northern Arizona. The Senate has the power to protect not only our Grand Canyon home and our waters, but also the health and safety of our tribe and the people of the Southwest from the life-threatening effects of uranium mining. We urge the Senate to do the right thing and protect the Grand Canyon now and for future generations.

Despite a decades-long fight to prevent uranium mining and all its negative consequences, including lawsuits supported by many non-governmental organizations, the uranium mining at the Pinyon Plain Mine (previously known as the Canyon Mine) obtained the final regulatory approval and started ore extraction in January 2024.

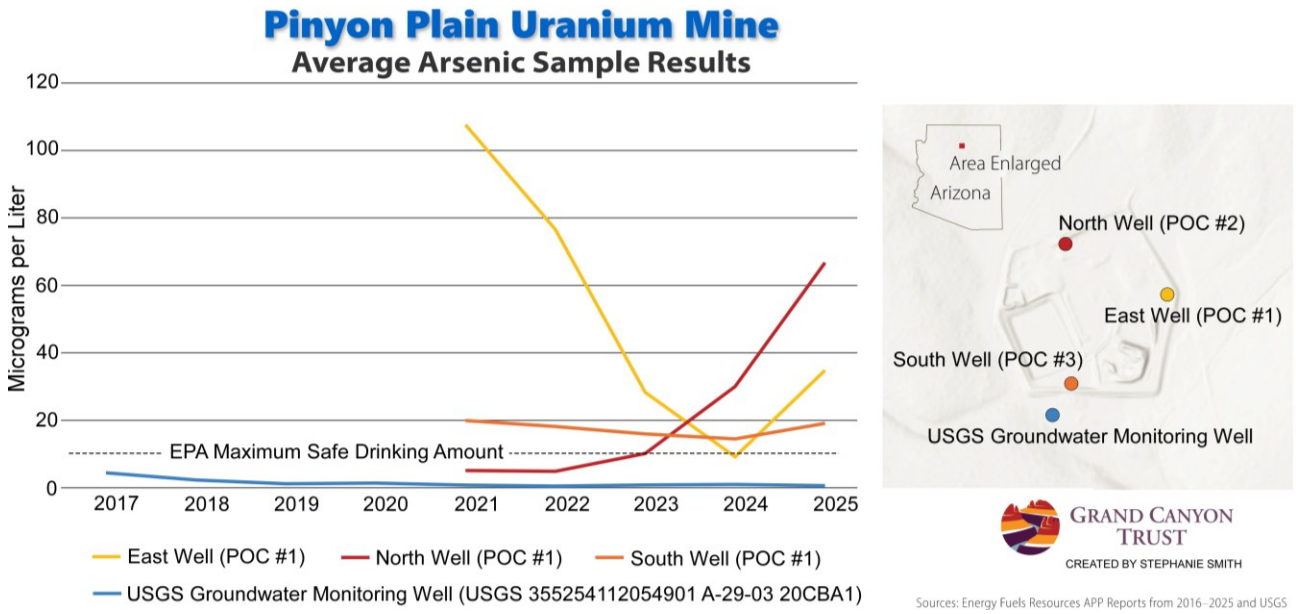


Figure 3.5.12 Graph showing a dramatic recent increase in arsenic concentration in the C aquifer at the Pinyon Plain Uranium Mine. Graphics by Stephaine Smith, https://www.grandcanyontrust.org/resources/arsenic-contamination-at-pinyon-plain-mine/, courtesy of Grand Canyon Trust. Naturally occurring arsenic is released from the host rock by the mining practices, in the process called chemical oxidation. Minerals in undisturbed rocks exist in an oxygen-depleted environment. Excavation brings these minerals into contact with atmospheric oxygen and moisture. When exposed to air and water, the sulfide minerals undergo rapid oxidation. This chemical reaction breaks the molecular bonds holding the arsenic, transforming it into highly mobile and water-soluble forms of inorganic arsenic. The oxidation of sulfide ores generates sulfuric acid. This acidic runoff further speeds up the dissolution and leaching of arsenic into local groundwater and surface water networks (Society for Mining, Metallurgy & Exploration, 2026).

Unfortunately, the Havasu Tribe did not stand a chance against the relentless assault of special interests, and the position of Arizona's regulatory agency in charge (Arizona Department of Environmental Quality, ADEQ). This agency astonishingly disregarded its own decades-long requirements and practices at the existing or potential groundwater contamination cases across the state of Arizona.

For example, in the process of permitting the Pinyon Plain Uranium Mine and replying to numerous public comments, ADEQ stated that "*From a hydrogeologic standpoint, the R-aquifer monitoring well at the Mine is essentially immediately below and downgradient of any potential release from the shaft or workings. The thickness, age difference, and hydraulic separation of the C-aquifer and R-aquifer, at the Mine, make the R-aquifer monitoring well an appropriate monitoring point for the low probability potential of leakage and impact above the AWQS. Additionally, the proximity of the R-aquifer monitoring/supply well to the breccia pipe makes it an appropriate monitoring point for detection of a potential release*." (Page 31 of 118 of the *Summary and Response to Public Comments, Energy Fuels Resources (USA) Inc. (EFRI) Pinyon Plain Mine Permit # 100333 LTF # 84446* authored by the Arizona Department of Environmental Quality, Aquifer Protection Program (APP) available at https://static.azdeq.gov/wqd/pinyonplain/2022_indpermit_rs.pdf).

In other words, ADEQ believes that one (emphasis added) well in the highly karstified R-Aquifer, which is known to have very complicated hydrogeologic characteristics and local and regional flow directions, is sufficient to determine, with a reasonable accuracy, the directions of groundwater flow and transport of any contaminant reaching the R-aquifer.

The professional judgment of this author is that ADEQ's position is nonsensical from a hydrogeologic standpoint and contrary to the entire professional and academic practice of groundwater science and engineering in any porous media and type of aquifer, not just in karst.



Figure 3.5.13 Redwall limestone, main member of the R-aquifer. In its approval of the *Pinyon Plain Mine* permit, ADEQ states that one (emphasis added) monitoring well in the R-aquifer is sufficient to determine the directions of groundwater flow and transport of any contaminant potentially reaching the R-aquifer. Photo courtesy of David Kreamer.

Below are excerpts from an article "Dept. of Environmental Quality allows increase in arsenic levels by Pinyon Mine" by Sam McLaughling, Arizona Daily Sun, July 12, 2026:

The Arizona Department of Environmental Quality (ADEQ) approved a change on Monday, July 6, to the allowable levels of arsenic in a groundwater monitoring well at the Pinyon Plain uranium mine near Grand Canyon National Park, despite the objections of the Havasupai Tribe.

Energy Fuels Resources Inc, the mine's owner, requested an amendment to the mine's Aquifer Protection Permit that would increase the maximum allowable arsenic content in the monitoring well by 10%, from 0.050 milligrams per liter (mg/L) to 0.055 mg/L. The company also sought to raise the "alert" threshold from 0.040 mg/L to 0.050 mg/L, an increase of 25%.

ADEQ originally classified the amendment request as "minor," then as "other" after the Havasupai Tribe spoke out against the proposed change. There is no public comment period for amendments classified as "other." ADEQ formally approved the amendment two weeks after publishing notice of the planned modification.

"Today, ADEQ made a choice. It could have stood with the Havasu 'Baaja and the people of Arizona and honored its obligation to protect our most precious resource. Instead, it chose to lower the bar for pollution and weaken environmental protections," Havasupai tribal chairwoman Melinda Yaiva said in a written statement released Monday afternoon.

"That decision does not simply affect the Havasu 'Baaja - it affects every Arizonan who depends on clean groundwater today and every child who will depend on it in the future," Yaiva continued. "We cannot replace a contaminated aquifer. We cannot recreate a sacred spring. We can only protect them before they are lost. The Havasu 'Baaja will never stop defending the waters that have sustained us since our beginning, because protecting water is protecting life itself."

According to Bills et al. (2010), at least fifteen springs and five wells in the region contain concentrations of dissolved uranium that exceed the U.S. Environmental Protection Agency maximum contaminant level for drinking

water (30 µg/L) and are related to the mining activities. Most of the groundwater samples represent conditions at the end of a groundwater flow path from areas of recharge to location of discharge (springs). Affected groundwater, if any, may not have flowed yet to the points sampled in this study.

The Pigeon Mine sump had a single reported value of 170 µg/L in 1986, and the Hermit Mine sump concentrations ranged from 3,310 to 36,600 µg/L (the highest reported value of any sample type in this study) in 1989−90. These high concentration mine shaft and sump waters may be sources of dissolved uranium for nearby spring sites if mine water is entering the regional groundwater flow system.

The results of hydrogeologic studies performed by Jones et al. (2017a) show that strong groundwater flow connectivity exists between the shallow groundwater system in the C-aquifer and the deep, regional karst system of R-aquifer which feeds all the largest springs in the Grand Canyon area including Havasu Spring.

Uranine injected into sinkhole in Kaibab plateau at the land surface was detected in tributaries to Bright Angel Creek downstream from Roaring Springs which discharge from the R-aquifer at thousands of feet of lower elevations, three months following injection. In addition, it stayed in the system for two more months. The timing and locations of the detected dyes indicate that the vertical connection between C and R aquifers occurs along fractures with high transmissive properties. However, the downward movement of groundwater along these vertical conduits and horizontal flow within the two aquifers is extremely complex (Jones et al. 2017a).

To further understand the complexities of the deep karst aquifer, Jones et al. (2017a) analyzed spring discharge at Roaring Springs, chosen because it is the sole water source for the park. Spring discharge hydrographs allow for the determination of aquifer water storage and discharge distributions over time because discharge can be directly measured at the spring.

Discharge analysis demonstrated a connection between the R-aquifer and surface recharge and characterized the vertical flow patterns through over 1,000 m (3,000 ft) of lithostratigraphic units, as well as horizontal flow patterns in the perched C-aquifer (Jones et al. 2017a). According to Jones et al., the hydrograph data for the deep karstic R-aquifer “indicate a complex, unknown history of transit time, flow paths and residence times” (Jones et al. 2017a, p. 16).



Figure 3.5.14 Supai Group's 800-foot rock layers as seen along Hermit Trail. NPS Photo by Kristen M. Caldon. The rocks of the Supai Group directly overly the R-Aquifer. Note numerous vertical fractures crossing the sediment layers, providing for a direct downward flowpath of any groundwater contamination coming from the overlying C-Aquifer, or from the land surface, or from a breccia pipe mined for uranium (see Fig. 3.5.17).

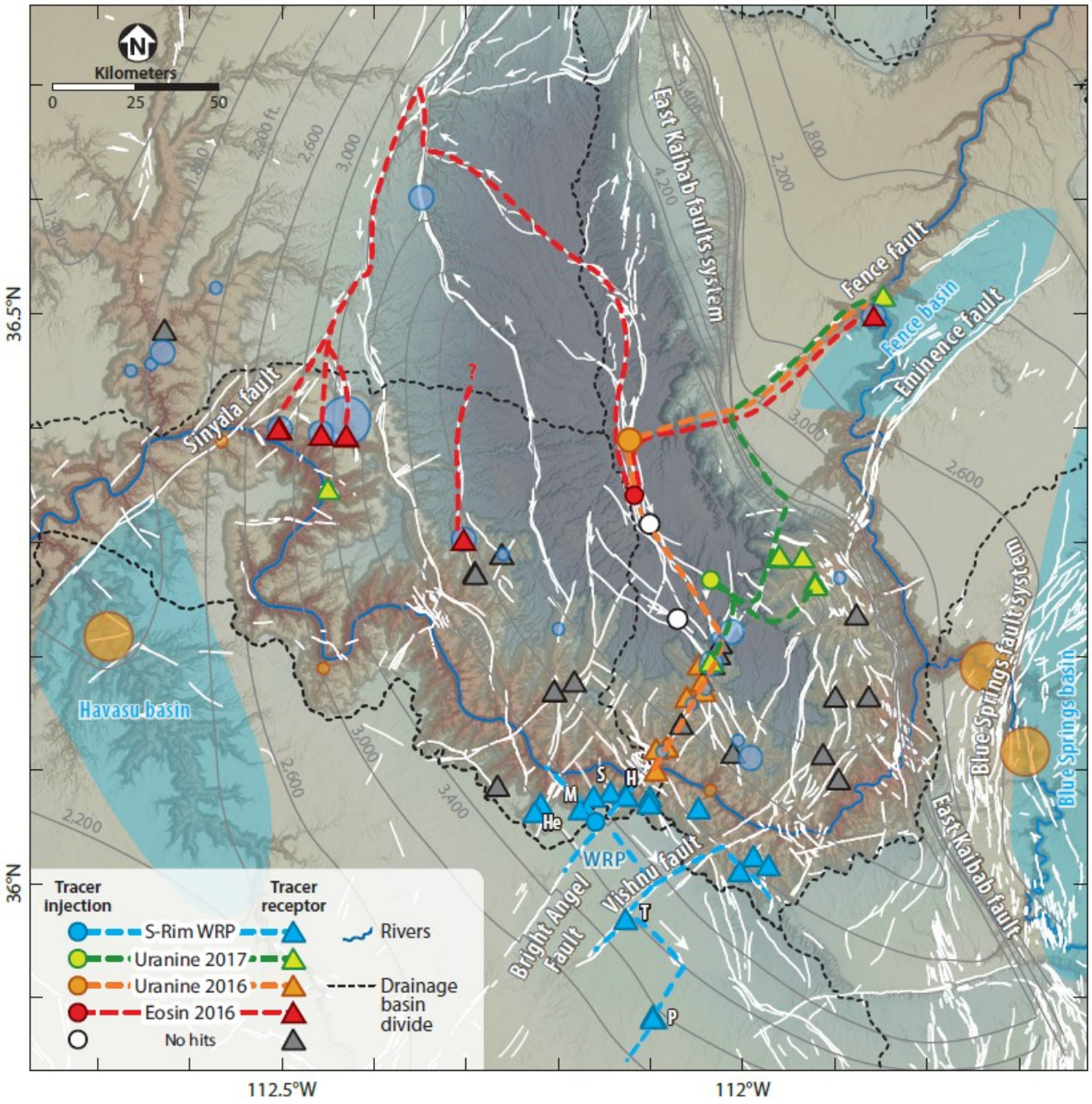


Figure 3.5.15 N-Rim dye tracer studies (Jones et al., 2018; Tobin et al., 2021) injected various dye tracers into sinkholes (*dots*); some reached detector locations (*triangles*) within weeks to months. On the S-Rim, pipeline water injected at the WRP along the Bright Angel fault is hypothesized to be affecting S-Rim springs and groundwaters based on using stable isotopes as time-averaged natural tracers (Curry et al., 2023). NW-trending faults mapped by Maxon (1968) may convey water to Hermit, Monument, Salt, Horn, and other S-Rim springs as well as to Tusayan and Pinon Plain Mine wells. Structure contours (in feet) are drawn on the top of the Bright Angel Shale aquitard (adapted from Huntoon 2000). Fence, Blue Spring, and Havasu sub-basins on either side of the Kaibab uplift are in synclinal or graben depressions and are discharging from the upper Redwall, with the R-M aquifer filled; most other springs are discharging from the Muav, with the aquifer mostly drained. Data are in Supplemental Table 1. Abbreviations: H, Horn; He, Hermit; M, Monument; N-Rim, North Rim; P, Pinon; R-M, Redwall-Muav; S, Salt; S-Rim, South Rim; T, Tusayan; WRP, Water Reclamation Plant. Copyright 2014 by L.J. Crossey, K.E. Karlstrom, B. Curry, C. McGibbon, C. Reed, J. Wilgus, C.J. Whyte, and T. Darrah.

A new look at Grand Canyon springs and possible threats from uranium mining

By New Mexico Sun, March 25, 2024

A new research paper published in Annual Reviews of Earth and Planetary Sciences sheds light on the significance of Grand Canyon's springs and groundwater, as well as the potential threats posed by uranium mining. The paper, titled Hydrotectonics of Grand Canyon Groundwater, emphasizes the importance of sustainable groundwater management and the need for better monitoring to address uranium mining threats.

Lead author Laura Crossey discussed the findings of the research, noting the role of faults as fluid superhighways that connect different aquifers. She stressed the importance of understanding these "hydro-tectonic" concepts in managing and protecting the Grand Canyon's springs and groundwater wells.

The paper also addresses the timely issue of uranium mining in the region, particularly the Pinyon Plain mine near the Park's South Rim Village. Karlstrom expressed concern over the lack of consideration for recent scientific findings in the permitting process, stating, "Tribes claim the permitting is ignoring recent peer-reviewed science and risks to culturally significant features."

The authors of the paper advocate for caution and oppose mining in the sensitive region due to the significant risk of contamination, especially of the Havasupai springs that supply water to the Havasupai Village. The paper serves as a comprehensive summary of the science surrounding Grand Canyon groundwater, urging state and federal agencies to prioritize the protection of this unique and vital ecosystem.

The paper in question authored by Crossey et al. (2024) is available at https://doi.org/10.1146/annurev-earth-080723-083513. Figure 3.5.15 is from it, illustrating the complexity of groundwater flow in the Grand Canyon area including the hydraulic connectivity of water flow between the ground surface and the springs discharging from the R-aquifer, throughout the entire thickness of both the C- and R-aquifers and any intervening fractured aquitards.



Figure 3.5.16 White Mesa community members participate in the 2019 White Mesa spiritual walk, an annual protest walk organized by Ute. Mountain Ute tribal members are concerned about the uranium mill's impacts on groundwater, air quality, and public health. Photo by Tim Peterson, from Reimondo, 2020; https://www.grandcanyontrust.org/Canyon-Mine-Report.

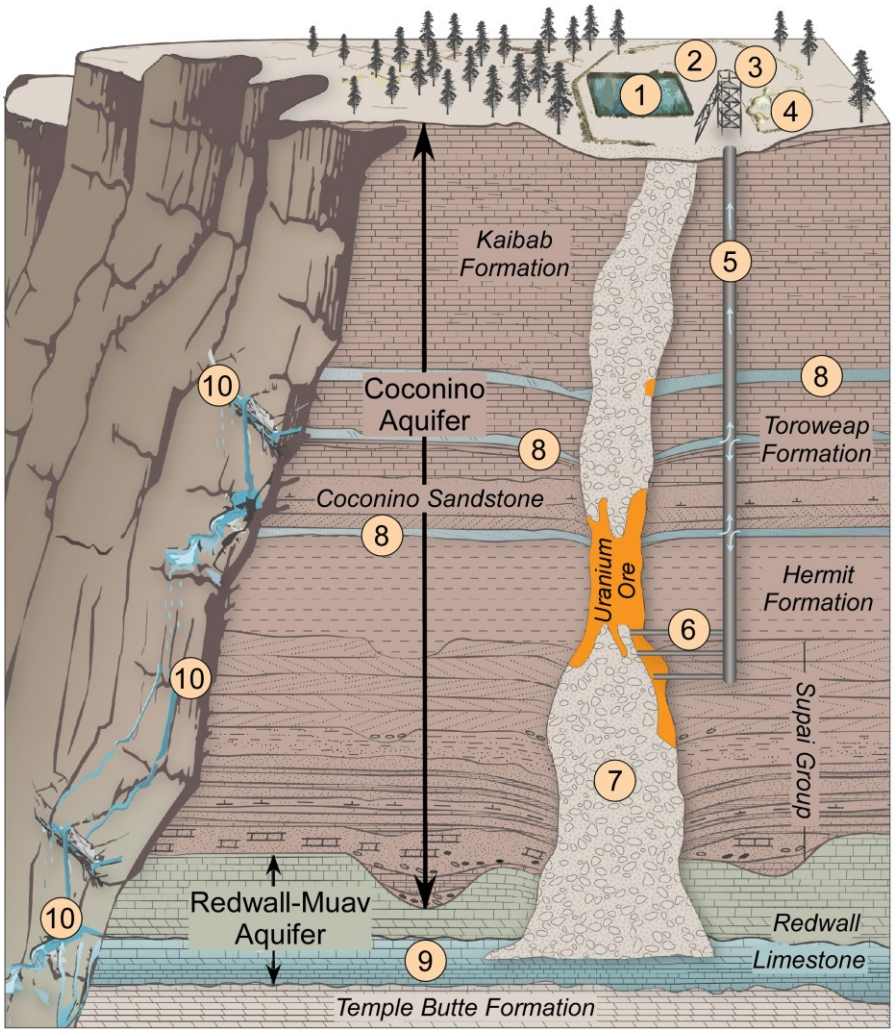


Figure 3.5.17 Lithology and mine features at breccia-pipe uranium mining sites in the Grand Canyon region. From Reimondo, 2020. (1) Containment pond; (2) Ventilation shaft; (3) Mine headframe; (4) Waste rock, ore pile, and top soil storage; (5) Mine shaft; (6) Horizontal shaft ("drifts"); (7) Breccia collapse feature; (8) Perched aquifer; (9) Regional aquifer; (10) Seep or spring. Potential water flow in mine shaft is shown with blue arrows. Generalized and modified from the USGS report Site Characterization of Breccia Pipe Uranium Deposits in Northern Arizona and Uranium Mine Conceptual Model. Graphic by Stephanie Smith, Grand Canyon Trust. Available at https://www.grandcanyontrust.org/Canyon-Mine-Report

High-grade uranium ore was discovered in geologic features called "breccia pipes" (Figure 3.5.17) in the Grand Canyon region during the late 1940s and became the subject of intense exploration during the 1970s (Otton and Van Gosen, 2010). Historical uranium mining of breccia pipes in northern Arizona has left a legacy of abandoned mines, mine waste, and waste rock exposed at the surface, allowing subsequent remobilization of minerals and radionuclides and contamination of both surface water and groundwater systems in the region. Although the advent of stricter environmental regulations and improvements in mining practices reduced the potential for environmental contamination, secondary permeability in fractured and karstified rocks provides pathways for dispersing remobilized radionuclides. Breccia pipes are particularly important in that regard since they have high vertical hydraulic conductivity as demonstrated by USGS scientists (e.g., Hayes, 1999; Epstein and Putnam, 2005).

Today, indigenous communities on the Colorado Plateau continue to bear the brunt of America's sordid past when it comes to uranium extraction. More than 500 abandoned uranium mines remain in need of careful

assessment and cleanup on Navajo Nation lands alone, where a recent study found that over a quarter of 781 Navajo women tested had elevated levels of uranium in their bodies, as did newborn babies. The Navajo Nation banned uranium mining on its 27,000-square-mile reservation in 2005, but the abandoned mines remain (Reimondo, 2020).

As explained by the U.S. Geological Survey (Tilousi and Hinck, 2024), growing public concern that uranium mining activities could have adverse environmental, cultural, and social impacts prompted a withdrawal of approximately one million acres of Federal lands from future mineral extraction to study the potential effects of uranium mining (Alpine, 2010) and to prepare an environmental impact statement (Bureau of Land Management, 2011). In 2012, the former Secretary of the Interior, Ken Salazar, closed the Federal lands in the Grand Canyon watershed to new mining claims under the 1872 Mining Law until the year 2032 (Bureau of Land Management, 2012; U.S. Department of Interior [DOI], 2012.



Figure 3.5.18 Travertine pools ("Beaver Falls") on Havasu Creek, Havasupai Tribe Reservation, Grand Canyon. The entire ecosystem supported by the Havasu Springs and the Creek is now under threat because of the uranium mining interests and operations of the Pinyon Plain Mine. The Havasupai Tribe calls themselves Havsuw' Baaja (or *Havasu 'Baaja*), which translates directly to "the people of the blue-green water." Photograph courtesy of the Havasu Tribe. Available at https://www.facebook.com/HavasupaiTribeTourismOfficial

The Kaibab Plateau Springs

The largest springs that discharge from the Kaibab Plateau, north of the Colorado River (see map in Fig. 3.5.3), are the Tapeats, Thunder, Cheyava Falls, and Deer Springs which drain most of the plateau via the West Kaibab Fault Zone. The base flows from these springs are about 50 ft^3/s, although flood flows can be several times this amount (Johnson and Sanderson, 1968). Bright Angel Creek Springs and Roaring Springs drain the south-central part of the Kaibab plateau via caves present on fractures or parallel to the Bright Angel Fault, which in turn is hydraulically linked to normal faults that trend along the axis of the Kaibab uplift.

Roaring Springs serves as the sole water supply for both the north and south rim developments in Grand Canyon National Park. In its current design, gravity plays a major role in the functioning of the waterline, creating sufficient pressure to flow water to Havasupai Gardens (formerly known as Indian Gardens) from the spring intake at an elevation of 5,200 feet. Water travels down along the North Kaibab Trail in the waterline, through Phantom Ranch. The waterline crosses the Colorado River and reaches its lowest elevation of 2,400 feet before it ascends the canyon and enters the Havasupai Gardens pump houses at an elevation of 3,700 feet. From there, water is pumped up 3,300 feet in elevation to the South Rim through the borehole section of the waterline.



Figure 3.5.19 Where the descending North Kaibab Trail meets the flatter bottom of Bright Angel Canyon, 5 miles from the North Kaibab Trailhead, **Roaring Springs** may be seen. Here, water gushes forth directly out of the cliffs, cascading over moss and fern to form Bright Angel Creek (which will be a constant companion all the way to the Colorado River). This large spring provides drinking water for every visitor and resident within Grand Canyon National Park. The water is delivered to the South Rim via a pipeline buried beneath the North Kaibab Trail (installed 1965-1970). NPS photos by Michael Quinn available at https://www.flickr.com/photos/grand_canyon_nps/albums/ under https://creativecommons.org/licenses/by/2.0/#. The park's new water bottle filling stations provide free spring water from Roaring Springs. Learn more about water bottle filling stations here: www.nps.gov/grca/planyourvisit/refilling_stations.htm



Figure 3.5.20 *Left*: **Deer Spring**. *Right*: Deer Creek Waterfall is a stop on most Colorado River Trips through the Grand Canyon. NPS Photos by Erin Whittaker, taken in August 2011.

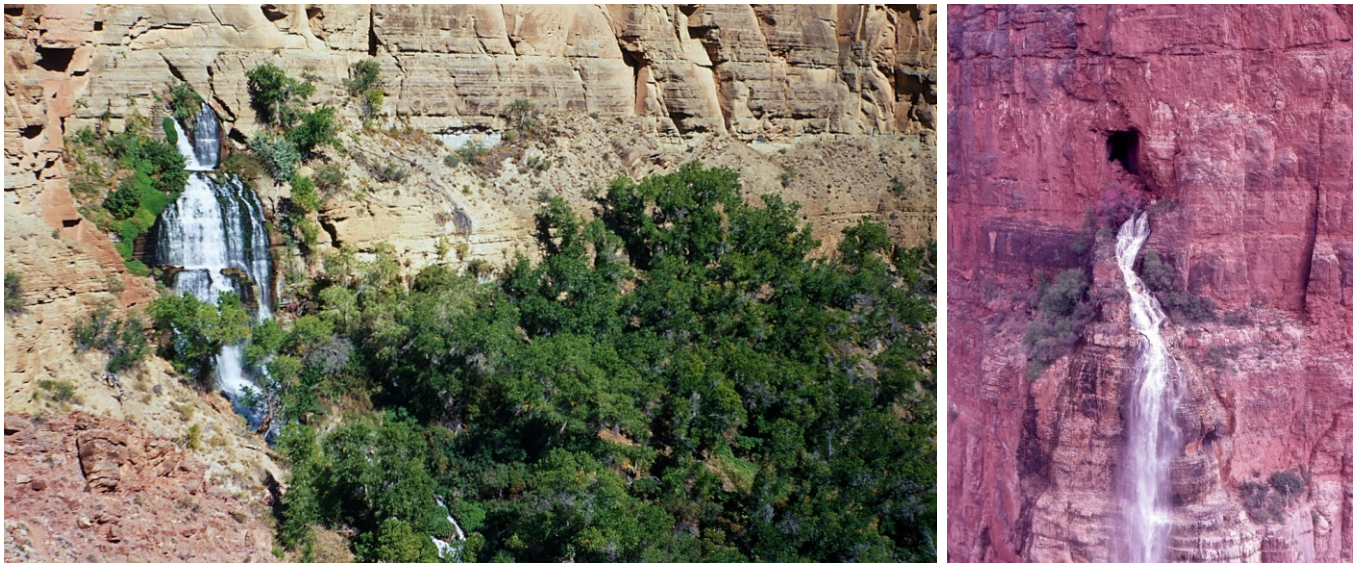


Figure 3.5.21 *Left*: View northeast toward **Thunder Spring** from Thunder River trail. Spring issues from fractures in Peach Springs Canyon and Kanab Canyon Members of Muav Limestone. *Right*: Intermittent **Cheyava Falls Spring** flow coming out of cave in Mooney Falls Member of Redwall Limestone, east wall of upper Clear Creek canyon. Photographs from Grand Canyon collection of Billingsley et al., 2019; USGS.

The only continuous streamflow records that exist for the Kaibab Plateau springs are for the Bright Angel Creek (USGS station 09403000) from 1924 to 1974 and from 1991 to 1993 (Fig. 3.5.22). These measurements serve as a useful proxy for spring discharge because the creek is fed by a series of karst springs including Roaring Springs.

The National Park Service (NPS) studies and design process has resulted in a decision to relocate the water intake to Bright Angel Creek near Phantom Ranch. This will create a second water source for the park with the

South Rim, Havasupai Garden and Phantom Ranch receiving water from the new intake at Bright Angel Creek while Roaring Springs will continue to provide water to the North Rim, Manzanita and Cottonwood.

Moving the water source to Bright Angel Creek will enable additional streams and watersheds to feed into the water intake and provide more resiliency for future water conditions. The location will also reduce the pipe length and eliminate the portion of the current waterline that experiences the most frequent failures, an area north of Phantom Ranch known as "The Box". The new waterline will stretch nearly 3 miles from Phantom Ranch to Havasupai Gardens (formerly known as Indian Gardens). With this design, natural flow would be restored to Bright Angel Creek from Roaring Springs.

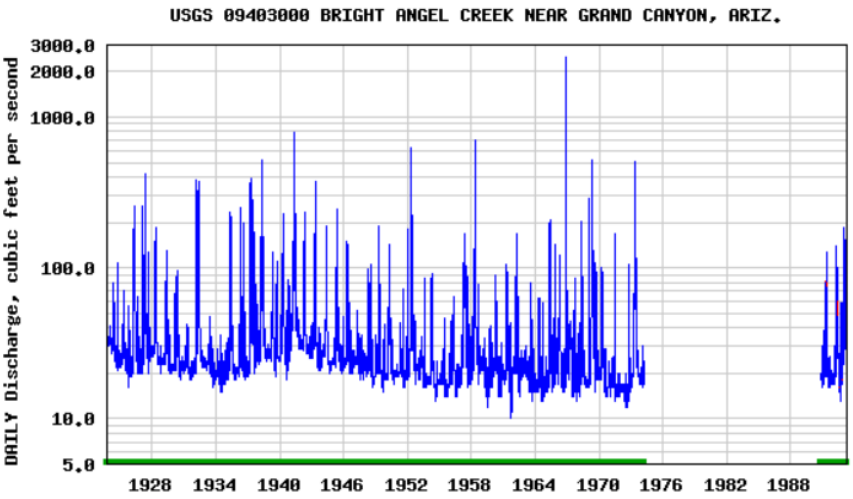


Figure 3.5.22 Hydrograph of the Bright Angel Creek flow rate for the period of record.



Figure 3.5.23 View west and upriver toward **Vaseys Paradise Spring** from river mile 32.2; Redwall Limestone in Marble Canyon. The Spring discharges from a cave more than 1.9 miles long. Photograph from the Grand Canyon collection of Billingsley et al., 2019; USGS.

Montezuma Well, Yavapai County, Arizona

Montezuma Well is a spring-fed sinkhole on the north bank of Wet Beaver Creek, about 8 miles northeast of the town of Camp Verde in the Verde Valley of Arizona. Montezuma Well is part of the Montezuma Castle National Monument, managed by the National Park Service (NPS).

As explained by NPS, Montezuma Well is a natural sinkhole 368 feet wide, with cliffs that tower 70 feet above the water's surface. Every day over 1.5 million gallons of warm (70-74°F) water flows into the Well, which is a near constant aquatic environment. The Well formed when the limestone collapsed sometime between 12 and 15 thousand years ago. Water is fed into the Well through four large underwater vents, and exits the Well through a swallet and a cave 300 feet long, with the outlet draining into a prehistoric irrigation canal. The canal was originally constructed almost 1,000 years ago by the people of the Sinagua (see-NAH-wah) culture who lived and farmed here for centuries (https://www.nps.gov/moca/planyourvisit/exploring-montezuma-well.htm; virtual tour of the Well is provided by NPS at Montezuma Well Virtual Tour - Montezuma Castle National Monument (U.S. National Park Service) (nps.gov).

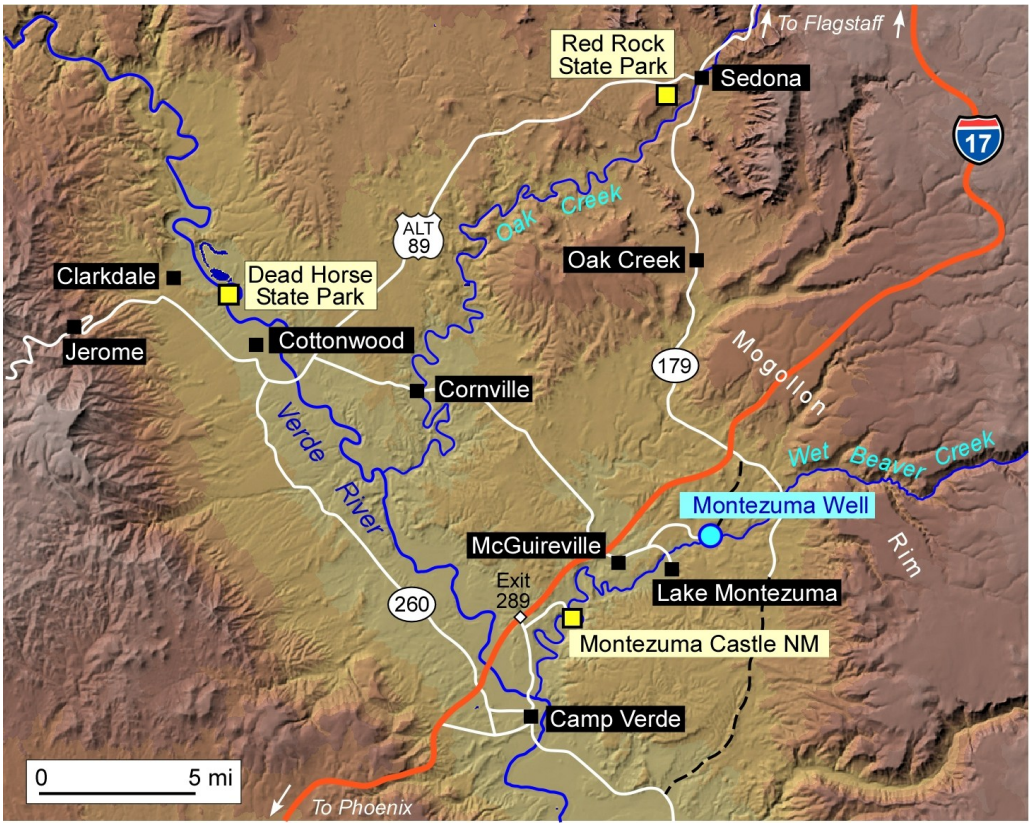


Figure 3.5.24 Location map for Montezuma Well, which is part of Montezuma Castle National Monument (NM). The monument lies in central Arizona, off interstate highway 17 (Exit 289). It is composed of two units: the Well Unit, which contains Montezuma Well, and the Castle Unit, which contains the Montezuma Castle cliff dwellings. Locations of two spectacular Arizona state parks in the vicinity are also shown. Shaded relief map courtesy of USGS.

Both the water level and temperature are nearly constant throughout the year, with the first fluctuating by only 6 inches and the latter averaging 70 degrees Fahrenheit (Blinn 2008). The water also has a stable pH because of high alkalinity and an excessive amount of dissolved carbon dioxide, which enters the well from the bottom vents in concentrations in excess of 500 mg per liter, more than 100 times higher than normal (Blinn 2008). In addition,

the concentration of dissolved arsenic of 114 micrograms per liter is an order of magnitude higher than any other groundwater tested in the area (Johnson et al., 2011). These extreme chemical conditions mean that no fish can survive in the Well, and that organisms living in the water have had to adapt to survive. This unique and isolated ecosystem enabled amphipods (small, shrimp-like animals), and the leeches which feed on them, to evolve into endemic species found nowhere else in the world.

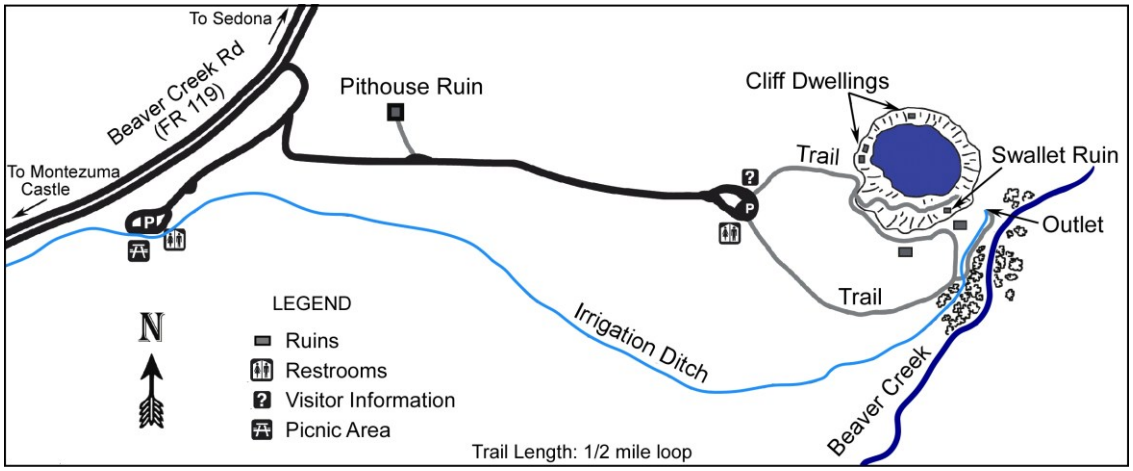


Figure 3.5.25 Montezuma Well is a natural sinkhole 368 feet wide, with cliffs that tower 70 feet above the water's surface. Every day over 1.5 million gallons of warm (70-74°F) water flows into the Well, fed by large underwater vents. The water exits the well through a cave 300 feet long, emerging on the southeast corner of the mound into an irrigation ditch originally constructed almost 1,000 years ago by the people of the Sinagua. Courtesy of NPS.



Figure 3.5.26 Montezuma Well on a clear day. NPS Photo by Ed Goodwin.



Figure 3.5.27 *Left*: People have been using the constant supply of warm, fresh water flowing from Montezuma Well to irrigate their crops for over 1,000 years. This section of a "fossilized" 800-year-old irrigation ditch can be seen near the picnic area at Montezuma Well. The walls of this ditch were formed from deposits of lime that built up over hundreds of years of use. *Right*: An irrigation ditch built by the Sinagua Indians near Montezuma Well about 800 years ago. The water it carries is still used today. Photographs courtesy of NPS. USGS reports that monthly mean discharge of the Montezuma Well outlet ranged between about 1 and 3 cfs for the 1977-1992 period of record (Konieczki and Leake, 1997, p.4).

The land around Montezuma Castle and Well has been home to many prehistoric groups of people since as early as 11,000 BC. The fascinating geologic and cultural history of Montezuma Well, Montezuma Castle and this part of the American southwest are described in a 1958 handbook by NPS that can be found at http://www.npshistory.com/handbooks/historical/27/hh27b.htm and https://www.gutenberg.org/files/49607/49607-h/49607-h.htm

"How could water be the most important player in a story about the desert? In an entire year, Montezuma Well receives less than 13 inches (33 cm) of rainfall—barely ⅓ of the national average for the United States. Yet the Well contains over 15 million gallons (56.8 million liters) of water! Where does it come from? How did it get there? Until 2011 the answers to these questions were a mystery. Though water may be easily turned aside, it is patient, persistent, and unrelenting. More than 10,000 years ago, the Well's water fell as rain and snow atop the Mogollon Rim, visible to the north. Over millennia, it has percolated slowly through hundreds of yards of rock, draining drop by drop through the path of least resistance. But here the water encounters an obstacle much harder than the others through which it has flowed. Beneath the Well, a vertical wall of volcanic basalt acts like a dam, forcing water back toward the surface. In its long trip toward daylight, it eroded an underground cavern until its roof collapsed and created the sinkhole you see today. The water still flows. Every day, the Well is replenished with 1.5 million gallons (5.7 million liters) of new water. Like a bowl with a crack in its side, the water overflows through a long, narrow cave in the southeast rim to reappear on the other side at the outlet. Side trails lead from the main loop down to cool, shaded benches at each end of this subterranean waterway."
(https://www.nps.gov/moca/planyourvisit/exploring-montezuma-well.htm)

Unfortunately, the above upbeat excerpt from the NPS narrative is based on various assumptions, sometimes contradictory to each other, and the Well remains an unsolved mystery. Figure 3.5.28 shows a cross section featured on various NPS webpages dedicated to Montezuma Well and reproduced widely on the Web by many. It shows that the Well is formed in thick continuous deposits of travertine, and water leaves it via a long "cave" in travertine.

The main source of water to the Well, shown in Figure 3.5.28, is deeply seated Redwall limestone. The upward groundwater flow to the Well basin is forced by an assumed, thin, vertical basalt dike located directly below the Well basin. The assumed dike cuts the entire thickness of all the sediments shown, including young travertine

deposits directly below the Well basin. Its existence was inferred from a geophysical survey. Notably, no dikes were mapped at the land surface near or farther away from the general area of the Well.

The cross section in Figure 3.5.28 contradicts other reports such as Johnson et al. (2011), Konieczki and Leake (1997), Vance (2014), and NPS' own narratives including pamphlet *Montezuma Castle National Monument, Montezuma Well* available at https://www.nps.gov/moca/planyourvisit/exploring-montezuma-well.htm:

"Over millions of years, underground streams dissolved away the soft limestone formed by sedimentation from Lake Verde, producing caverns below the surface of the Verde Valley. 11,000 years ago one of these caverns collapsed into a sunken pool, creating what we now call Montezuma Well."

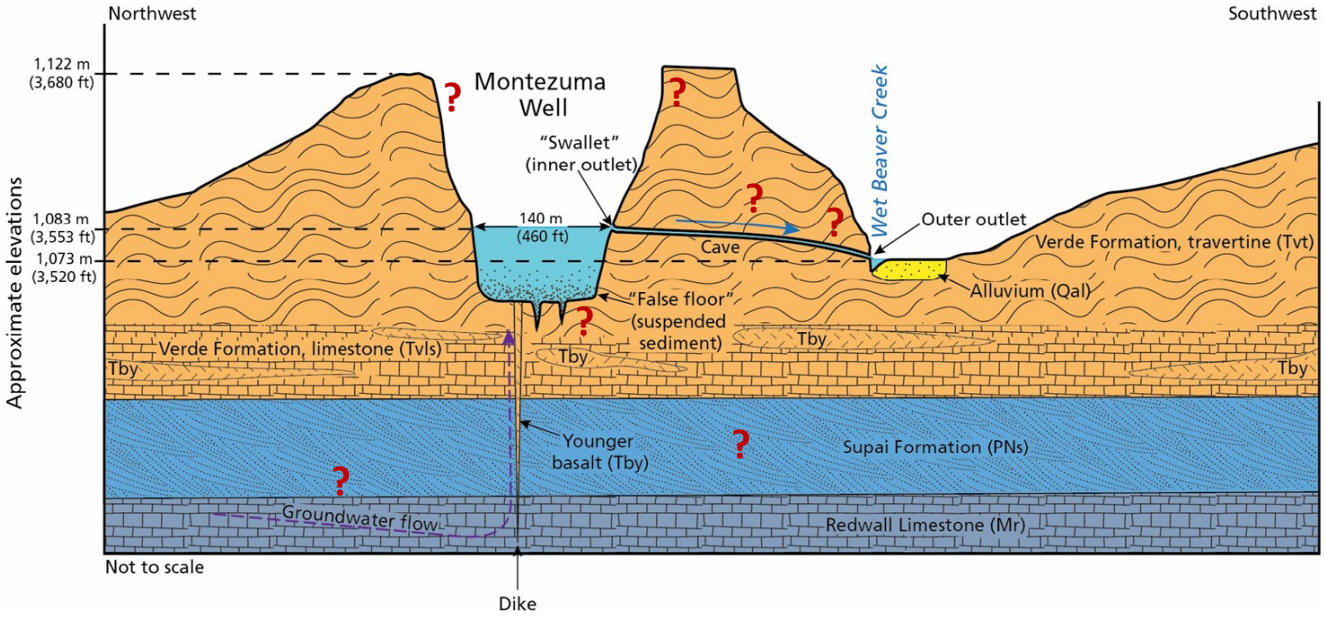


Figure 3.5.28 Figure and caption from KellerLynn, 2019 (question marks are added by the author of this book): "Generalized cross section through Montezuma Well. Montezuma Well occurs in the Verde Formation, travertine (Tvt). Groundwater (represented by the purple arrow on the graphic) follows flow paths reaching permeable fractures along and above a basalt dike below the well. These fractures serve as conduits that carry groundwater and deep sourced carbon dioxide (CO2) to the surface, which dissolves carbonate minerals along the transport path in response to the added CO2; this mechanism likely formed the cavity of Montezuma Well. At the surface, CO2 degasses, depositing travertine. As evidence, travertine coats the current irrigation ditch that transports water away from Montezuma Well. Groundwater enters Montezuma Well with such force to keep sand particles in suspension, creating the "false floor." At the surface, water flows through the swallet, cave, and outer outlet into an irrigation ditch. Under natural conditions, Wet Beaver Creek would transport this water away from the well. Graphic by Trista Thornberry-Ehrlich (Colorado State University) after Konieczki and Leake (1997, figure 15) and Johnson et al. (2012a, figure 9) with information from Lange (1957) and Lenihan (2011)."

Johnson et al. (2011) state that "The source of water to Montezuma Well, a flowing sinkhole in a desert setting (see picture on cover), is poorly understood. Water emerges from the middle limestone facies of the lacustrine Verde Formation (Twenter and Metzger, 1963), but the precise origin of the water and its travel path are largely unknown (McKee et al., 1947; Lange, 1957). Some have proposed artesian flow to Montezuma Well through the Supai Formation, which is exposed along the eastern margin of the Verde Valley and underlies the Verde Formation (Konieczki and Leake, 1997). The groundwater recharge zone likely lies above the floor of the Verde Valley somewhere to the north or east of Montezuma Well, where precipitation is more abundant."

According to Konieczki and Leake (1997), "The Supai Formation, consisting primarily of sandstone and siltstone underlies (emphasis added) the Verde Formation beneath the Montezuma Well pond and is exposed east of the pond. Near Montezuma Well, drillers' logs indicate that the Supai Formation is overlain by the Verde Formation and stream alluvium to a depth of about 52 feet. Near Montezuma Well, the Verde Formation is thin and has been removed by erosion to the east. Drillers' logs show a difference in composition of the Verde Formation between the area around Montezuma Well and the area near Camp Verde. The lithology near Montezuma Well is mainly limestone with some basalt flows, and the formation has extensive fractures and solution features. (emphasis added; see also Fig. 3.5.29).



Figure 3.5.29 Outlet of Montezuma Well in Verde Formation limestone (not in travertine as shown in Figure 3.5.28). Photo by Meghann M. Vance, 2010. From Vance, 2014.

Most of the water supply wells in the area were drilled into the Verde Formation, and a few wells in the Beaver Creek drainage basin penetrate the Supai Formation. Water levels in wells completed in the Verde Formation range from flowing at the land surface to about 61.0 m (200 ft) below land surface. Well owners report that some wells drilled into the Verde Formation flow during the winter and not in the summer."

To make their groundwater flow model work, Konieczki and Leake (1997), assumed a hypothetical horizontal flow barrier and a vertical conduit to force groundwater flow upwards and discharge at Montezuma Well. A basalt dike was never mentioned in the entire report, nor were travertine deposits.

"The possibility remains, however, that the flow system in the Verde Formation could include solution channels that transmit water to Montezuma Well from other areas within the Verde Formation. Although the mechanism for inflow from the Verde Formation is not understood, the Verde Formation and the underlying Supai Formation probably are the sources of water to Montezuma Well." (Konieczki and Leake, 1997.)

Vance (2014) states that "The Well is fed by two powerful underwater vents and maintains a constant temperature year round. Water exits the Well at the southern edge, traveling through a swallet (a dissolved channel in the stone that drains the surface water) some 300 feet through the limestone (emphasis added) surrounding the Well and exiting into a prehistoric irrigation canal on the other side, near Beaver Creek. Roughly 1.5 million of gallons, 74 degrees in temperature regardless of the time of year, flow out of the Well every day."

On the cross-section AA' by Johnson et al. (2011; Fig. 3, p.16), there is not Supai formation directly below travertine and limestone; instead, it is Tertiary Verde Formation limestone which overlies several hundred feet thick basalt (including an assumed thin and vertical basalt dike, not mapped anywhere in the vicinity of the Well). Only below this thick basalt is sandstone of Supai Formation, and below all that are Mississippian Naco and Redwall limestones.



Figure 3.5.30 The people who lived at Montezuma Well over 700 years ago spent their winters living in these southeast-facing cliff dwellings. They built their homes here to take advantage of passive solar heating during the coldest months of the year. Note horizontal, fractured and cavernous limestone layers at the top of the cliff. Photo courtesy of NPS.

Importantly, in Johnson et al., (2011; Fig. 50, p.60), the groundwater flow direction is from the east-northeast (from the Transition Highlands of the Mogollon Rim– see map in Fig. 3.5.24) to the west-southwest and the Verde Valley, and not from northwest to southwest as shown on the cross-section in Fig. 3.5.28.

Travertine is a young sediment deposited from warm groundwater, supersaturated with calcium carbonate, flowing to the land surface. It is virtually impermeable as evident by numerous travertine dams around the world which impound flowing water and create pretty waterfalls. Recently deposited travertine is not readily or quickly dissolved by water leaving "empty" spaces or voids behind. One cannot expect long, dissolutional, tubular caves to be formed in it.

Hypothetical transmissive vertical fractures in travertine, shown in Fig. 3.5.28 to extend below the Well pond and facilitate an exceptionally strong inflow of groundwater through the "vents" are unlikely to develop in travertine, but could arguably be formed in areas of strong seismic and tectonic activity.

Before a sinkhole collapse occurs, the material that has been eroded and/or dissolved must be transported away by groundwater, via karst conduits and their network, so that the cavity can be formed. When a sinkhole collapse occurs, the material that collapsed remains at the bottom of the cavity and can also be transported away via a conduit network. None of these mechanisms is likely to exist within the travertine deposits shown on Fig. 3.5.28.

Below are several excerpts from a fascinating story about various Well diving expeditions told by Steve Ayers, courtesy of Verde Valley Newspapers (https://www.nps.gov/moca/learn/photosmultimedia/dive-to-the-bottom-of-the-well.htm). Especially telling is the encounter with the force of groundwater coming up to the Well at its bottom, effectively preventing any downward removal of the geologic material by groundwater, before or after a "collapse" in the mischaracterized travertine deposits.

- *The report of diver H.J. Charbonneau seemed to verify a conclusion drawn by the famous newspaper reporter and adventurer, Charles Lummis, some 60 years earlier, that the well was a "creepy place."*
- *In 1962, diver G.J. Murray became the first to report that the mysterious bottom was perhaps not the bottom.*
- *In an article in Skin Diver magazine, Murray reported of the eeriness of swimming "in a 'bottomless pit' with thousands free swimming leeches."*
- *Murray labeled it a bottomless pit after observing that the bottom of the well appeared as "an irregular boiling surface, like that of thin mush cooking."*
- *Numerous other divers have since reported the strange layer of sandy sediments, describing the "false bottom" as "a white lava flow moving on top of a suspended bottom with a silica gel consistency," or "quick sand" or even "boiling oatmeal."*
- *A probe dropped into the west vent went down 69 feet through the fluidized sand. Added to the 55 feet from the water's surface to where the sand layer starts, makes the west vent 124 feet deep overall. The east vent measured 19 feet, or 74 feet overall.*
- *Of all the discoveries made during their 2006 dive, perhaps the most interesting for the dive team was the correlation of their scientific research with the traditional stories of the Yavapai and Apache people.*
- *One day while Conlin and Lenihan were studying the well, they met a group of tribal elders, who told of stories from their past that said there was a place at the bottom from which once something emerged, it could never return.*
- *"It was a very interesting view for us, in all our scientific splendor, and it kept coming back to us," Lenihan said. "Everything we tried putting down those holes -- cameras, rovers, sensors -- kept being pushed back out. It was an interesting convergence of how they viewed the well and how science looks at it."*

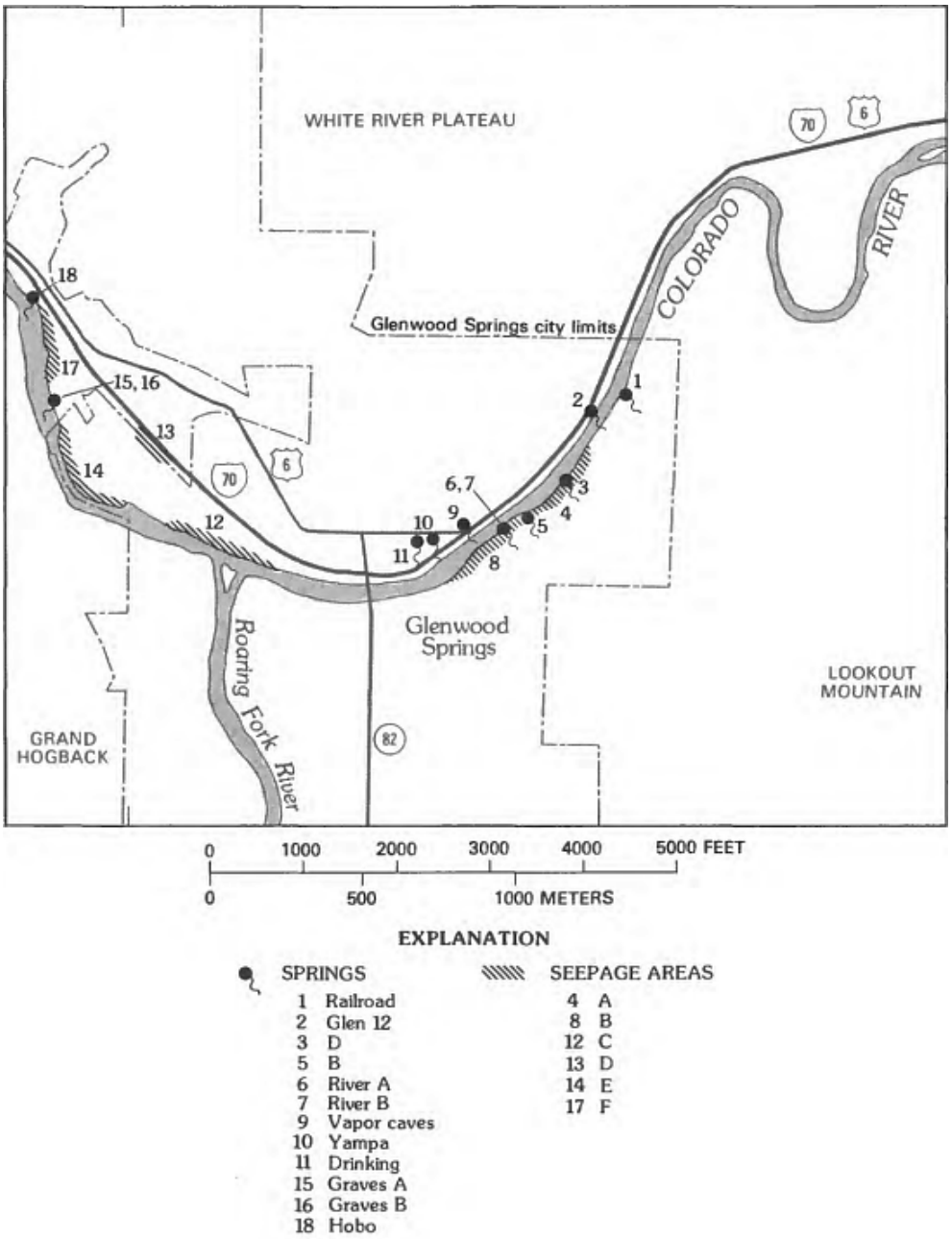
Whatever the case may be regarding various geologic and hydrogeologic inconsistencies mentioned earlier, Montezuma Well remains a great mystery. Before this mystery is solved by experienced karst hydrogeologists, one can only read with awe the following from a Native American oral tradition:

The Hopi have been an agricultural people for a long time. However, even though we have always grown our own food, life was still difficult. When our ancestors (who people now call Sinagua) found the Well, they realized life could be easier. So they stayed and made it their home. However, life became too easy and the people became complacent. It was as though they had forgotten who they were and how they should live. The village was in a state of shameless corruption. One day, the village began to shake and the houses caved in. Water gushed in and filled the hole where the village once stood. Suddenly, a great serpent rose up out of the water in anger at what the people had become. Upon seeing the serpent, everyone became very scared and ran away to the north and east, never to return.

Summary of “The Legend of Montezuma Well”, as told by
Fred Lomayesva of the Hopi Snow Clan

Glenwood Hot Springs, Garfield County, Colorado

Twelve hot springs and 6 seepage areas, collectively known as the Glenwood Hot Springs Group, occur along both banks of the Colorado River on the northeastern and northwestern sides of the town of Glenwood Springs. These springs and seepage areas cumulatively discharge at a rate between 4000 and 5000 gpm. All springs issue from either the Mississippian Leadville Limestone (equivalent to Madison Limestone) or alluvium overlying the Leadville Limestone and Early to Middle Pennsylvanian Belden Formation; springs issuing from the alluvium are believed to be supplied by water migrating up from the Leadville Limestone through fractures and faults (Barrett and Pearl, 1977, p. 92).



The largest spring in the area, the Yampa (Big) Spring was first developed for recreational use in 1888. Successive modifications have made measurement of its discharge difficult. Currently, the spring flows upward into a 60-ft diameter, rock-lined caisson with four outlets, one each to the Glenwood Springs Lodge and Pool, and two to the Colorado River.

The spring discharge varies as the water level in the caisson is changed by manipulation of gates to the swimming pool and river. Barrett and Pearl (1977, p. 92) report a discharge of 2263 gpm from this spring. Average discharge from the caisson when all flow is diverted through the river gate is about 2500 to 2700 gpm.

Figure 3.5.31 Locations of springs and seepage areas at Glenwood Springs. From Geldon, 1989.

The Devonian and Mississippian carbonate rocks hydrostratigraphic unit in the Glenwood Springs area consists of the Dyer Dolomite and the Leadville Limestone. The temperature and chemistry of water discharging from the Leadville Limestone at Glenwood Springs indicate that most of the water originated in highlands to the south. Moisture from precipitation and snowmelt in the Lookout Mountain and Grand Hogback areas descends to the Leadville Limestone and Dyer Dolomite through several thousand feet of upper Paleozoic, Mesozoic, and Tertiary sedimentary rocks; locally, the moisture descends through Tertiary basalt. The water is heated as it descends through the overburden and becomes saline by dissolution of halite and gypsum in the Eagle Valley Evaporite. The hot, saline water moves through fractures and solution channels in the Leadville Limestone and Dyer Dolomite toward the Glenwood Springs area, where it mixes with cool, freshwater from the White River Plateau, and discharges as hot springs or seeps into stream alluvium. Artesian flow from wells completed in the Leadville Limestone at Glenwood Springs affects discharges from nearby hot springs and water levels in the alluvium (Geldon, 1989).

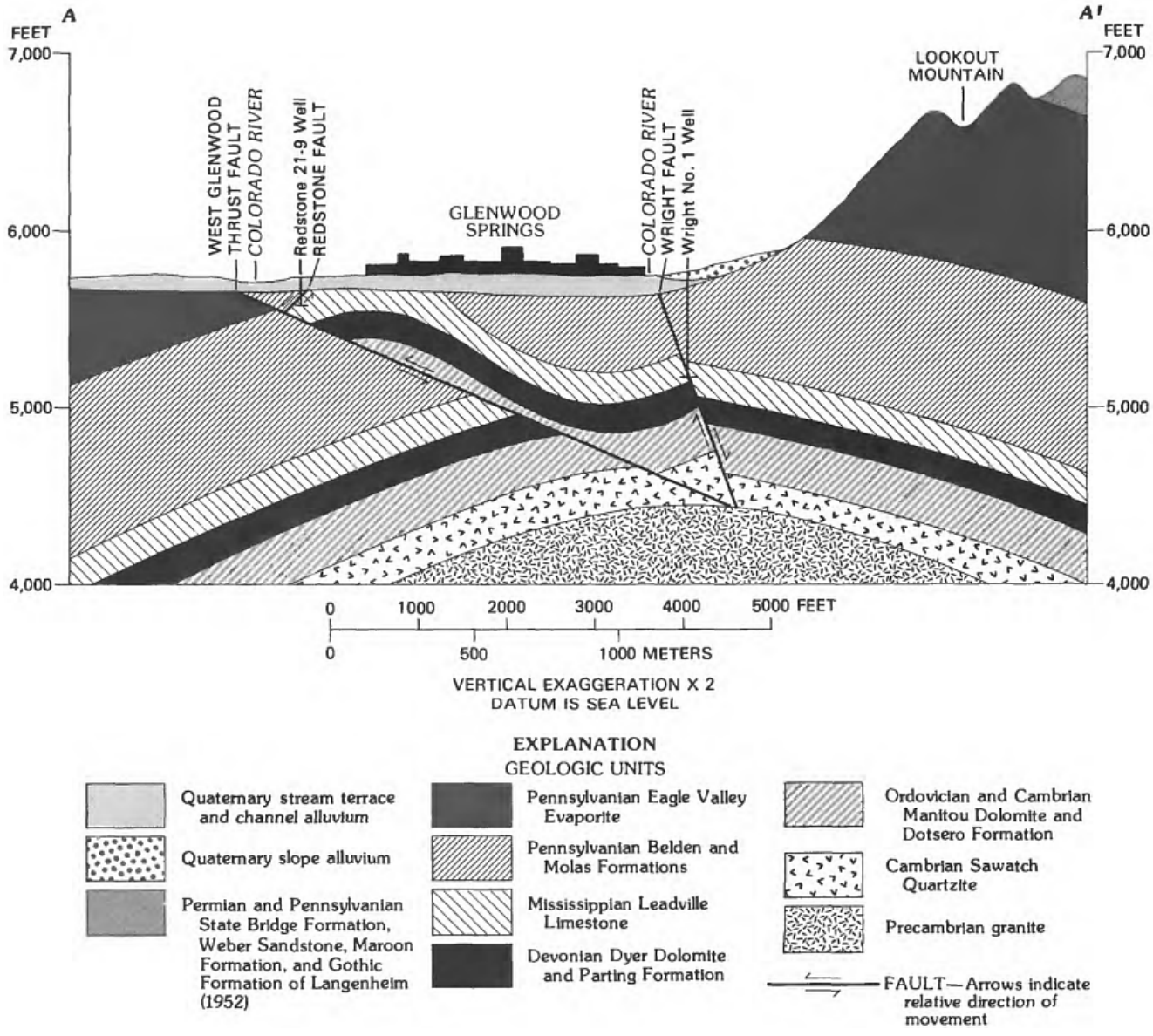


Figure 3.5.32 Geology and structure at Glenwood Springs. From Geldon, 1989.

The Ute Indians were a group of nomadic hunter-gatherers. For countless generations, they made regular treks through the Glenwood Springs region following the animal herds and the weather. The geothermal wonders of the area were well known to them and they named the spring Yampah or Big Medicine in the Ute language.

The Utes believed the springs were a magical gift from the earth. They sanctified the location to Manitou, their great creator spirit. In addition to soaking in the warm spring waters along the Colorado River, they used the nearby vapor caves as natural steam baths to sweat out sickness. Tribal warriors would immerse themselves—and sometimes even their horses—in the hot springs to gain strength and energy before going on big hunts or facing rival tribes. The waters were so highly valued that the Utes occasionally had to fight other tribes, like the Comanche and Arapaho, who traveled to the area wanting to use the healing springs.

Sadly, by the late 1800s, the U.S. government forced the Ute people off their ancestral lands and onto reservations, paving the way for developers to turn the sacred waters into a commercial resort.

A legend of the American West, John Henry "Doc" Holliday is buried in Glenwood Springs. By the time Holliday arrived in Glenwood Springs, his health was in severe decline. He hoped the miraculous mineral hot springs and vapor caves could provide a cure or at least offer some relief.

Holliday spent the last 57 days of his life in Glenwood Springs. On November 8, 1887, he was said to have asked for a glass of whiskey, which he thoroughly enjoyed. Then, looking down at his bare feet, Doc said, "This is funny," and died. Apparently, Doc never expected to die in bed, but rather with his boots on, succumbing to his end in a gunfight. (https://visitglenwood.com/history/doc-holliday-in-glenwood-springs/)

Glenwood Springs was one of the first places in the country to have electric lights (1887), with power supplied to the town even before many major cities like New York. Opened on July 4, 1888 (together with an opulent bathhouse complete with a men's only gambling casino), it remains the world's largest mineral hot springs pool. Measuring 405 feet long and 100 feet wide at the widest point, it contains 1,071,000 gallons of water, and is kept at about 90°F. The town leverages its geothermal resources to power local facilities, maintaining a reputation as a leading green and sustainable mountain community.



Figure 3.5.33 Glenwood Springs' main thermal pool. Photo courtesy of Glenwood Hot Springs Resort.

3.6 New Mexico, and Oklahoma

Santa Rosa Blue Hole, Guadalupe County, New Mexico

Santa Rosa Blue Hole is the largest spring in New Mexico, located in the town of Santa Rosa just off Interstate 40 (historic Route 66), roughly halfway between Albuquerque and Tucumcari. Below are several excerpts from the Santa Rosa Blue Hole webpage describing the spring.

Blue Hole appears in the midst of the desert like a great blue gem. (Agua Negra Chiquita) Once known as Blue Lake, it is one of seven sister lakes connected underground by a vast system of water.

Nomadic tribes, cowboys driving their dogies across the Pecos, and Americans going west on the Mother Road, Route 66, all sought respite here.

In 1932, Blue Hole became a National fish hatchery, morphed into the Blue Hole Recreation Area in the seventies, and more recently expanded to become the Blue Hole Dive and Conference Center.

Now it's a destination meeting site for everyone from brides to board chairs. Halfway between Albuquerque and Amarillo, it's within two hours drive of 80 percent of all New Mexicans.

Not just for drivers wanting to get out of the fast lane or divers eager to get on down, it's more than just a watering hole.



Figure 3.6.1 Santa Rosa Blue Hole spring in Santa Rosa, New Mexico. Ball-shaped buoys are anchored to the *cenote* bottom and provide orientation for divers. Courtesy of https://www.santarosabluehole.com/

Still, we never forget that the real sapphire is the incredible beauty of the Blue Hole itself.

The lake is unsurpassed for its clear, pure water. That's why we work so hard to protect it, ensuring, for example, that surface runoff won't impair its purity. Like scuba divers who drive ten hours to get here, we know it's the crystalline water that draws them in.

Visibility is an astonishing 100' due to the fact that the water completely renews itself every six hours—it's truly never the same lake twice.

What doesn't change is the temperature—a constant 62 degrees (Fahrenheit)*, ideal for storing a fine Cabernet or tossing in the kids on a hot summer's day. Within a few minutes of exiting the interchange, they'll literally be climbing the walls—the lovely rock surrounds of Blue Hole, that is—and jumping in.*

https://www.santarosabluehole.com/about-santa-rosa-blue-hole/

The 90 foot deep Blue Hole is an example of a *cenote*, a water-filled sinkhole in Spanish. It is just one of the roughly 190 gypsum and limestone sinkholes lying within a six-mile-wide subsidence feature known as the Santa Rosa sink in the Pecos River basin. Blue hole was formed by collapse of Santa Rosa sandstone into a void created by dissolution of the underlying San Andres and Artesia Group carbonates including gypsum. Inflow and outflow of groundwater from Blue Hole is a constant 3,000 gpm.

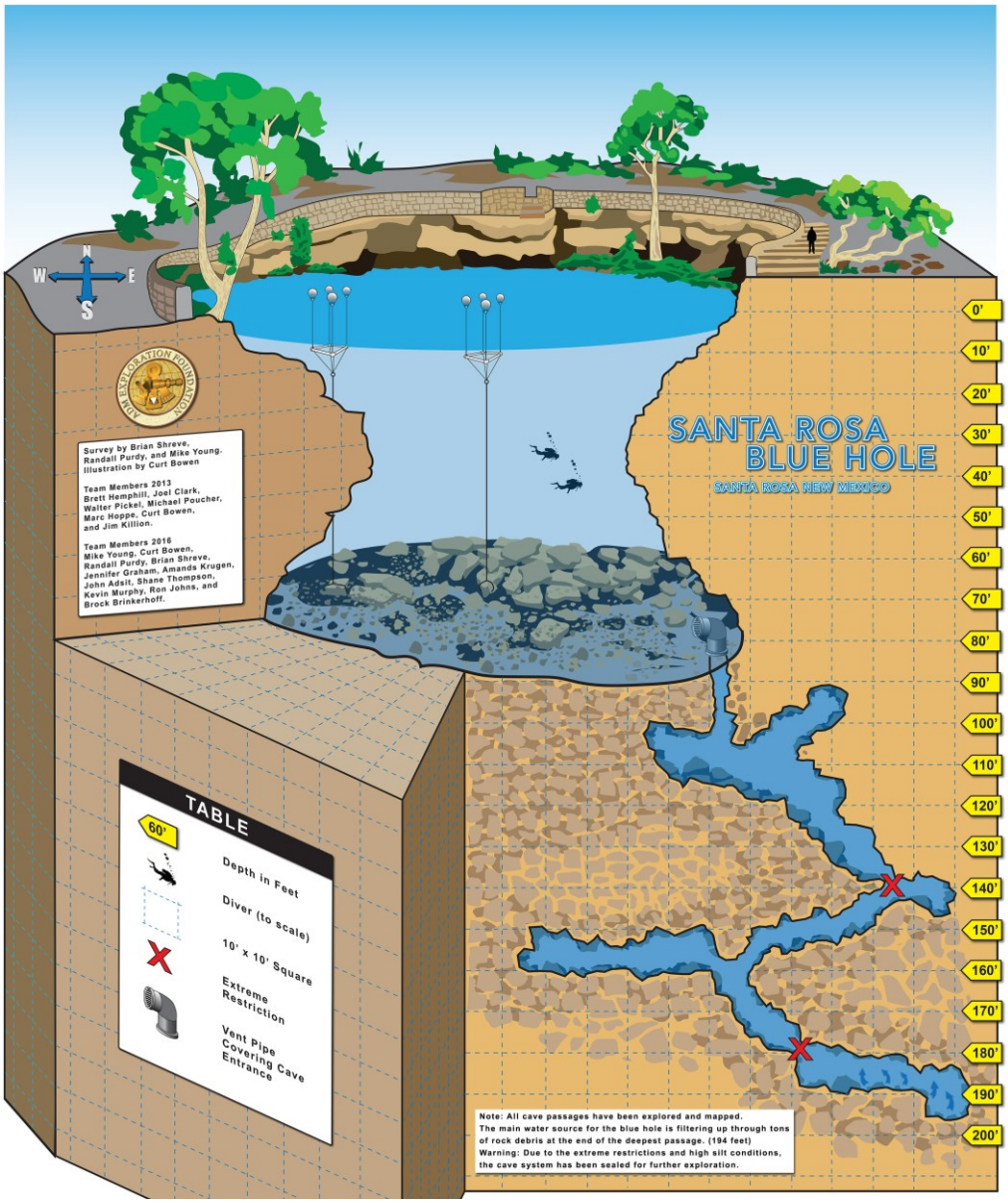


Figure 3.6.2 Divers assembled by ADM Exploration Foundation surveyed cave passages below the sinkhole on several occasions. The expeditions are documented on an excellent web site which features diving videos as well. Note at the bottom of the map reads: All cave passages have been explored and mapped. The main water source for the blue hole is filtering up through tons of rock debris at the end of the deepest passage. (194 feet) Warning: Due to the extreme restrictions and high silt conditions, the cave system has been sealed for further exploration. Available at http://www.admfoundation.org/projects/santarosa/santarosaexpedition.html



Figure 3.6.3 One of the roughly 190 sinkholes lying within a six-mile-wide subsidence feature known as the Santa Rosa sink in the Pecos River basin.



Figure 3.6.4 Freedivers in the Blue Hole's famously clear water. The combination of consistent water conditions (62°F year-round), 100-foot visibility, and easy access has made it a destination for divers from across New Mexico, Texas, Colorado, and the rest of the Four Corners region. Roughly 8,000 diving permits are issued every year. Photo courtesy of William Drumm. Detail description of diving highlights and conditions at the Santa Rosa Blue Hole, including more photographs, is provided by William Drumm at Santa Rosa Blue Hole, New Mexico: Diving & Visitor Guide

Byrds Mill Spring, Pontotoc County, Oklahoma

Byrds Mill Spring is the largest spring in Oklahoma. It is located near Fittstown in south-central Oklahoma and it is the primary water supply for the city of Ada. The water travels entirely via gravity, flowing from the spring site down to the storage and pumping stations on the south side of Ada. The system consists of two parallel 12-mile-long raw water lines: an older 24-inch pipeline, and a newer 30-inch line completed in 2017 to replace a pipe dating back to 1928.

Based on the combined daily spring discharge rate measurements (diversion for the Ada water supply plus the spring overflow) between December 1989 and September 2008, the mean is 18.5 cfs, the minimum is 4.6 cfs, and the maximum is 43 cfs (Mashburn et al., 2025).

The City of Ada is the county-seat of Pontotoc County and home to East Central University and headquarters of the Chickasaw Nation. The public water supply and wastewater system serves a population of 37,000 city customers and areas of several rural water districts. Byrd's Mill Spring provides 5 MGD on average for Ada's water supply, but 2060 demands project a need for nearly 10 MGD. The Arbuckle-Simpson Aquifer, which also has other demands from surrounding communities, feeds Byrd's Mill Spring which results in unreliable flows and potential water emergencies for the City of Ada.

In 2017, the City announced a feasibility study which would focus on water reuse alternatives and a distribution system to various potential customers in the surrounding area. As stated in the document announcing the study:

“The City of Ada's primary source of water is Byrd's Mill Spring fed by the Arbuckle-Simpson Aquifer. Byrd's Mill Spring typically flows at 10 MGD, and Ada's average demand is 5 MGD. Flow from Byrd's Mill Spring is highly variable, ranging from 3 MGD to 20 MGD. During droughts, Ada's demand has reached 10 MGD which can require supplemental water from one of three wells near Byrd's Mill Spring.

Many communities pull directly or indirectly from the same Arbuckle-Simpson Aquifer. Streams and springs fed by Arbuckle-Simpson Aquifer are featured within the Chickasaw National Recreation Area and the communities of Sulphur, Davis, Wynnewood, and Ardmore rely on the aquifer for water supply.



Figure 3.6.5 Historic photograph of Byrds Mill Spring, Ada's original water supply source. This photograph is part of the collection entitled: Oklahoma Historical Society Photograph Collection and was provided by the Oklahoma Historical Society to The Gateway to Oklahoma History, a digital repository hosted by the UNT Libraries. Byrd's Mill - The Gateway to Oklahoma History

Based on the Arbuckle-Simpson Aquifer hydrology study initiated in 2003, the Oklahoma Water Resources Board proposed a pumping limit of 0.2 acre feet per acre which is one-tenth of the current 2 acre-feet per acre limit. With this reduction, Byrd's Mill Spring is protected, but groundwater reserves that Ada relies on will no longer be available. Pulling 2 acre-feet per acre provides Ada with 18 MGD while pulling 0.2 acre feet/acre provides Ada less than 2 MGD (million gallons per day) from the Aquifer. With this new pumping limit, Ada would not have enough water supply for the average 5 MGD demand. Conservation, system improvements, and additional water rights may not satisfy future water demand in the area." (City of Ada, 2017).

The City has determined several long-term options, or a combination thereof, for water reuse: 1) Continue to pull water from Byrd's Mill Spring 2) Continue to pull water from groundwater sources 3) Water conservation 4) Enhanced recharge of Arbuckle-Simpson Aquifer 5) Use surface water from nearby Scissortail Lake 6) Purchase water from Oklahoma City Pipeline, and 7) Desalination of brackish groundwater.

Based on the results of the feasibility study, Ada selected water recycling as a primary infrastructure upgrade to reduce its baseline reliance on the aquifer. Rather than processing recycled water to drinking-water standards, the city opted to build a modernized reuse system that routes treated effluent directly to commercial and industrial demands. This recycled water will supply high-volume, non-potable needs such as industrial cooling, site-use, and commercial fire suppression systems.

The City is also exploring options for Enhanced Aquifer Recharge (EAR) in partnership with the Oka Water Institute at East Central University, the Chickasaw Nation, Oklahoma State University, and University of Oklahoma ("Oka" is the Chickasaw and Choctaw word for water; https://www.okainstitute.org/research).



Figure 3.6.6 Byrds Mill Spring's natural discharge pool is now covered for safety reasons. *Left* photo courtesy of USGS. *Right* photo by Richard R. Barron, Ada Evening News.

Byrds Mill Spring is situated in the northeastern region of the Eastern Arbuckle-Simpson aquifer (ASA). Geologically, this aquifer consists of two main formations: the Arbuckle Group and the Simpson Group. The Arbuckle Group, approximately 3,300 feet thick, is predominantly composed of karstified dolomite that stores water within cavities, solution channels, fractures, and intercrystalline pore spaces. (Fairchild et al., 1982; from Richins et al., 2026). Overlying the Arbuckle Group, the Simpson Group is composed of sandstone, shale, and limestone. Structurally, the Hunton Anticline, a geological feature formed during a period of uplift in the Pennsylvanian era, defines the Eastern ASA. This uplift resulted in fractures and folds, prominently shaping the aquifer's structure (Christenson et al., 2011, from Richins et al., 2026).

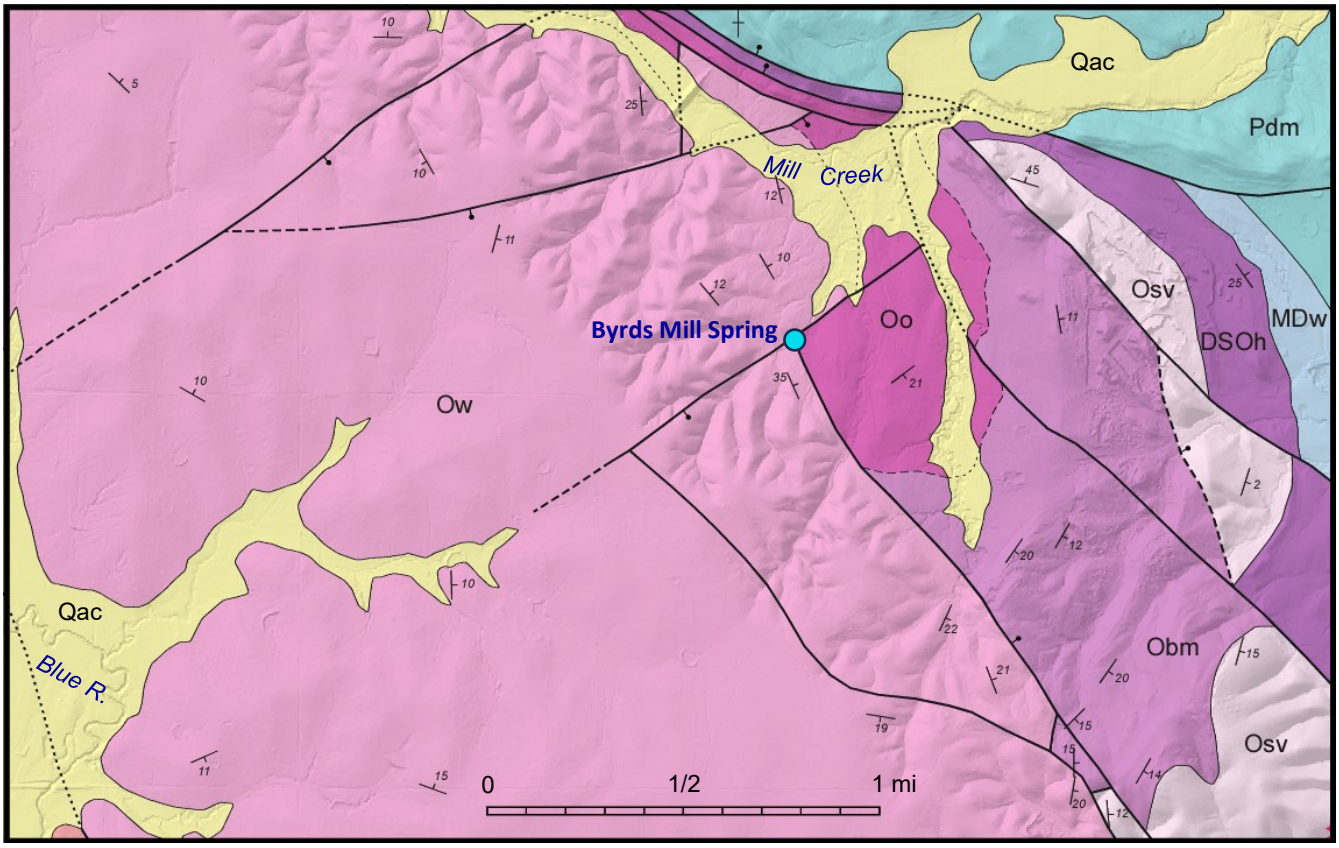


Figure 3.6.7 Geologic map of the Byrds Mill Spring area. Geologic units, from youngest to oldest: Qac—Quaternary alluvium and colluvium. Pdm—Pennsylvanian Desmoinesian Series formations: shales, siltstones, sandstones, coal, some limestones. PMs—Lower Pennsylvanian to Upper Mississippian Springer Group: shales and sandstones. MDw—Welden Limestone and Woodford Shale, undifferentiated (Lower Mississippian to Devonian). DSOh—Lower Devonian to Upper Ordovician Hunton Group: limestones, dolomites, marlstones, and calcareous shales. Osv—Ordovician Sylvan Shale and Viola Group: shale, claystone, limestone, and chert. Obm—Ordovician Bromide and McLish Formations: sandstone, limestone, and shale. Oo—Ordovician Oil Creek Formation of Simpson Group: sandstone, shale, and limestone. Ow—West Spring Creek Formation of Arbuckle Group (Lower Ordovician): limestone and dolomite. Modified from Lidke and Blome, 2017.

Byrds Mill Spring (BMS) discharges at fault contact between the West Spring Creek Formation carbonates of the Arbuckle Group and the underlying less permeable sediments of the Oil Creek Formation of the Simpson Group (Fig. 3.6.7). According to Richins (2020), and Richins et al. (2026), the spring, located at the intersection of two major faults, is an end point of two general flow systems shown in Fig. 3.6.8. This concept is based on the results of a comprehensive study including continuing monitoring of the spring discharge rate, water temperature and electrical conductivity (EC), analytical modeling using a tank siphon and mixing model, and Electrical Resistivity Imaging.

The study also identified two intermittent streams upstream of the spring, which serve as rapid flow paths during storms, consistent with the observed EC fluctuations. The authors emphasize how even small recharge features, such as these, can significantly impact water quality by overwhelming regional flow systems during extreme weather events. From a management perspective, the results indicate that intermittent losing streams upstream of BMS should be designated as protected recharge features, with land-use restrictions or buffer zones implemented to limit potential contaminant sources within the identified ~300 ft recharge corridor (Richins et al., 2026).

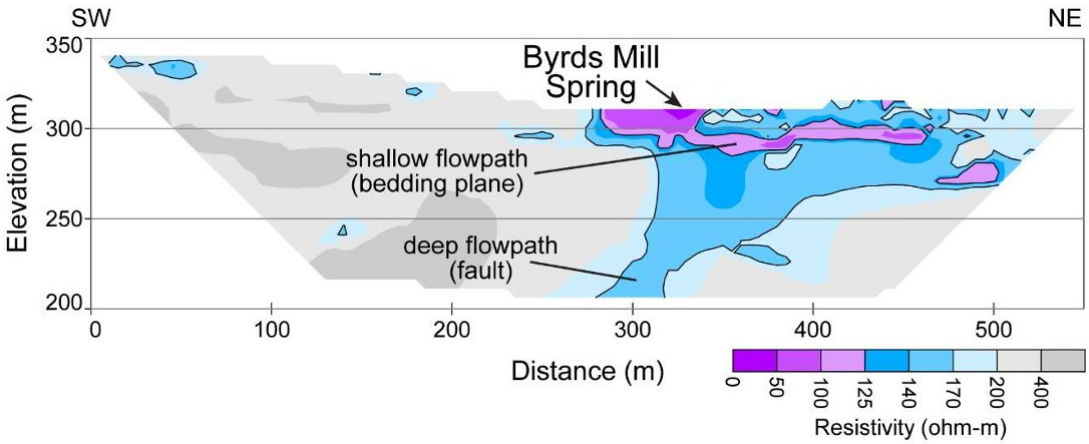


Figure 3.6.8 Electrical Resistivity Imaging of Byrds Mill Spring showing shallow and deep flow paths. The deep flow path is interpreted as a siphon. From Richins, 2020.



Figure 3.6.9 Looking away (downstream) from the Byrds Mill Spring. Photograph by Hayden A. Lockmiller, USGS. From Mashburn et al., 2025.

Continuous spring discharge hydrographs from the Antelope Spring, Byrds Mill Spring, and Sheep Creek Spring gages were analyzed by USGS for the highest discharge periods during 2015–16 to identify flow regimes of the Arbuckle-Simpson (ASA) karst aquifer. Linear regression equations (regression curves) for spring discharge and time were examined for breaks (inflection points) in the recession slope, which are indicative of a change from one flow regime to another within the karst continuum. Recession curves were classified as different flow regimes based on their recession coefficient (α) (Fig. 3.6.10)

If the recession coefficient is 0.18, then conduit flow is the predominant flow type. If the recession coefficient is 0.09, then flow type consists of fracture flow or a mixture of conduit flow and diffuse flow (with conduit dominant). If the recession coefficient is 0.02, then flow type consists of fracture flow or a mixture of conduit flow and diffuse flow (with diffuse dominant). If the recession coefficient is 0.008, the diffuse flow is the predominant

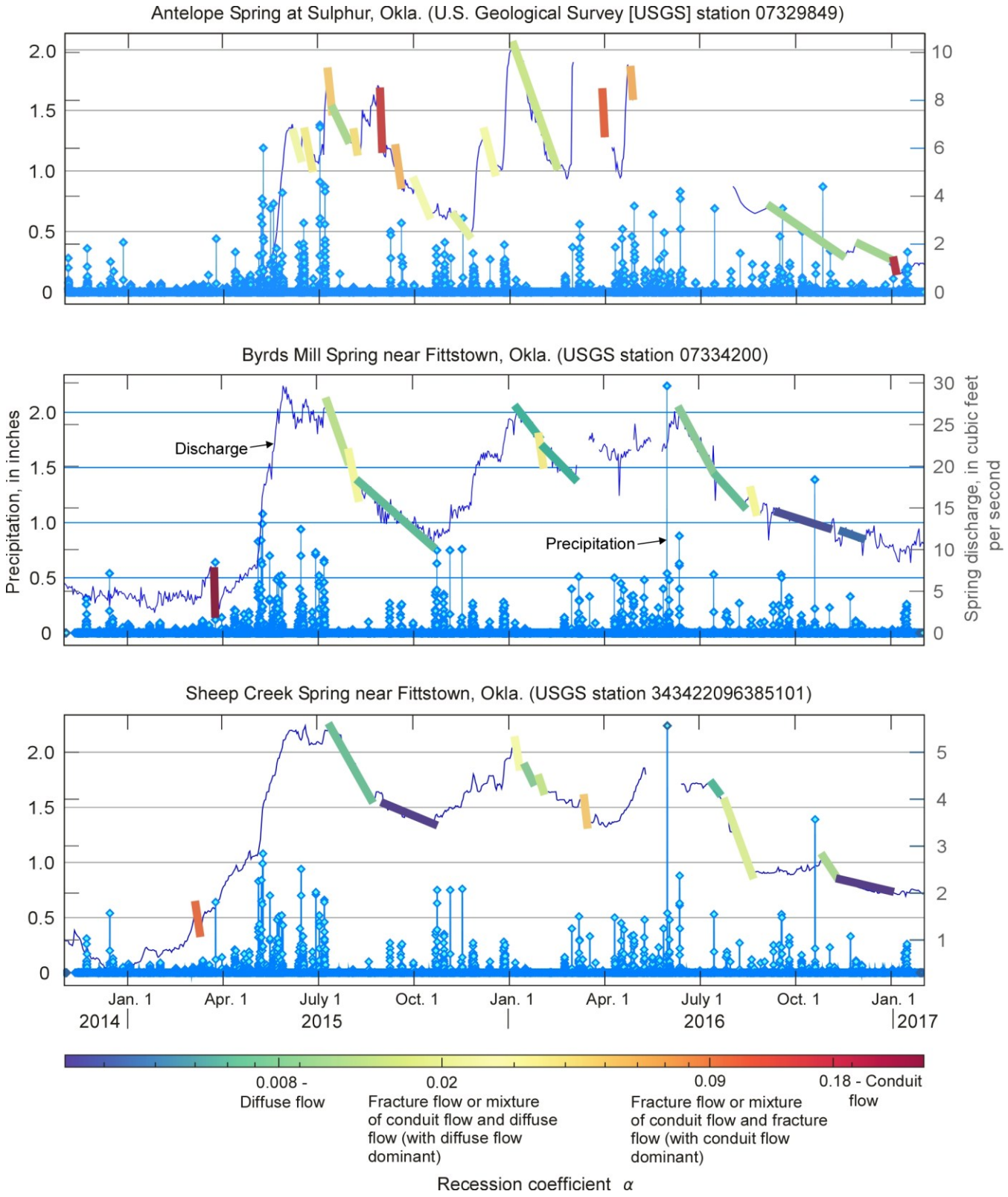


Figure 3.6.10. Discharge and spring recession curves classified by flow regime and precipitation for Antelope Spring, Byrds Mill Spring, and Sheep Creek Spring, south-central Oklahoma, 2014–17. Modified from Mashburn et al., 2025.

flow type (Milanović, 1976). As indicated by the recession curves at the highest discharge periods, Antelope Spring yielded a higher recession coefficient than during the other periods of recession and exhibited more conduit flow than diffuse flow, followed by greater portions of mixed and diffuse flow. This higher recession coefficient indicates that during periods of high-discharge recession, the groundwater system supplying Antelope Spring might include the upper layers of the aquifer which are dominated by fractures and conduits, with a middle and lower section of the aquifer dominated by fractures, primary porosity, or both (Mashburn et al., 2025).

As indicated by recession curves for the highest discharge periods for the Byrds Mill Spring and Sheep Creek Spring gages, recession coefficients are relatively lower and flow is mostly diffuse compared to Antelope Spring. These lower recession coefficients indicate that the groundwater system near Byrds Mill and Sheep Creek Springs, near the upper levels of the formation, is likely dominated by diffuse drainage through primary porosity (Mashburn et al., 2025).

Sheep Creek Spring, Pontotoc County, Oklahoma

Because the neighboring Byrds Mill Spring supplies municipal water to the city of Ada, Sheep Creek Spring is protected under strict groundwater regulations managed by the Oklahoma Water Resources Board (OWRB). To prevent over-pumping of the Arbuckle-Simpson Aquifer, laws limit groundwater extraction to a maximum annual yield. This specific legal threshold ensures that baseline discharge from springs like Sheep Creek never drops below the point of ecological collapse.

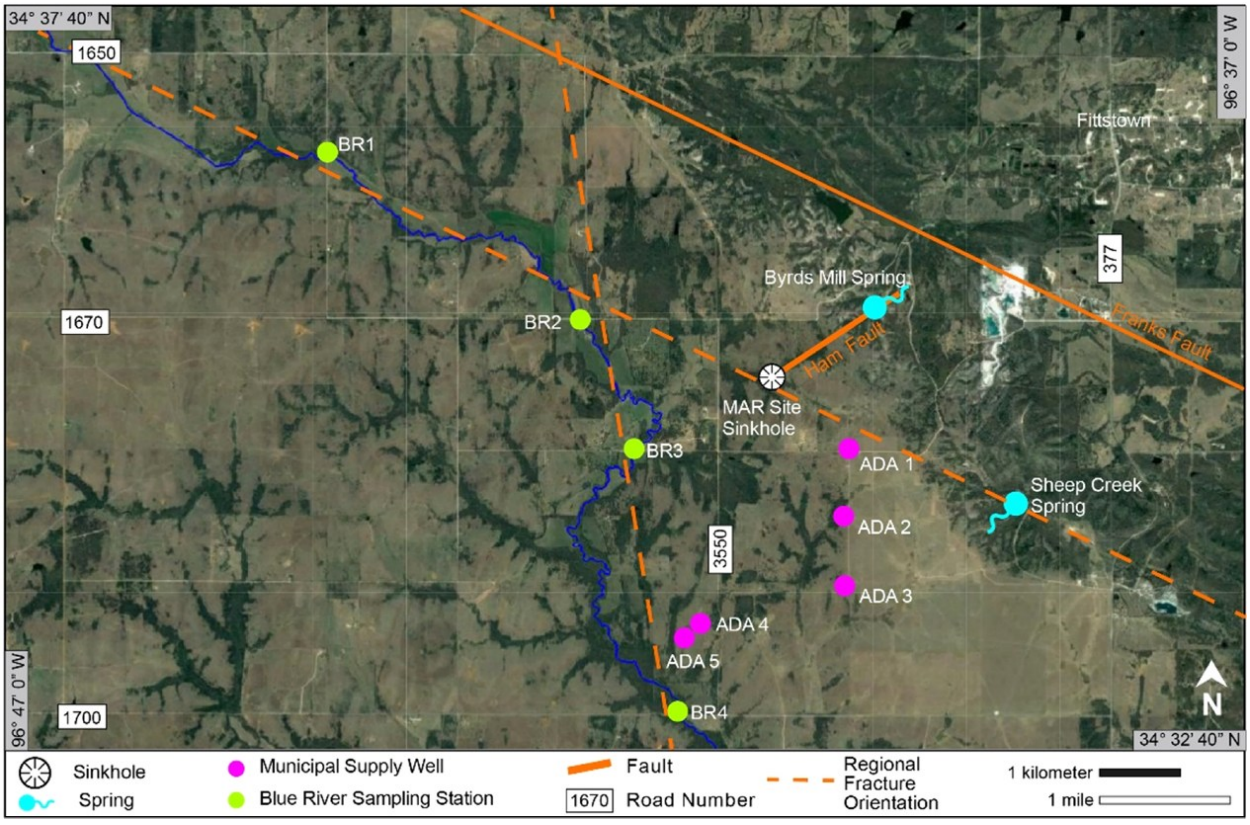


Figure 3.6.11 Map of the Sheep Spring and Byrds Mill Spring areas showing hydrologic features of interest. From Richins et al., 2026.

Sheep Creek Spring is not open to the public for recreation, swimming, or tourism. The spring orifice sits on restricted, privately held property within the jurisdictional boundaries of the Chickasaw Nation. Trespassing is strictly prohibited, meaning the physical site can only be accessed by landowners and authorized federal or tribal hydrologists conducting water monitoring tasks. The spring is a major field laboratory for groundwater science. The Chickasaw Nation actively partners with the USGS Oklahoma City Office to operate the continuous monitoring gage on-site.

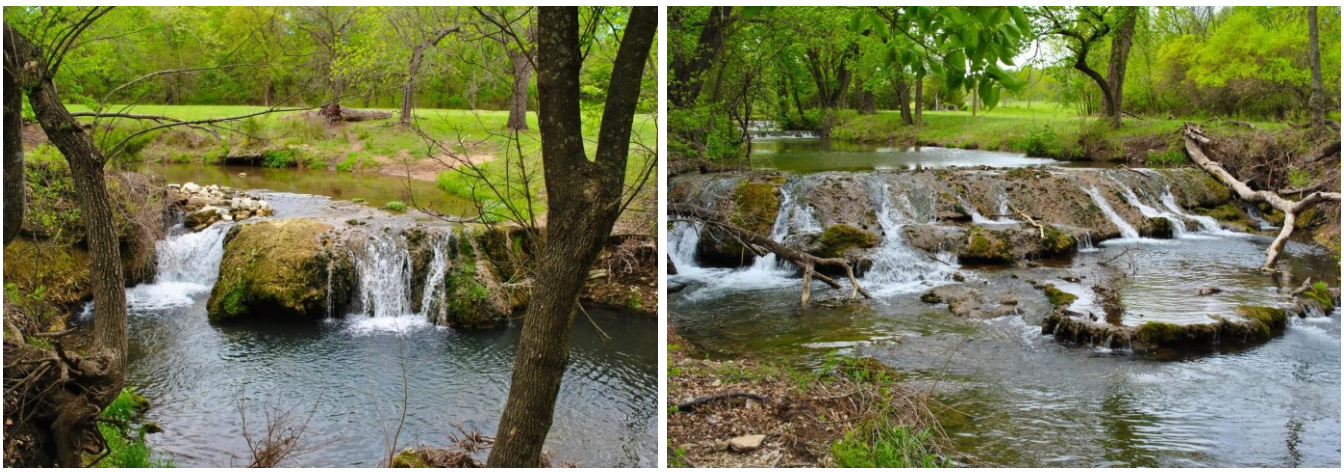


Figure 3.6.12 *Left*: Sheep Creek Spring. *Right*: Tufa deposits on Sheep Creek create numerous natural dams, including areas where the creek splits around islands and then rejoins in a double waterfall setting with large splash pools. Photographs courtesy of The Land Doctors https://landdoctors.com/property/spring-creek-ranch-pontotoc-oklahoma/8301/

Antelope Springs and Bufalo Springs, Chickasaw National Recreation Area, Murray County, Oklahoma

As explained by Hanson and Cates of USGS (1994), the Travertine District (Park) of the Chickasaw National Recreation Area, operated and maintained by the National Park Service, is near the City of Sulphur in south-central Oklahoma. The Park was established in 1902 because of its unique hydrologic setting, which includes Rock Creek, Travertine Creek, numerous mineralized and freshwater springs, and a dense cover of riparian vegetation. Since the turn of the 20th century several flowing artesian wells have been drilled within and adjacent to the Park which led to a substantial decline in discharge for many of these springs. The flow from the wells themselves has declined substantially such that, in 1990s, they were estimated to yield only about 10 percent of the total flow reported in 1939.

The principal geologic units in the area are the Arbuckle Group and the overlying Simpson Group. These rocks, of Upper Cambrian to Middle Ordovician age, are composed of dolomitic limestone, with some sandstones and shales in the Simpson Group. The two Groups are not readily distinct in some parts of the Chickasaw National Recreation Area because of the similarity of their rocks; thus, both water-bearing units are referred to frequently as the Arbuckle-Simpson aquifer. The aquifers are confined under the Park, but are unconfined east and south of the Park. Precipitation on the outcrop area of the Arbuckle aquifer northeast and east of the Park recharges the freshwater Antelope and Buffalo Springs near the east boundary of the Park. The source of water from mineralized springs located in the central part of the Park, and flowing wells within and north of the Park, is believed to be a mix of waters from rocks of the Arbuckle and Simpson Groups. The source of water from two highly mineralized springs, Bromide and Medicine, that ceased to flow in the early 1970s, is believed to be from the Simpson Group.

Antelope and Buffalo Springs provide most of the water in Travertine Creek, with a combined flow of about five million gallons per day during high flow years. From the beginning of the park's history, the water of Travertine Creek was very important. Records of the stream flow and condition were maintained since 1900s. The two springs have dried up over 20 times in the last 120 years, most significantly in the 1910s, 1930s, and 1950s. At times they were dry for only a couple of weeks, and other times, for over two years. In contrast, they ran for over twenty years straight from the mid-eighties to the early 2000s, though there were several times they slowed down during that period (https://www.nps.gov/chic/index.htm).

Antelope and Buffalo Springs are the most significant features of the former Platt National Park (a National Historic Landmark District), which now lies within Chickasaw National Recreation Area. The district is important for its association with the development of national park landscapes completed during the Great Depression and funded as part of Franklin Delano Roosevelt's "New Deal" programs.



Figure 3.6.13 Buffalo Spring, located in Chickasaw National Recreation Area; photo courtesy of USGS. The springs is surrounded by stone structures built by the Civilian Conservation Corps (CCC) in the 1930s.



Figure 3.6.14 Stone footbridge over Antelope Springs. Photo by K. Ruhnke, NPS.

3.7 Wisconsin, Minnesota, and Iowa

Large karst springs in Minnesota, Wisconsin, and Iowa are primarily in the so-called "Driftless Area" which also extends into the far northwestern corner of Illinois as shown in Fig. 3.7.1 Notable karst springs in Wisconsin are also present outside of the Driftless Area, whereas in Illinois, karst is more extensive in the southern and western parts of the state, outside of the Driftless Area, where all large karst springs are located.

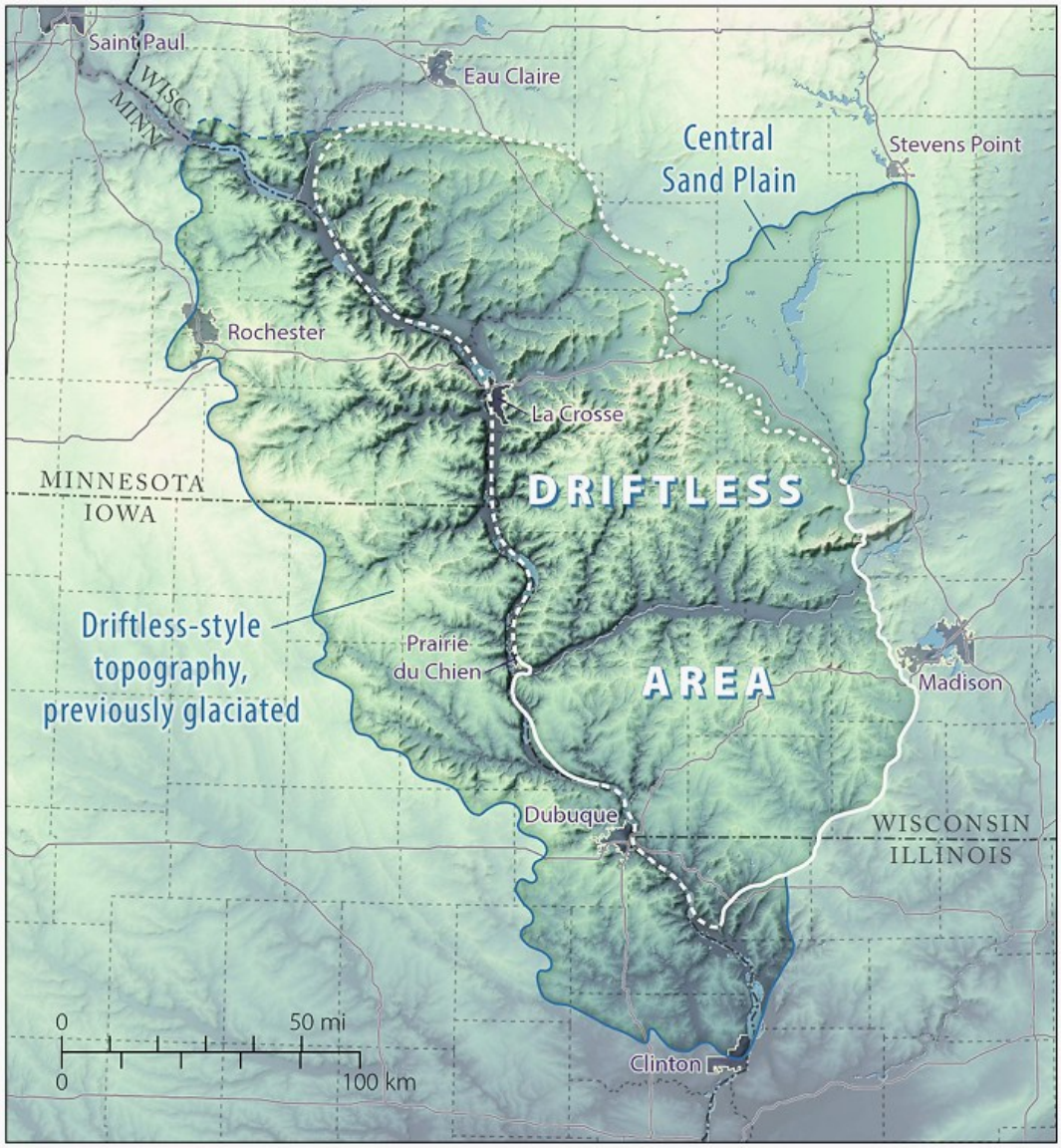


Figure 3.7.1 The Driftless Area: The extent of unglaciated and similar terrains in Wisconsin, Illinois. Iowa, and Minnesota. In southwestern Wisconsin and northwestern Illinois, the term "Driftless" refers to the area free of glacial sediment, or "drift", as it was once called. There is no evidence that the region has ever been covered by glaciers. The limit of glaciation is interpreted from ice-marginal landscape features including moraines, erratics, and glacial till. Dashed lines indicate uncertain or inferred boundaries. From Wisconsin Geological and Natural History Survey Educational Series 057, https://doi.org/10.54915/qapq2678. Highways and inset data courtesy of Natural Earth Data. Boundaries, hydrography, and elevation data courtesy of the U.S. Geological Survey.

As explained by Eagle Bluff Environmental Learning Center (https://eaglebluffmn.org/driftless/), while no formal boundaries define the Driftless Area, scientists agree it encompasses much of western Wisconsin and extreme northwestern Illinois. The areas in southeastern Minnesota, northeastern Iowa and (a very small area in) southwesternmost Wisconsin identified as having "Driftless Area" topography were glaciated, as documented by the identification of pre-Illinoian till in the subsurface. The relative antiquity of these glacial deposits has allowed them to be buried by younger sediment and for fluvial incision to become the predominant landscape-forming process. As a result, the landscape has primarily been formed by rivers cutting down into the nearly flat-lying sedimentary bedrock, forming valleys as deep as 400 feet from ridgetop to river level. While the area of the Central Sand Plain that represents the ancient bed of glacial Lake Wisconsin was almost certainly never glaciated, it has been excluded from the Driftless Area here because it lacks the characteristic fluvial incision commonly associated with the Driftless Area.

There have been many glaciations over the last 2.58 million years and although the last one did not reach this area, glaciers did reach SE Minnesota during earlier glaciations. The absence of glacial "drift" (moraines, erratics, and glacial till) in this area during the last glacial period is why part of Wisconsin and Illinois has been deemed the *drift*-less.

The failure of the most recent glaciers to penetrate the Driftless Area is responsible for the region's defining geomorphologic characteristics: steep hillsides, narrow valleys, and a land marked by unevenness. The glaciers were unable to blanket this landscape because they were blocked by the highlands of northwestern Wisconsin and simultaneously led away from the area through troughs created by Lake Superior and Lake Michigan. As the glaciers melted for the final time along the edges of the Mississippi some 13 thousand years ago, massive quantities of water released by melting ice sheets eroded the sedimentary rocks below. Known as "inverse topography", this left the land with flat plateaus above bluff tops, dissected by numerous streams and rivers, all flowing towards the Mississippi.

The widespread presence of soluble carbonates, in particular dolomites of St. Lawrence Formation (Upper Cambrian), Prairie du Chien Group (Lower Ordovician), and the dolomites and limestones of Galena Group (Middle to Late Ordovician), has led to numerous caves, sinkholes, and springs in the Driftless Area.



Figure 3.7.2 A small stream has eroded a limestone wall near Spring Valley, MN. Copyright © 2026 Eagle Bluff Environmental Learning Center. Available at https://eaglebluffmn.org/driftless/

3.7.1 Wisconsin

Highland Big Spring

Highland Big Spring is a must-visit-spring located off of Big Spring Road, about 2 miles west of Highland in Iowa County, in Wisconsin's Driftless Area. This hillslope spring cascades around thirty feet into the Big Springs Branch of the Blue River from a single orifice in Ordovician Prairie du Chien Group Limestone. The average flow of Highlands Big Spring is 3.2 cfs or 1,440 gallons per minute (Working Together To Collect Wisconsin Water Quantity Data). The spring has a fairly constant cold temperature, typically 42–50°F)

Highland Big Spring is a fishery area, where the Wisconsin DNR is monitoring the fish population in association with a beaver dam built nearby. The area is designated as Exceptional Resource Waters (ERW; "Surface waters that provide outstanding recreational opportunities, support valuable fisheries and wildlife habitat, have good water quality and are not significantly impacted by human activities.") in order to protect the Class 1 and Class 2 brown trout stream (https://dnr.wisconsin.gov/topic/Lands/FisheriesAreas/4058bigspring)

Highland Big Spring is located on public land owned by the Wisconsin DNR. There is a small, designated parking area and a short foot trail leading to the spring and waterfall cascades. While the area is open for public recreation, visitors should remain on the state-owned property to avoid trespassing on adjacent private farms.



Figure 3.7.3 Highland Big Spring, about 2 miles west of Highland in Iowa County is in Wisconsin's Driftless. Photo credit: DNR Water Use Section: Highland Big Spring Waterfall (Iowa County). Working Together To Collect Wisconsin Water Quantity Data.

Kelly Spring

Kelly Spring, near the town of River Falls in St. Croix County, is at the headwaters of Kelly Creek which flows into the main branch of the Kinnickinnic River. The springs and the creek are in Kelly Creek Preserve, a public 70-acre protected area known for its clear, cold, spring-fed stream, restored prairies, and around 2 miles of hiking trails. The preserve is managed by Kinnickinnic River Land Trust (https://www.kinniriver.org/nature-preserves-kinnickinnic-river-land-trust). Kelly Spring emerges from an outcrop of early Ordovician Prairie du Chien Group dolomite and has the average flow of 1.1 cfs or 495 gallons per minute (Working Together To Collect Wisconsin Water Quantity Data) which helps sustain cool temperatures for Brown and Brook Trout in Kelly Creek.



Figure 3.7.4 Kelly Spring outlet in winter, spring and summer. Photographs courtesy of Wisconsin DNR.

Paradise Springs

Paradise Springs is in southeastern Wisconsin's Kettle Moraine State Forest, approximately two miles northwest of the Town of Eagle, Waukesha County, and about four miles west of Jericho Creek. This spring complex supported a trout fishery until 2015 when the dam which created the pond failed and most of the water depth and fish were flushed downstream. The spring flows at about 500 gallons per minute or 1.1 cubic feet per second (WDNR, 2015), placing it in the protected category according to the groundwater quantity legislation, 2003 Wisconsin Act 310 (Hart et al., 2008). Paradise Springs is about 5 feet deep and maintains a temperature of about 47°F year-round.

Paradise Spring likely captures most of its water from the underlying Silurian dolomite aquifer, a conclusion reached by Gittings (2005) for the springs in the Mukwonago River Watershed where Paradise Spring is located. Land surface around the spring is covered by unconsolidated glacial deposits (sand and gravel), which together with the bedrock form the shallow aquifer system.

Since the late 1800s, people have enjoyed the pure water of Paradise Springs. Early owners called it Minnehaha and Eagle Rock Springs. The last bottling plant at the spring closed in the late 1960s. Spring water was pumped from the spring house to the bottling plant and sold by various companies under the names of Natural Spring Water and Lullaby Baby Drinking Water.

Paradise Springs is now all but a forgotten roadside attraction. The ruins of the spring house still attract urban explorers, but this once popular strolling park is empty even on busy holiday weekends (Kenneth Casper, https://wisconsin-explorer.blogspot.com/2016/07/exploring-paradise-springs.html.) WDNR (Wisconsin Department of Natural Resources (WDNR, 2015) has published a pamphlet describing the history of Paradise Springs and featuring a self-guided nature trail.



Figure 3.7.5 Ruins of the spring house at Paradise Springs. *Left photo* courtesy of Kenneth Casper; *right photo* by Michael Leland; available at https://www.flickr.com/photos/milelandpix/15070241184/. Under Creative Commons license https://creativecommons.org/licenses/by-nc/2.0/deed.en

Three Springs, 2 miles outside of downtown Sister Bay in Door County, is a headwater of Three Springs Creek that flows into North Bay and eventually into Lake Michigan. Much of the area surrounding the springs is managed by the Door County Land Trust as part of the Harold C. Wilson Three Springs Nature Preserve, a designated Wetland of International Importance (https://doorcountypulse.com/hike-this-three-springs-nature-preserve/.) The springs provide a consistent source of cold, hard water that supports rare species, most notably the federally endangered Hine's emerald dragonfly, which relies on these specific wetland conditions.

Glaciation has removed much of the soil, leaving the Silurian dolomite bedrock exposed in many places or covered by only thin soil or drift.



Figure 3.7.6 Pond fed by Three Springs, located on the Door Peninsula in Door County. Springs issue from shallow karst aquifer developed in Silurian dolomite. The average flow rate of Three Springs was 1.9 cfs or 855 gallons per minute between 2015 and 2022. (Working Together To Collect Wisconsin Water Quantity Data)

Powell Spring

Powell Spring, located near Green Lake in Green Lake County, discharges at approximately 4.5 cubic feet per second (2,000 gpm) during the spring season. It serves as the primary headwater for Powel Creek, a 10-foot-wide stream discharging directly from the fractured Middle Ordovician Platteville-Galena dolomite. Powell Creek meanders downstream about a half mile until it merges with White Creek, which winds another mile or so (passing under Hwy. A) until it empties into Green Lake at its southeast corner. The spring flows nearly continuously year-round but fluctuates somewhat during wet or dry periods. During times of snowmelt or spring rain it often reaches its maximum flow, and on rare occasions it may stop flowing for a time until the groundwater is recharged by additional precipitation. (https://greenlakeconservancy.org/ps-naturalhistory/)

Green Lake Conservancy has stewardship over this fragile ecological resource and believes it is vitally important to protect the Powell Spring in perpetuity. It is for that reason that The Conservancy joined other environmental groups including The Green Lake Association (GLA), the Green Lake Conservancy (GLC), and the Green Lake Sanitary District (GLSD) in successfully appealing a conditional use permit for the Skunk Hollow Mine, which was awarded to Kopplin & Kinas Co., Inc. This permit would have allowed construction of a proposed 40-acre limestone quarry, roughly one farm field away from Powell Spring and the adjacent Mitchell Glen, another outstanding and ecologically sensitive natural feature in the area. (https://www.riponpress.com/news/)



Figure 3.7.7 *Left*: Powell Spring pours out from the side of a dolomite outcrop located just below the cabin shown in this image. Courtesy of Commonwealth Ripon Press. *Right*: Powel Creek downgradient of the spring source. Courtesy of Green Lake Conservancy; https://greenlakeconservancy.org/ps-naturalhistory/

Mitchell Glen is a 12.2-acre protected site in the town of Brooklyn in Green Lake County. It is best known for its 40-foot waterfall within a dolomite and sandstone gorge, as well as its unique flora and fauna. Mitchell Glenn is considered a "crown jewel" of the Green Lake Conservancy.

The glen was formed approximately 12,000 years ago by post-glacial meltwater that cut through Platteville-Galena dolomite and St. Peter sandstone, resulting in steep, timbered cliffs. The ravine features a spring-fed stream called Glen Creek and sits at an elevation that drops roughly 100 feet from its crest to its base. Several small springs that start the creek emerge at the contact of dolomite with the underlying sandstone. The site is situated along the Platteville-Galena escarpment, a significant geologic feature in southern Wisconsin and Northern Illinois, formed by the differential erosion of Ordovician-age carbonate rock. It is often associated with the region's hilly terrain, karst topography, and historically rich lead and zinc mining areas. It is one of several parallel escarpments in the Upper Midwest, including the more famous Niagara Escarpment. Mitchell Glen can be visited only by guided tours.



Mitchell Glenn, as well as Powel Spring nearby, are now threatened by the future operation of a limestone quarry located a bare mile away, as illustrated by the following excerpts from an article by Roger Mathews entitled *County K Quarry receives green light*, published in Commonwealth Rippon Press on June 19, 2024: *A conditional use permit for the proposed limestone quarry known as the "County K Quarry" was approved by the Green Lake County Land Use Planning and Zoning Committee last week Friday. The approval comes after a similar proposal for a quarry by Kopplin & Kinas, called the "Skunk Hollow Quarry," was initially approved by the Green Lake County Land Use Planning and Zoning Committee and then rejected by the Green Lake Board of Appeals (BOA) in December of 2022. A marathon session of public comments preceded the decision to deny the conditional use permit for that project*.

Figure 3.7.8 Mitchel Glenn waterfall. Photo courtesy of Green Lake Sanitary District.

Castle Rock Spring

Castle Rock Spring in the Driftless Area of Wisconsin is near town of Fennimore in Grant County, just north of Church Road on County Highway Q. It discharges at the base of Early Ordovician Prairie du Chien Group dolomites, which overlie less permeable Cambrian sandstone layers—a geological setting common for regional springs in southwestern Wisconsin. Castle Rock Spring is one of the largest in the state, with an average discharge rate of approximately 3,000 gallons per minute (gpm). It forms headwaters of Castle Rock Creek (also referred to as Fennimore Fork of the Blue River), the main tributary of the Blue River. The spring maintains a relatively constant year-round temperature of about 48–51°F (9–10°C), which is why it provides a critical cold-water habitat for brown trout.

A Facebook post at https://www.facebook.com/groups/558782617615194/posts/2933308786829220/ includes the following historic description of the spring:

1) Fennimore Times, June 1, 1904 PICTURESQUE CASTLE ROCK AND GAS HOLLOW (Editor Henry Roethe)

"We drank out of the spring, the largest in the county, whose waters, clear as the sky above, come gushing out in such large volume that a city of ten thousand inhabitants could be supplied with it. A little farther along the road is the Norwegian schoolhouse, and ten rods or so back of it is the big cave, which no one yet has fully explored. We found the entrance under a large rock, but did not venture in. A little distance farther along, on the town line between Hickory Grove and Castle Rock, is the site of the new parsonage of the Norwegian Lutheran church. The rock has been hauled and building operations will soon commence. Then you cross an iron bridge over Blue River, and soon are in the village of Castle Rock"





Figure 3.7.9 *Left*: Castle Rock Spring, headwaters of the Blue River (https://www.wisconsinrivertrips.com/segments/castle-rock-creek). *Right*: Diver at the entrance to the underwater cave of Castle Rock Spring. The cave extends approximately 270 feet back into the rock with a maximum depth of only 12 feet. Despite its short length, it is notoriously difficult to navigate due to its extremely tight restrictions. Because of the narrow openings, divers must use "no-mount" techniques, pushing their oxygen tanks in front of them rather than wearing them on their backs. Photo Copyright ©2011 Superior Dive Training. Available at https://web.archive.org/web/20180407231551/http://superiordivetraining.com/videos-2



A major indicator of Castle Rock Spring's karstic nature is Castle Rock Cave, located just a quarter mile north of the spring. Historically described as having "wonderful features," the cave's presence confirms the existence of the subterranean solution cavities typical of karst systems. The cave is on private property with restricted access.

Figure 3.7.10 Photograph showing geologist R.H. Wood seated near the entrance to a cave at Castle Rock, Wisconsin, ca 1920. University of Wisconsin-Madison Libraries. Geological and Natural History Survey: photograph 1901, album p245.

3.7.2 Minnesota

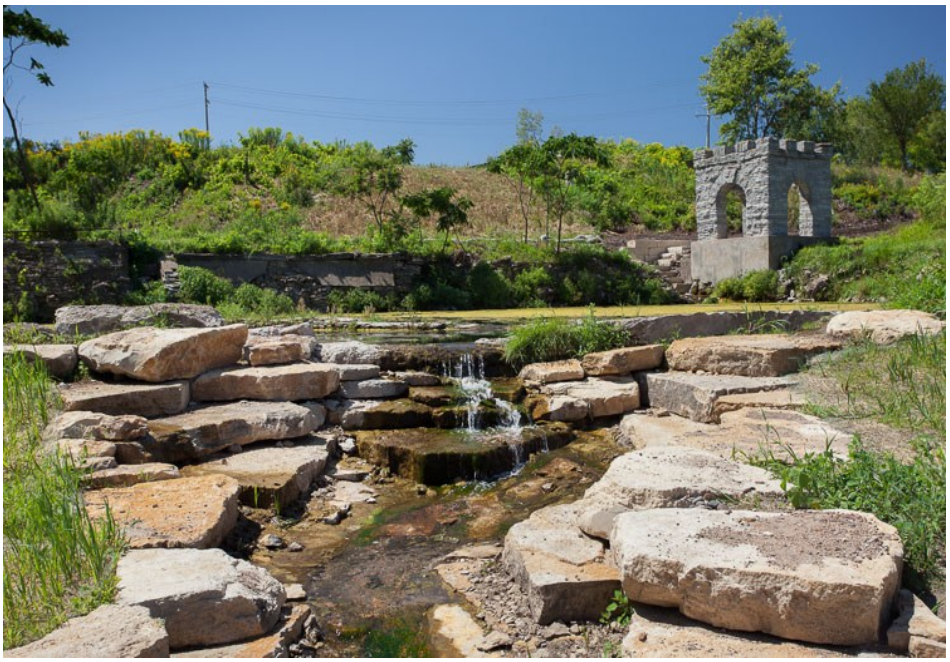
Coldwater Spring

The site of the Coldwater Spring (*Mni Owe Sni*) in the southeast corner of Minneapolis north of Fort Snelling was made part of the National Park Service's Mississippi National River and Recreation Area in 2012. It is a sacred area for the Dakota and other Tribes. In 2023, it was officially designated a Traditional Cultural Place listed in the National Register of Historic Places, recognizing its deep and ongoing significance to Dakota people. Mni Owe Sni is not merely a historical site—it continues to be an active and vital part of Dakota life. Mni Owe Sni encompasses the spring itself and the surrounding landscape down to the riverbank.

Mni Owe Sni's issues from Ordovician-age Platteville Limestone and flows down to its confluence with the Mississippi River. Mni Owe Sni lands are managed in partnership by the National Park Service (NPS), federally recognized Tribes, and agency landowners. The area is steeped in history, but its National Park Service chapter began in 2010 when the NPS started ecological restoration of what was an abandoned US Bureau of Mines administrative campus.

Mni Owe Sni has been and continues to be a place of peace, ceremony, and gathering, where Dakota, Anishinaabe, Iowa, and Ho-Chunk Tribes, met to nurture relationships, and participate in cultural practices. The arrival of European settlers prompted the US government to station soldiers here, choosing this location, in part, because of the spring's clean and abundant water. (NPS; https://www.nps.gov/places/mni-owe-sni.htm)

Originally, the area was an important crossroads for Native Americans and traders using both the Mississippi and Minnesota Rivers for commerce and travel. During the construction of Fort Snelling, United States soldiers camped at Coldwater Spring, making it the first American settlement in Minnesota. After the completion of Fort Snelling, the spring continued to provide water to the fort and the later Upper Post. The fort relied on horse-drawn water wagons and eventually railcars to deliver water from Coldwater Spring until the 1870s. (https://mntrips.com/walk-coldwater-spring-national-park-site/).



When the fort was expanded in the late 1870s, it required a more efficient water supply system and, in 1880, the Army established a formal waterworks at the spring. Coldwater served as the water source for the Army until the 1920s when the City of Saint Paul took over the water supply for the fort.

Figure 3.7.11 Coldwater Spring (*Mni Owe Sni*). Photo courtesy of NPS.

Fountain Big Spring

Fountain Big Spring is about a mile northwest of Fountain and the largest of a complex of four typical conduit springs. Many of the sinkholes in and around the town of Fountain drain to this spring complex within hours. All four springs emerge near the base of the Cummingsville Formation in the Galena Group just above the contract with Decorah Shale. The spring complex acts as an exit point for groundwater that can travel at speeds of 1–3 miles per day in the Galena-Spillville karst, causing the spring to respond quickly to rain or snowmelt, often turning turbid.

Fountain Big Spring used to act as the city's water supply. A significant portion of the water that previously flowed through it was diverted over 40 years ago by mining operations at the nearby Big Spring quarry in Harmony and now discharges in the quarry.



Figure 3.7.12 Fountain Big Spring. Photographs courtesy of John Barry.

Odesa Spring

Odesa Spring is the largest spring in Minnesota; its discharge rate is estimated to vary from 20,000 to 90,000 gallons per minute (Alexander et al., 1995). It is located at the base of a limestone cliff in the lower Prosser Formation limestones of the Ordovician Galena Group along the Upper Iowa River near the Minnesota/Iowa border. The spring is the primary discharge point for the York Blind Valley in southern Fillmore County which is the largest blind valley in Minnesota. Blind valleys are landscape features that abruptly end with surface water disappearing into the subsurface. The York Blind Valley accepts 7,520 acres of Canfield Creek's surface watershed, draining into the longest subsurface conduit system documented by dye tracing in Minnesota. Water that disappears

in the York Blind Valley resurges approximately 10.5 miles away at Odessa Spring near the town of Granger in Filmore County as shown by groundwater dye-traces conducted by the Minnesota DNR (Larsen et al., 2018).



Figure 3.7.13 Minnesota Caving Club members in Grand Central Station within Holy Grail Cave. A dye trace was conducted in 2009 documenting a connection from cave drips to Odessa Spring. Holy Grail Cave is located approximately midway between York Blind Valley and Odessa Spring. Photo courtesy of Martin Larsen.

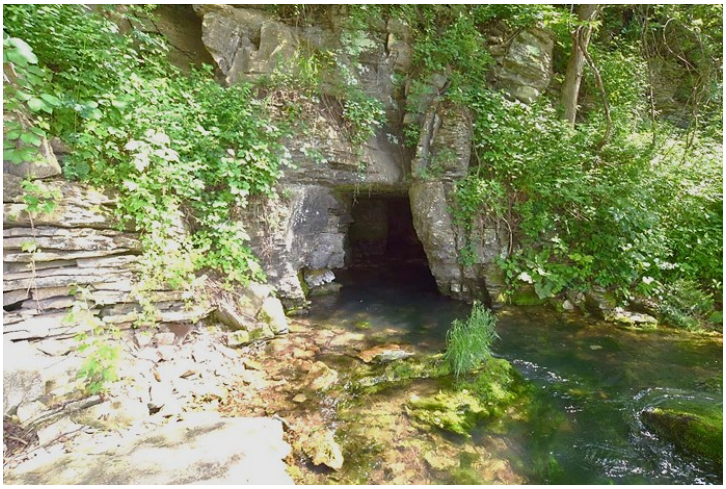


Figure 3.7.14 The largest spring in Minnesota, Odessa Spring, flows into the Upper Iowa River just before the river enters Iowa. Photo courtesy of Martin Larsen.



Figure 3.7.15 Odesa Spring West (on the left) and East. Photo courtesy of John Barry.

South Branch Root River Springs

Seven Springs, Crayfish Springs, and **Saxifrage Springs** are significant resurgence points for the South Branch of the Root River in Fillmore County, the largest sinking stream in Minnesota which sinks through Mystery Cave (Fig. 3.7.16). These springs emerge from the Galena carbonates aquifer system, acting as primary groundwater outlets that feed the river's flow in this karstic region.

Moth and Grabau Springs are the resurgences of sinking South Branch of the Root River under high flow conditions, and a large springshed including other caves. They issue from Ordovician Cummingsville Formation at the base of the Galena Group and are the headwaters for Forestville Creek.

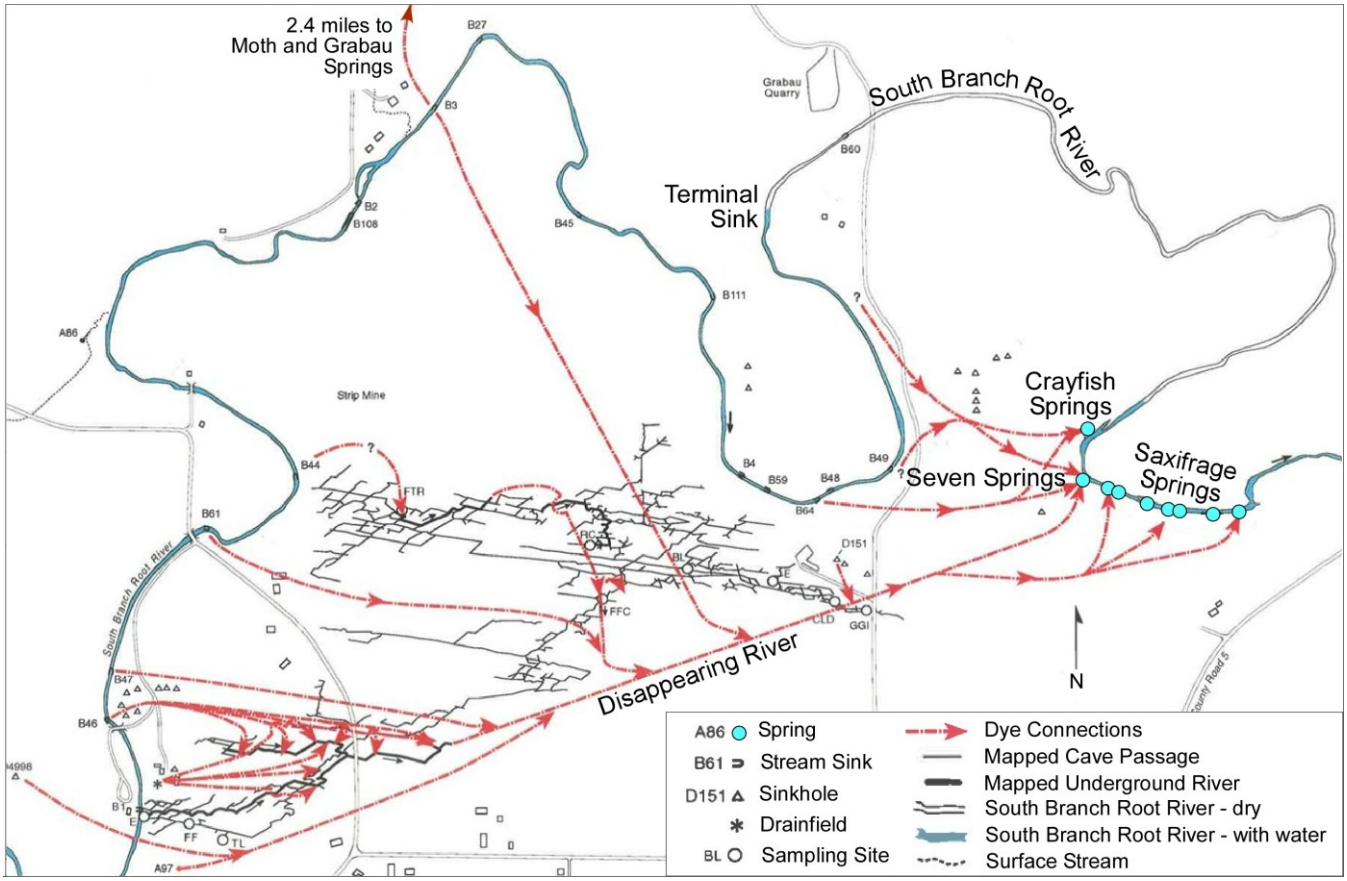


Figure 3.7.16 The Underground Rivers of Mystery Cave; Moderate Flow, Root River < 20-25 cfs. Modified from Calvin, 1996. Available at https://conservancy.umn.edu/items/7a596e71-7a92-4ac3-982d-8eda756ccb0f

Mystery Cave functions as a meander cutoff for the South Branch of the Root River. Surface water in the South Branch drains underground through parts of Mystery Cave, short-circuiting the longer surface course (Fig. 3.7.16). The Mystery I entrance is adjacent to the South Branch of the Root River on its south side (at the Mystery I entrance the river flows north and locally the entrance is on the east side of the river). Water sinks at discrete locations which start near Mystery I and occur for several miles downstream. The sink points are vertical joints that have been solutionally enlarged and filled with sediment. The sink points have collective capacity to accept surface water. When the flow in the river exceeds that capacity, it continues to flow through the entire surface reach of the South Branch near the cave. As the flow in the river recedes toward and then below the capacity of the sink points, a terminal sink develops on the South Branch. This terminal sink migrates upstream as the flow in the river decreases.

During all but the wettest years several miles of the South Branch are dry during much of the summer and fall. The permanent flow in the South Branch resumes at Seven Springs about 1.5 miles east-northeast of the Mystery I entrance and about 0.5 miles east-northeast of the Mystery II entrance (Jameson and Alexander, 1994).

Mystery Cave is primarily in the dolomite of the Stewartville Formation, and in the limestone, dolomite, and shale of the overlying Dubuque Limestone. Some of the highest parts of the cave are near-surface breakdown rooms in dolomite and shale of the lower of the Maquoketa Formation (Palmer and Palmer, 1993). Many passages appear to have originated as vertical fissures at or near the Dubuque/Stewartville contact, then enlarged upward by a combination of collapse with dissolution and downstream transport of the debris. Most of the present void space of passages is in the Dubuque Formation (Jameson and Alexander, 1994).



Figure 3.7.17 A highlight of Mystery Cave is its bluish Turquoise Lake, colored by dissolved calcite. Mystery Cave is the largest cave in Minnesota and is in Fillmore County in the heart of southeast Minnesota's karst land. It is a joint-controlled network maze (Milske et al., 1983). A network maze consists of a net of intersecting passages with closed loops that formed contemporaneously (Palmer, 1975; 1991). Mystery Cave has two entrances (Mystery I and II) and is owned and managed by the Minnesota Department of Natural Resources as part of Forestville State Park. The State Park staff conduct tours through the commercial parts of the cave from May to September. The blue color of Turquoise Lake is caused by Rayleigh scattering by calcite molecules (CaCO3) and by complexes of up to a few thousand molecules in the water. Molecules and particles much smaller than the wavelength of light selectively scatter the blue wavelengths relative t9 the red wavelengths. When a light beam shines into the water, the calcite molecules selectively scatter the blue light toward the observer. This is the same process that causes the sky to appear blue, except in the sky nitrogen and oxygen molecules and very small aerosols produce the scattering (Jameson and Alexander, 1994). Photo courtesy of Explore Minnesota. https://conservationcorps.org/updates-stories/forestville-mystery-cave/.

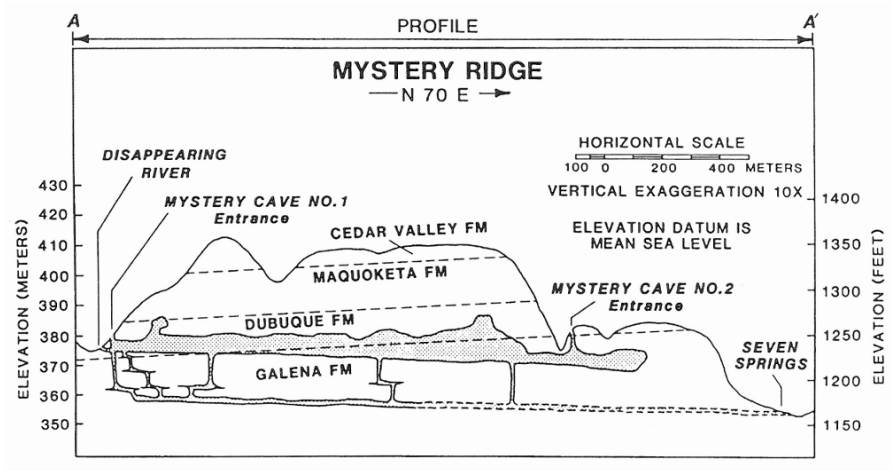


Figure 3.7.18 Profile of Mystery Cave (from Milske, 1982).



Figure 3.7.19 The largest of **Seven Springs**, one of three spring clusters that drain Mystery Cave (see map in Fig 3.7.16). When surveyed in September of 1992, nine springs were identified at Seven Springs. Many of the springs are at the base of bedrock cliffs. Rock falls from cliff collapses periodically block individual spring orifices or divert water so that flow is from two or more orifices. Most of the springs issue from joints or bed-joint intercepts, but some issue from rubble piles so that it is uncertain whether flow is actually from a single solutionally-enlarged fracture or several (Jameson and Alexander, 1994). Photo courtesy of Jake Stacken.



Figure 3.7.20 **Crayfish Springs** are ephemeral and dry up when the terminal sink of the South Branch retreats upstream of the riverbed immediately north of the Mystery II entrance. Photo courtesy of Jake Stacken.



Figure 3.7.21 **Saxifrage Springs** are volumetrically the largest of the Mystery Cave springs much of the year but may also dry up when the terminal sink has retreated above the bridge at the east end of the Mystery I driveway. Photo courtesy of John Barry.

The Nature Conservancy preserved 337 acres in Fillmore County to protect Seven Springs and a trout stream connected to Forestville-Mystery Cave State Park. The property, which also includes bluff and native prairie, is now owned and managed by Minnesota DNR as Seven Springs Wildlife Management Area (WMA). Seven Springs WMA is open to the public for hiking, photography and bird-watching. The Minnesota State Constitution requires property purchased with Outdoor Heritage Fund dollars to be open to the public taking of fish and game during the open season (https://www.nature.org/en-us/get-involved/how-to-help/places-we-protect/seven-springs-wma/).



Figure 3.7.22 **Moth Spring** is the resurgence of South Branch of the Root River under high flow conditions. Together with Grabau springs, it constitutes the headwaters for Forestville Creek. Photo courtesy of Martin Larsen.



Figure 3.7.23 **Stagecoach Spring** is a significant karst spring located in Fillmore County in southeastern Minnesota, 4.5 miles northwest of Preston. It is the primary source of Watson Creek, a designated trout stream, providing a vital input of cold groundwater, which is essential for maintaining the water temperatures required by the brook, brown, and rainbow trout that populate the 16.9 miles of Watson Creek. Stagecoach Spring is not located within a state park, making public access more difficult or potentially restricted by private property. *Left photo* courtesy of Martin Larsen. *Right photo* courtesy of John Barry.



Figure 3.7.24 **Canfield Big Spring**. Part of the spring's water comes from Goliath's Cave in the Cherry Grove Blind Valley Scientific and Natural Area, approximately five miles away. It is accessible via the Big Spring Trail within Forestville-Mystery Cave State Park. Photo courtesy of Jake Stacken.
"*Adventurers looking for a bit of an outing should consider a trek to Canfield Big Spring, the headwater of Canfield Creek in Forestville/Mystery Cave State Park, which requires a four-mile round-trip hike on the park's Big Spring trail. The hardwood forest in the valley is lovely any time of year, but it is particularly beautiful when it displays the oranges, reds, and yellows of autumn or the vivid wildflowers of early spring. When you are within a few hundred yards of Canfield Big Spring, the trail goes through Canfield Creek. You will gain a new appreciation for the term "cold-water stream" as you wade through its spring-fed water! After crossing the stream, it is a quick hike to the Big Spring, which emanates from a cave at the base of a limestone cliff.*
https://www.dnr.state.mn.us/mcvmagazine/issues/2022/sep-oct/springs.html

Boiling Springs

As explained by Greg Brick in his *Encyclopedia of Noteworthy Minnesota Springs by County* (Brick, 2017), Schwartz (1936), a renowned Director of the Minnesota Geological Survey, devotes one paragraph to Boiling Springs, once a famous tourist attraction. This information is repeated verbatim by Schwartz and Thiel (1954) in Minnesota Geological Survey Bulletin 37 (from Brick, 2017):

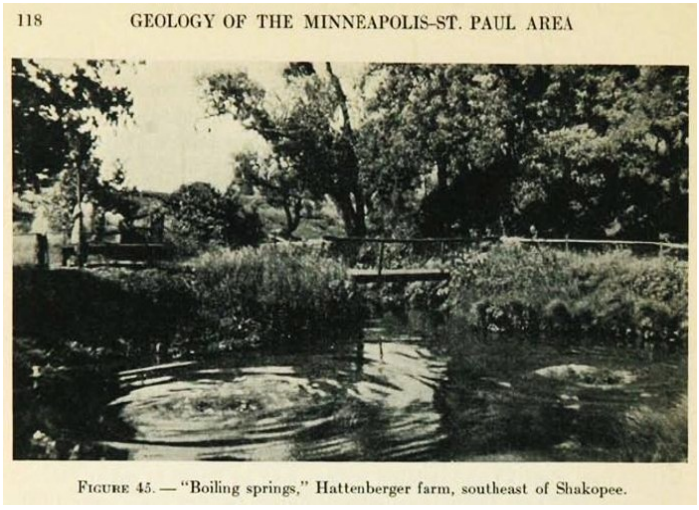

118 GEOLOGY OF THE MINNEAPOLIS-ST. PAUL AREA

FIGURE 45. — "Boiling springs," Hattenberger farm, southeast of Shakopee.

The so-called "boiling springs" on the Hattenberger farm, about three and a half miles west of Savage, are probably of this same type. Tests indicate that the dolomite lies only 30 feet below these springs, which form a pool along a small creek valley. These springs "boil" vigorously at intervals of a few minutes. Normally the "boiling" raises the water a foot or more above the surface of the pool, but occasionally the agitation is much more violent, reaching a height of two or three feet. The water is of normal temperature and the bubbling is merely an upwelling under pressure, probably owing to a very fine suspended clay in the pool, which settles down and confines the water until the pressure builds up sufficiently to burst through the clay. This would also explain why the "boiling" shifts from place to place in the pool. It is possible that the water in this spring may be supplied from fissures in the dolomite rather than from deposits of drift.

According to the 1982 Geologic Atlas of Scott County, the bedrock at Boiling Springs is Prairie du Chien Group and the surficial geology is Holocene "Dune Sediment". Groundwater flowing upwards under artesian conditions from the underlying dolomite bedrock passes through the unconsolidated overburden onto the surface.

The heyday of tourism at Boiling Springs seems to have been the first half of the 20th century. Photographic postcards of the spring were made and sold in the 1910s especially. Every now and then, articles advertising Boiling Springs as a "natural wonder" would appear in local newspapers. Three generations of Hattenbergers hosted visitors to the site and offered homemade wine, replaced by ice cream and candy in the '20s through '40s. Today, Boiling Springs lies within a grove of trees, almost concealed from the view of the houses nearby. Apart from minor trails made by the tread of occasional visitors' feet, nobody would ever know the springs were widely visited in the past. Historically known to the Dakota people as *Maka Yusota*, Boiling Springs is nonetheless one of the most significant places in all of Scott County. The "Boiling Springs" of Savage — Scott County Historical Society



Figure 3.7.25 Boiling Springs, 3.5 miles west of Savage in Scott County. Water discharges through unconsolidated sediment with upwelling under pressure from the underlying Prairie du Chien Group dolomites, which creates the boiling effect. Photo by McGhiever; https://commons.wikimedia.org/wiki/File:Maka_Yusota.jpg; under Creative Commons Attribution-Share Alike 4.0 International license.

3.7.3 Iowa

Big Spring

Big Spring along the Turkey River in northwestern Clayton County, 10 miles northwest of Elkader, is the largest spring in Iowa. It reportedly discharges as much as 295 cfs after heavy rains, with the minimum discharge of 32 cfs (Halberg et al., 1984). Big Spring is fed by a groundwater basin developed in the Ordovician Galena Group limestones and dolostones comprising about 103 square miles. It serves as the water source for the Big Spring Trout Hatchery, with average flows ranging from 20,000 to 30,000 gallons per minute.

The Big Spring Galena aquifer is made up of only the three youngest Formations of the Galena Group: the Dunleith, Wise Lake, and Dubuque Formations. These three units are comprised of interbedded limestones and dolomites. Regionally the degree of dolomitization decreases toward the north. Some shale interbeds occur, principally in the Dubuque Formation. The Dubuque Formation tends to be well bedded with shaly partings, while the Wise Lake is more massive. The Dunleith tends to be cherty. The rocks of the Galena aquifer average about 220 feet (67 m) in total thickness in the spring drainage area. All the units are jointed or fractured. Karst-solutional activity along joints and bedding planes is obvious in nearly any exposure or quarry visited. Major joints are clearly widened by solutional activity, and along many joints, deposition of secondary calcium carbonate (flowstone, travertine, etc.) is evident. Many exposures show sinkholes, small "dome pits," and even small conduits or caves formed in the rocks. Investigations of newly formed sinkholes by Iowa Geological Survey staff have revealed natural, open, vertical solution shafts which go down 30 to 120 feet (10-35 m) below land surface, into the aquifer (Halberg et al., 1983).



Figure 3.7.26 Big Spring, Photo courtesy of Joe Honings.



Figure 3.7.27 The Big Spring Hatchery near Elkader in Clayton County, seen in an aerial photo, raises more than 200,000 rainbow trout a year. The fish are part of the 130 million fish a year the Iowa Department of Natural Resources stocks into lakes and streams. Photo courtesy of Nick Rohlman/The Gazette, Cedar Rapids, Iowa.

Coldwater Spring

Coldwater Spring is in Coldwater Cave State Preserve, a 60-acre public area which encompasses four miles of Coldwater Creek, one of several cold-water trout stream tributaries to the Upper Iowa River that is open to public fishing. Coldwater Spring issues from Coldwater Cave, at the base of a 150-foot-high bluff and flows a short distance into Coldwater Creek, and from there to the Upper Iowa River. The cliff is composed of Galena limestone, a Devonian (375-million-year-old) rock formation.

Coldwater Spring is the natural exit of an underground stream flowing through Coldwater Cave, the largest and most elaborate cave system in Iowa. It is in northern Winneshiek County, five miles from the Minnesota border, on land owned by the Iowa Department of Natural Resources since 1943 (managed as part of the Cold Water Creek Wildlife Area). The presence of the larger cave system was not discovered until 1967. In 1969, a 60-acre tract around the spring was dedicated as a geological state preserve. The spring itself is presently the only part of the cave system that is contained within the state preserve. Part of the cave was designated as a National Natural Landmark in 1987 due to geologic and ecological features considered to be of national significance. The cave is located on private land, and exploring by hobbyists is strictly prohibited (https://www.iowadnr.gov/places-go/state-preserves/cold-water-spring-state-preserve.)

Coldwater Cave system is relatively young and still forming. It is part of a fluvio-karst drainage basin located in northeast Winneshiek County, Iowa and southeast Fillmore County, Minnesota. The cave proper is situated in the Iowa part of that basin. Formed almost entirely within the Dunleith Formation of the Galena Group (Ordovician), the cave consists of over four miles of borehole and stream gallery, nearly a mile of parallel stream passage and another 11 miles of in-feeders and their associated offshoots. The cave system, which is dendritic in its layout, is developed within a subtle carbonate ridge bounded by surface drainages; some of the side passages cross under these drainages. The entire area is mantled with loess and glacial till.



https://www.onlyinyourstate.com/nature/iowa/secret-cave-ia

There is only one natural entrance to the cave, the Coldwater Spring itself. Visitors can hike along a designated path to see this entrance, but this underwater entrance is currently gated. This cave is not open to tourists. Due to extremely difficult access, fluctuating water levels, and risk of injury, only a few experienced adult cavers are granted limited, guided access (Winneshiek County Discovery Area - Driftless Karst Trail.)

Figure 3.7.28 Coldwater Spring issues from Coldwater Cave, the longest cave in Iowa. It has average flow rate of 24 cfs. Photo by Roberta Bumann, available at

Dunning Spring

Dunning Spring is in Dunning's Spring Park which might be one of the prettiest attractions in all of Iowa and is home to the most visited waterfall in the state. This small park has a rich history in both commerce and now as a place of recreation. Before it was donated to the City of Decorah, it was once the site of a grist mill and a marble quarry. Today, although no sign of commerce remains, visitors walk to the park to see the continuously flowing 200-foot waterfall, which starts from a small cave in bluff of Ordovician-age Galena Group limestones, cascades down the limestone hill, and feeds a small cold-water trout stream that tumbles into the Upper Iowa River. A bridge, boardwalk and staircase provide access to the top of the bluff to see the spring. This park is part of a larger system of Decorah parks that connect to each other and is officially named "Dunning's Spring Park and Ice Cave Hill", which have trails and other recreation opportunities (https://driftless.caves.org/index.php/iowa-discovery-areas/winneshiek-county-ia-discovery-area/).

Decorah Ice Cave is open to the public as an "enter at your own risk" attraction, meaning there is no supervision, admission cost, or set hours to visit. It is located beneath a south-facing wooded bluff of Ordovician-age Galena Group limestones exposed along the valley of the Upper Iowa River. Decorah Ice Cave is believed to be the largest

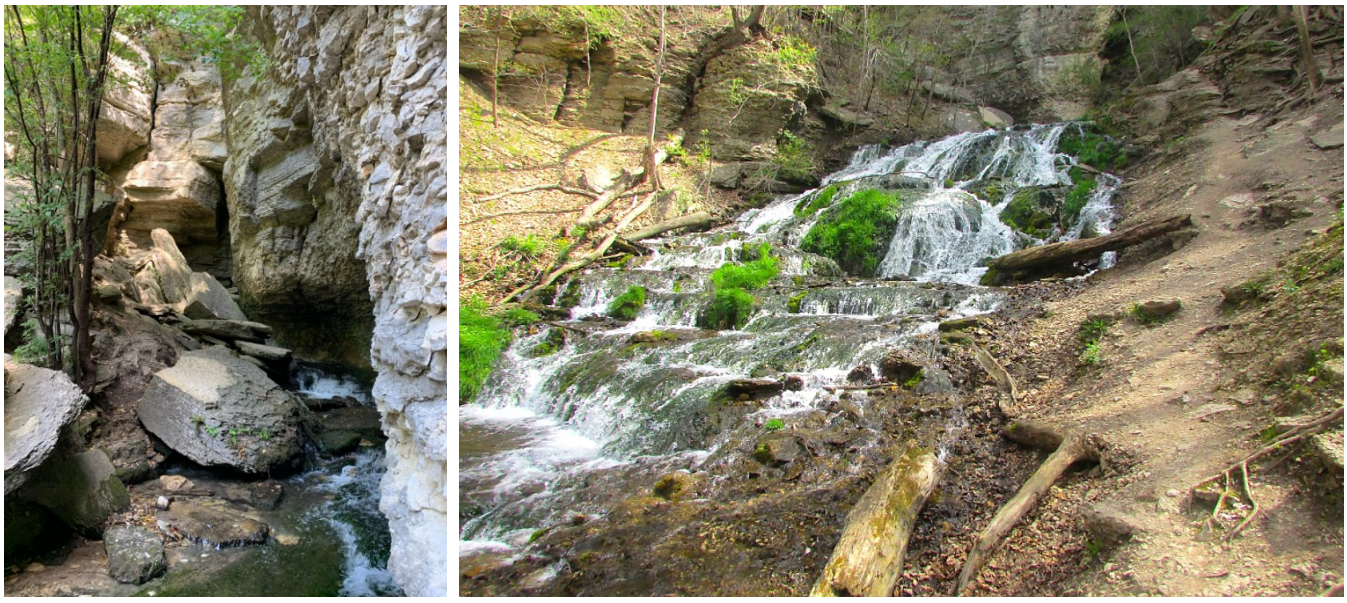


Figure 3.7.29 Dunning Spring starts from a small cave in Ordovician-age Galena Group limestones and has average flow rate greater than 10 cfs. Water flows down a 200-foot-long waterfall that starts a small cold-water stream. Water temperature is constant 45° Fahrenheit. Left photo courtesy of Joe Honings. Right photo courtesy of https://visitdecorah.com/

known ice cave in eastern North America, unique for the unusual deposits of ice which coat its walls. The cave passage follows a prominent vertical fracture through the limestone, enlarged by the slipping of large rock blocks downslope. Ice formation is caused by sinking of cold air in the cave during winter along a network of crevices which extend back into the uplands north of the bluff. The air drainage cools the surrounding rock to temperatures well below freezing. Ice build-up occurs with the spring thaw when surface water seeps into the cave and is frozen by contact with the cold rock walls and entrapped cold air. Ice begins to accumulate in March, reaches a maximum thickness of 8 to 10 inches by June, and remains until August or September. The cave has an interesting history of scientific investigation and was shown as a commercial enterprise from 1929 until 1941.



Figure 3.7.30 Decorah Ice Cave; courtesy of Travel Iowa, https://www.traveliowa.com/places/ice-cave-state-preserve/7631/

Siewers Spring

Siewers Spring in Winneshiek County is recognized as the second largest spring in Iowa (after Big Spring in Clayton County), with an average flow rate of 23 cfs. It is located near Decorah at the Decorah Fish Hatchery, provides the cold water necessary for trout rearing, and flows into Trout Run Creek. The spring issues from Ordovician Galena limestone.

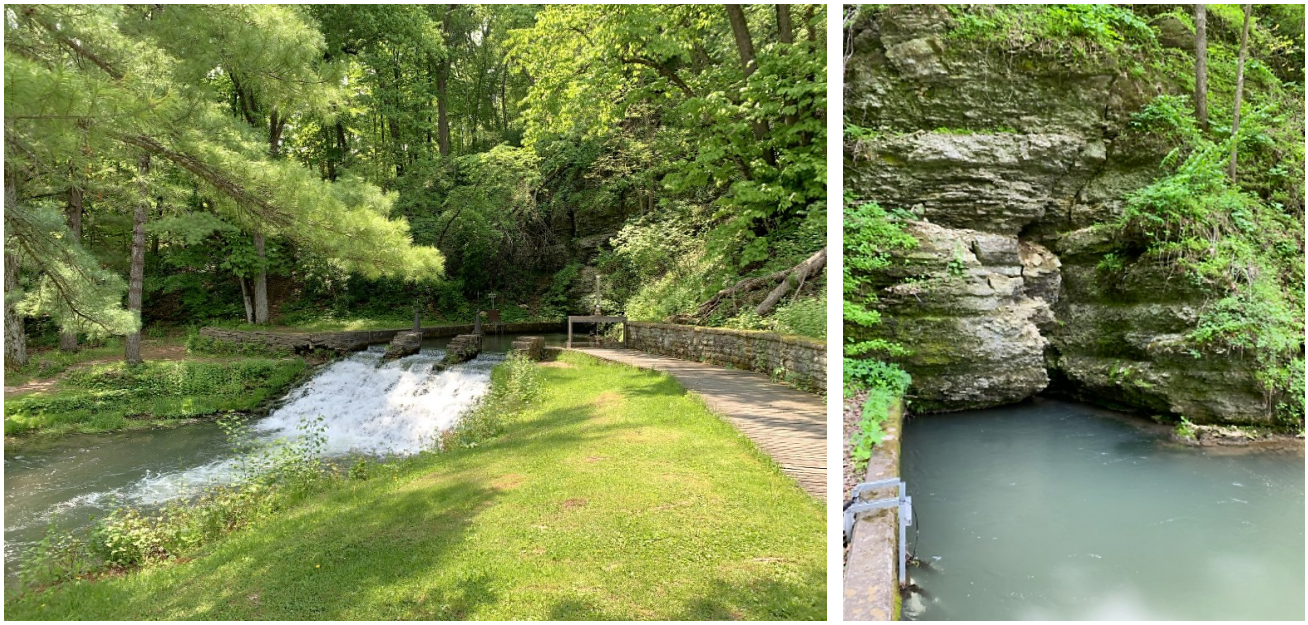


Figure 3.7.31 Siewers Spring. Photographs courtesy of Joe Honings.

Malanaphy Springs

Malanaphy Springs is in Malanaphy Springs State Preserve, a 64-acre public area along the Upper Iowa River near Decorah. The springs issue from crevices in high cliffs of Galena Group limestone, which is approximately 450 million years old (Ordovician age). The spring water remains cold year-round, typically under 60°F (15°C), and flows strongly even during dry summer months.



Figure 3.7.32 Malanaphy Springs emerge from a cave at the bottom of a limestone cliff (*left*) and water tumbles over limestone colluvium rocks forming a series of cascading falls and small pools for a 60-foot drop, until it reaches the river. The last 10 feet is the most impressive part of this waterfall (*right*) and what paddlers on the Upper Iowa River see. Photographs courtesy of The Waterfall Record. Available at https://waterfallrecord.com/2014/07/20/malanaphy-springs-falls-iowa/

The Preserve is located on the Paleozoic carbonate platform and houses more than 300 native vascular plants including sugar maple and basswood in the forest canopy and ironwood in the understory. Although there are dozens of springs that flow into the waters along the river float from Bluffton to Decorah, Malanaphy Springs is the best-known spring along the Upper Iowa River.

3.8 Illinois, and Michigan

3.8.1 Illinois

Most springs in Illinois are located in the Mississippian, Devonian, and Pennsylvanian limestone and sandstone of the Shawnee Hills of southern Illinois and along the western border of the state, extending north to the Driftless area of Carroll and Jo Daviess counties (Fig. 3.8.1). Several springs also originate from the base of the bluffs along the Fox, Illinois, and Rock rivers. Only a modest number of springs, however, occur in the central plains of Illinois (Springs of Illinois – Introduction and Overview – Mark J. Wetzel.) Unlike in the driftless area of Minnesota, Wisconsin, and Iowa, there are no significant karst springs in the driftless area of Illinois which occupies the northwestern corner of the state.



Southwestern Illinois contains a prominent karst region known as the Sinkhole Plain, which primarily covers Monroe County and portions of St. Clair and Randolph Counties. It is specifically recognized as the Illinois part of the physiographic province Salem Plateau of the Ozark Uplift. The Salem Plateau, most of which is in Missouri, was divided this way by incision of the Mississippi River valley.

Figure 3.8.1 Karst areas of Illinois. The bedrock aquifer of the Driftless Area of northwestern Illinois is Ordovician-age Galena Dolomite. Shawnee Hills karst is underlain mainly by Salem, St. Louis, Ste. Genevieve, Glen Dean, and Menard Limestones. Salem Plateau (Sinkhole Plain) is composed of Mississippian strata ranging from the Valmeyeran Salem Limestone to the Chesterian Kinkaid Limestone; karst features are most prominent in St. Louis, Salem, and Ste. Genevieve Limestones. Most of the west part of the Lincoln Hills karst region occur in either the Kimmswick or St. Louis Limestones. The sinkholes in the east part of the region, in and near Alton, are associated primarily with the St. Louis and Ste. Genevieve Limestones (Panno et al., 1997). Map is from Panno and Kelly, 2015, modified from Weibel and Panno 1997.

The narrow area of the Sinkhole Plain exhibits classic karst geology and hydrogeology, contains a high density of cover-collapse sinkholes (Figs. 3.8.2 and 3.8.4), and hosts the largest karst springs in Illinois. Steep bluffs up to 200 feet high separate the highlands of the Sinkhole Plain from the lowlands of the Mississippi River valley to the west. Bedrock in the Sinkhole Plain is Mississippian-age limestone overlain by a relatively thin layer (0 to 60 ft thick) of Pleistocene-age glacial till and loess. The area is replete with cover-collapse sinkholes, large flashy springs, and large branchwork cave systems (see image in Fig. 3.8.4). Outcrops and road cuts in the area reveal limestone bedrock festooned with solution-enlarged crevices and overlain by a well-defined epikarst (Panno et al. 1996; 2011).

Karstification in the sinkhole plain occurs primarily in the St. Louis and Ste. Genevieve Limestones, which are 97% calcite; consequently, these formations are highly soluble. Numerous fractures in the bedrock, roughly trending north to south and east to west have developed into solution-enlarged crevices that, along with numerous



Figure 3.8.2 Aerial view of the Illinois sinkhole plain looking west across the Mississippi River valley toward Missouri's Salem Plateau of the Ozark Uplift. Photograph by Joel M. Dexter; from Panno et al., 2011.

branchwork-type cave systems, make up the shallow karst aquifer. The carbonate rocks dive beneath Pennsylvanian age shale to the east; on the west, the sinkhole plain is bounded by 60- to 100-m bluffs of the Mississippi River and its floodplain. Structures in the area include the northwest-southeast–trending Waterloo-Dupo Anticline and Columbia Syncline to the north and Valmeyer Anticline to the south (Nelson 1999).

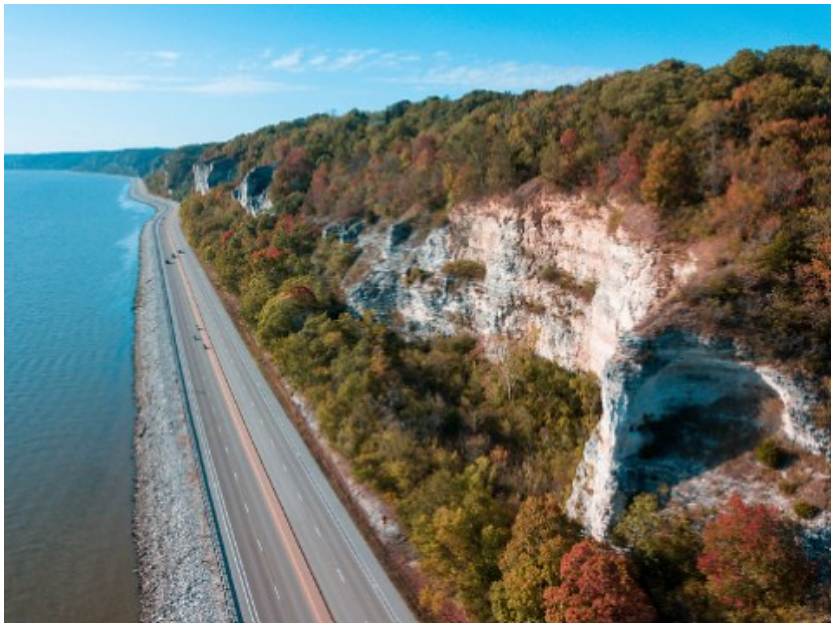


Figure 3.8.3 The stretch of River Road between Alton and Grafton is unlike any other place in Illinois. This area is the confluence of the Mississippi, Missouri and Illinois Rivers, Complete with a full view of the Mississippi and towering white limestone cliffs. From *I Like Illinois; Great River Road: Limestone Cliffs in Alton and Grafton*. Available at Great River Road: Limestone Cliffs in Alton and Grafton

Many sinkholes in the sinkhole plain remain open to water flow and become direct conduits that focus surface runoff into the karst groundwater systems with little filtration. This direct connection makes groundwater, springs, and cave streams quite vulnerable to contamination and pollution. A 2014 study conducted by scientists at the University of Illinois Prairie Research Institute, with support from Clifftop and Southwestern Illinois College,

analyzed water from 10 caves and springs in the Sinkhole Plain and discovered widespread presence of fecal coliform bacteria and chemicals from pharmaceuticals and personal care products (PPCPs), like the lipid-lowering drug Gemfibrozil and an antibacterial chemical common in soaps, Triclocarban. Stemler Cave and its resurgence point, Sparrow Spring, had the highest concentrations of PPCPs: 7 of the 15 chemicals tested for were detected.

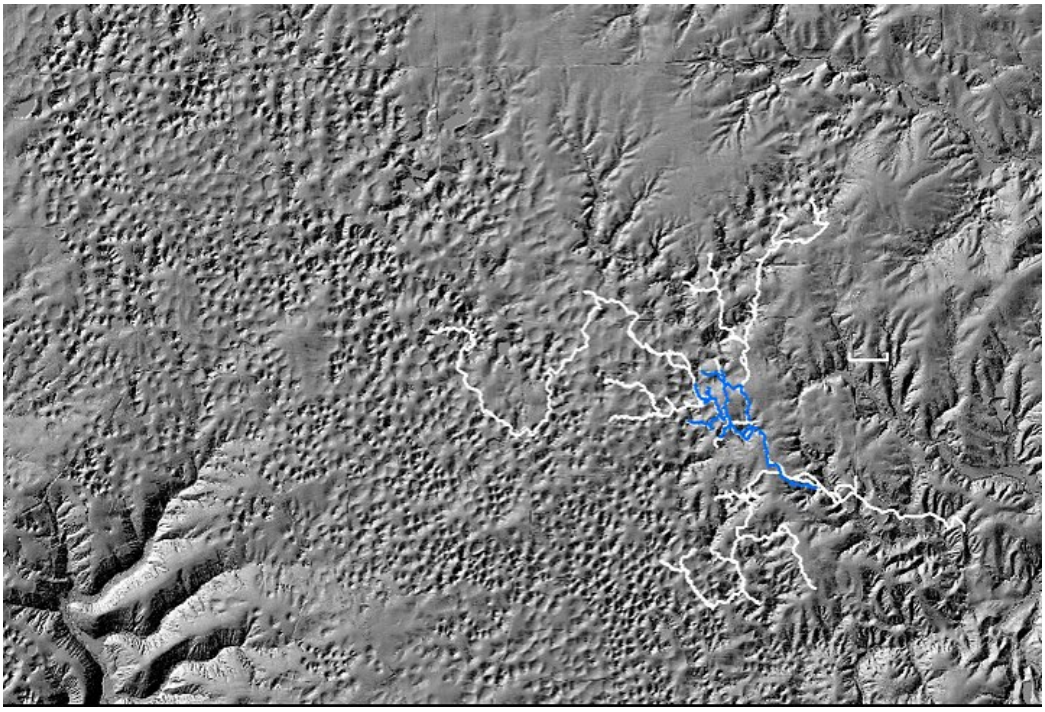


Figure 3.8.4 Lidar (light detection and ranging) image of Monroe County, Illinois, reveals thousands of sinkholes. White and blue lines denote Fogelpole Cave, the largest and longest cave in Illinois with over 15 miles of surveyed passages. From "Mapping Fogelpole. Washington University geologists are helping to survey a sinkhole plain cave that drains a big chunk of Illinois"; available at Mapping Fogelpole - The Source - WashU.

The study suggests septic systems (poorly maintained home aeration systems and faulty leech fields) are the source of the PPCPs and some of the bacteria that are contaminating the groundwater systems in the area (Thousands of Sinkholes, Hundreds of Caves, Dozens of Rare Species…. Recent Discoveries and New Data About Our Karst Landscapes | Clifftop).

Indian Hole Spring (Fig. 3.8.5) in Monroe County is the primary resurgence point for the groundwater basin of Fogelpole Cave, the largest and most biologically diverse cave in Illinois. Located in the sinkhole plain of southern Monroe County, the spring drains approximately 10,000 acres of karst landscape dotted with hundreds of sinkholes. Indian Hole Spring and the Fogelpole Cave entrance are located within the Fogelpole Cave Nature Preserve, which is only open to the public on the surface. The cave itself is gated and strictly closed to unauthorized visitors to protect its fragile ecosystem and endangered species.

Indian Hole Spring is a relatively small, shallow pool; its water flows up through solutionally sculpted Mississippian-age St. Louis Limestone at the head of a blind valley (box canyon) leading to South Fork Creek. The results of dye-tracing experiments in Fogelpole Cave indicate that Indian Hole Spring is the resurgence of Fogelpole Cave (Wightman 1967, Aley and Aley 1998). Discharge from this spring under baseflow conditions was estimated at approximately 1,500 gpm. Indian Hole Spring is separated by a low ridge from a larger blind valley containing a circular, water-filled sink hole most often referred to as "Indian Hole" (Fig. 3.8.6). Dye-tracing experiments

showed that the two features are hydraulically connected (Aley and Aley 1998). The Indian Hole sinkhole flows only during and immediately following significant rainfall events, but when it does flow, it pirates most of Fogelpole Cave's discharge from Indian Hole Spring. During one of these pirating events (February 17, 1998), discharge from the sinkhole and spring was measured at 130,000 gpm (Panno et al., 2003).

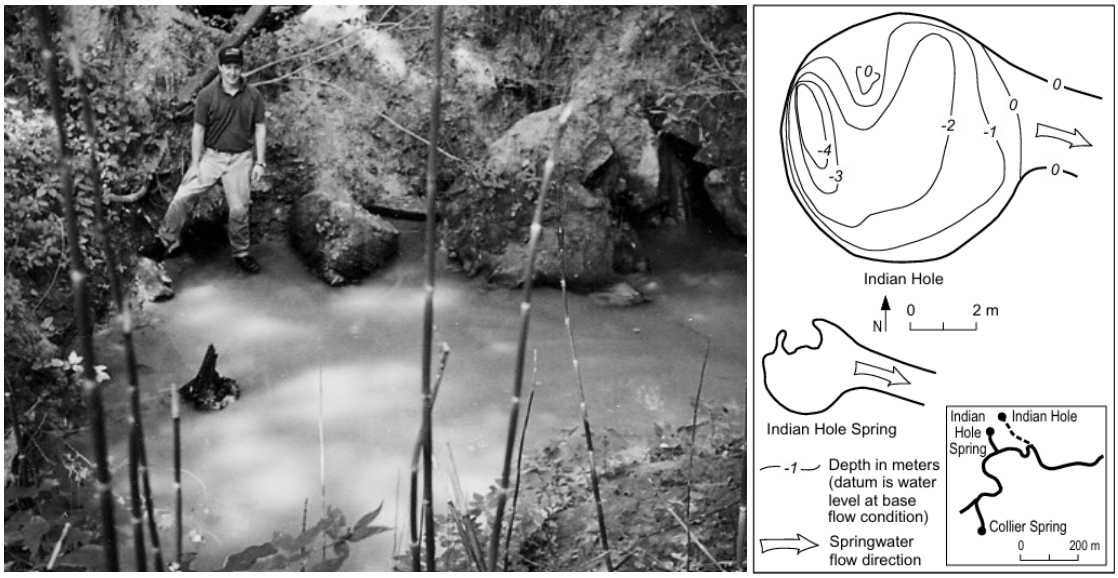


Figure 3.8.5 *Left*: Indian Hole Spring is located at the head of a blind valley and is the resurgence of Fogelpole Cave. *Right*: Contour map of Indian Hole (top) and Indian Hole Spring (bottom). Datum is water level under baseflow conditions. The two features are hydraulically connected, and Indian Hole flows only during periods of flooding when it pirates water from Indian Hole Spring. The deepest part of Indian Hole is a fissure from which water discharges during high-water conditions. From Panno et al., 2003.



Figure 3.8.6 Indian Hole does not flow under baseflow conditions (*left*) but pirates water from Indian Hole Spring during periods of flooding (*right*). Photographs courtesy of Samuel A. Panno.

Collier Spring

The Collier Spring groundwater basin discharges at Collier Spring located about 1,300 ft from Indian Hole Spring. Water from this spring flows upward from a circular pool bounded by limestone bedrock (Mississippian-age St. Louis Limestone) at the head of a blind valley and forms the headwaters of South Fork Creek. Discharge from Collier Spring under base flow conditions was estimated to be about 1,300 gpm (Fig. 3.8.7-*Left*), and under high flow conditions was about 160,000 gpm (Fig. 3.8.7-*Right*).



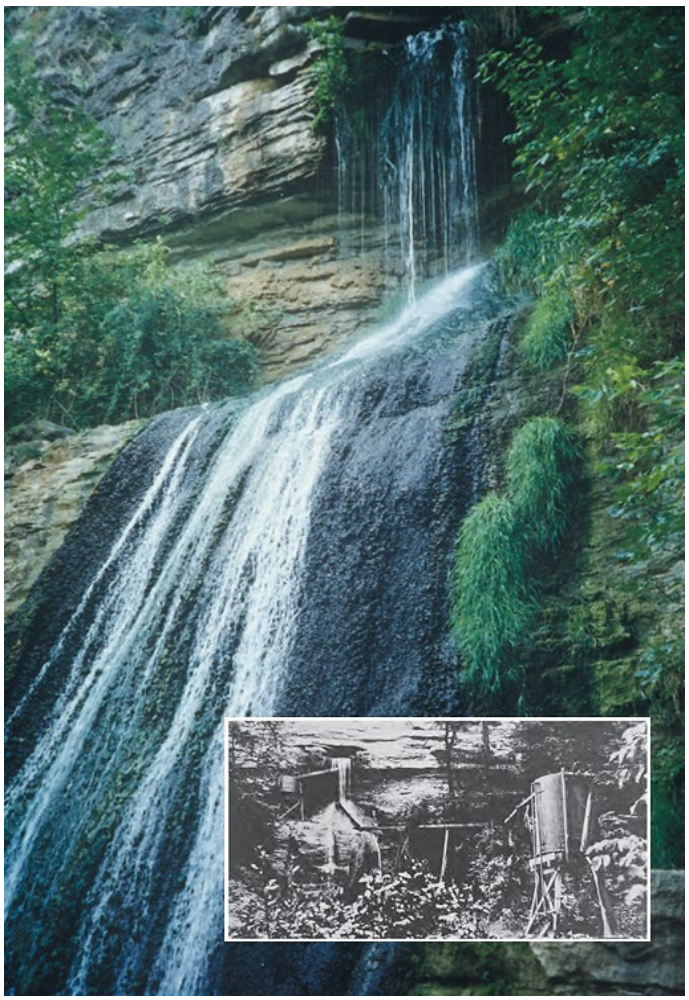
Figure 3.8.7 Collier Spring, a circular feature and the headwaters of South Fork Creek, is also the resurgence of a cave or conduit system that, at some point, flows beneath Fogelpole Cave. Collier Spring under base flow conditions (*left*) and under flood conditions (*right*). Photographs courtesy of Samuel A. Panno.

Falling Springs

Falling Springs near the town of Dupo in St. Claire County is a perched cave spring that discharges from a 3-by 6-feet cave along a bedding plane about midway up a steep, 150-ft bluff (Fig. 3.8.8). The bluff is predominantly Mississippian-age St. Louis Limestone capped by several meters of St. Genevieve Limestone (Zakaria Lasemi and Rodney Norby, ISGS geologists, personal communication 1998) and is located along the eastern edge of the Mississippi River Valley. The spring water cascades over a reddish black tufa deposit that has encrusted local vegetation. Tufa is a porous deposit of calcium carbonate that is associated with calcareous springs and seeps. Historically, the spring was used as a source of fresh water. Among other uses, during the late 1800s and early 1900s, the spring supplied water to steam locomotives that served nearby limestone quarries (Figure 3.8.9 inset). More recently, a foot-wide iron trough near the base of the tufa apron was installed in about 1970 and currently supplies water to adjacent fishponds. Based on the fact that the tufa has now grown over the trough and extends beyond it by about 1.5 ft, the growth rate of the tufa near the base of the apron is approximately 1 in./yr. Precipitation of minerals from the spring water is so rapid that leaves from the fall may be found totally encrusted with calcite before they get washed away by flooding in the spring. Tufa on the right of the apron is continually removed with a backhoe so that water continues to fill the trough (Panno et al., 2011.)



Figure 3.8.8 Falling Spring in St. Claire County discharging from a bluff in Mississippian-age St. Louis Limestone. Photo courtesy of https://www.cecinc.com/project/surface-limestone-mine-expansion/



Discharge from the spring at base flow is clear and was measured at approximately 600 gpm (Panno and Weibel 1999). During and following heavy rains, discharge typically is laden with suspended solids, and the volume of discharge appears to be one to two orders of magnitude greater than that of base flow.

Figure 3.8.9 Falling Springs is a perched cave spring near Dupo, St. Claire County that discharges from a bedding plane cave in the upper St. Louis Limestone. The perched cave spring bears witness to the rapid entrenchment of the Mississippi River valley, probably during the Pliocene to early Pleistocene Epoch. In the early 1900s, the spring was used to supply water for domestic use and for steam locomotives (inset). The railroad, steam locomotives, and water towers are gone today, but the spring now provides water for nearby fishing ponds. From Panno et al. 2011.

As with most springs in the sinkhole plain, water quality from Falling Springs is extremely poor. The water is contaminated with a variety of bacteria derived from both natural and anthropogenic sources. Given the likely recharge area of the spring, bacteria sources probably include wastes from livestock, pets, and private septic systems as well as naturally occurring bacteria. Given the abundance of suspended sediments in the spring water under high flow conditions, it would appear the dominant source of water during high flow comes via sinkholes. Sinkhole density in the highlands above Falling Springs is estimated to be as high as 56 sinkholes/km^2 (Panno et al. 2009b)



Figure 3.8.10 A well-manicured sinkhole pond on uplands above Falling Springs showing a nearly perfect circle. From Panno et al., 2011.

Sparrow Creek Spring,

Sparrow Creek Spring, located on private property in Union County near East Cardondelet, is a partially submerged, 15 feet wide, 7 feet high opening at the base of a steep, loess-covered, blind stream valley. It is the resurgence of the Stemler Cave groundwater basin, the headwaters of Sparrow Creek, and likely the largest spring in the state. The groundwater basin is about 4.3 mi long and 1.2 mi wide with an area of approximately 10 mi^2 (Aley et al. 2000). Sparrow Creek Cave is hydrologically connected to Stemler Cave to the south. Sparrow Creek Cave consists of an 800-ft-long, northwest-trending passage that leads to a sump and a 1,000-ft-long side passage that becomes constricted at its distal end (Sherrill 1989). At low flow, water discharges from the cave at a rate of about 7,450 gpm (Panno et al. 2001). The highest discharge measured at Sparrow Creek Spring was 180,000 gpm following a heavy rainfall (Sherrill 1989).



Figure 3.8.11 Sparrow Creek Cave Spring, the largest spring in the sinkhole plain of Illinois, is located near Carondelet, which is just northeast of Columbia, Illinois. It issues from Mississippian St. Louis Limestone. Photograph by Samuel V. Panno; from Panno et al., 2011.

In 1986, the Stemler Cave Woods Nature Preserve, a 120-acre remnant of old growth, dry upland forest was dedicated and is now maintained by the Illinois Department of Natural Resources. Sinkholes in the area create microclimates that result in black and white oak trees dominating the upper parts of the sinkholes and red oak trees dominating the lower parts. In 2004, thanks to extensive efforts by The Karst Conservancy of Illinois, the nature preserve (now over 4,500 acres) and the Stemler Cave groundwater basin received a Class III: Special Resource Groundwater designation from the Illinois Environmental Protection Agency to help protect the biodiversity of the cave and karst system.

Camp Vandeventer Spring

Camp Vandeventer Spring is located at Camp Vandeventer 3.7 miles west of Waterloo and is owned by the Boy Scouts of America (BSA). It issues from a small cave in the St. Louis Limestone at the base of a 40–ft-high bluff (Fig. 3.8.12). Discharge from the spring was measured at 550 gpm at low flow (Aley and Aley 1998). The spring flows directly into Fountain Creek through an entrenched channel that at one time was the water source for the camp. As with most springs in the sinkhole plain, water samples from this spring are always contaminated with fecal coliform bacteria; the spring is among the most severely contaminated in the sinkhole plain. The dominance of Escherichia coli and the presence of optical brighteners in the spring water suggest that private septic systems are a primary source of contamination (Panno et al. 2001; Joan Bade, sanitarian, Monroe and Randolph County Health Departments, personal communications 1997). The rRNA source tracking suggested both human and pig sources (W.R. Kelly, Illinois State Water Survey, unpublished data).



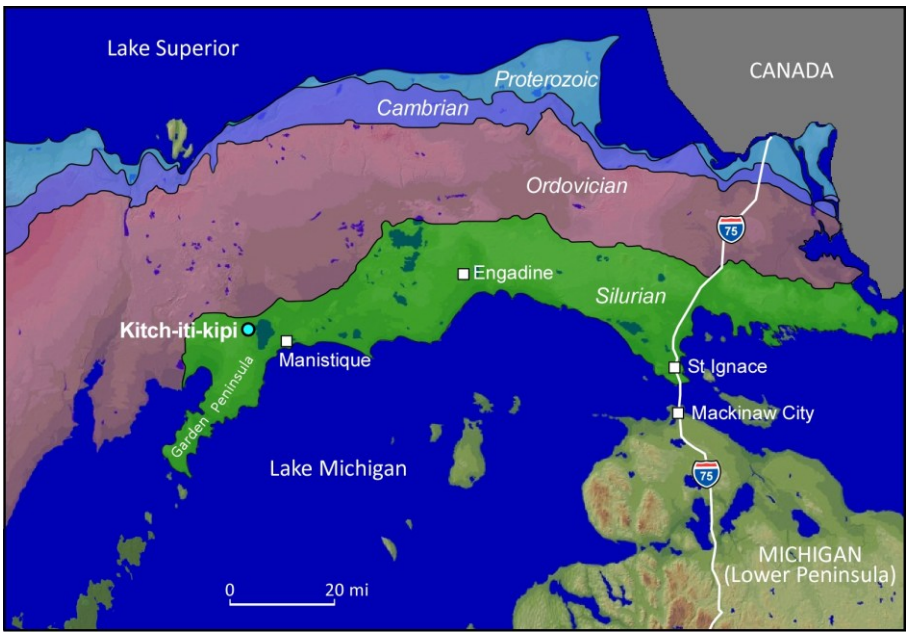
Figure 3.8.12 *Left*: Camp Vandeventer Spring, located at the base of the Mississippian St. Louis Limestone bluff is a typical large cave spring in Illinois' sinkhole plain. *Right*: Karst window at Camp Vandeventer formed when the roof of a cave collapsed and formed this 30-ft-deep cavity. Groundwater flows from the far side of the window, disappears into the near side, and flows to Camp Vandeventer Spring. Photographs taken after a storm event. From Panno et al., 2011.

3.8.2 Michigan

Kitch-iti-kipi

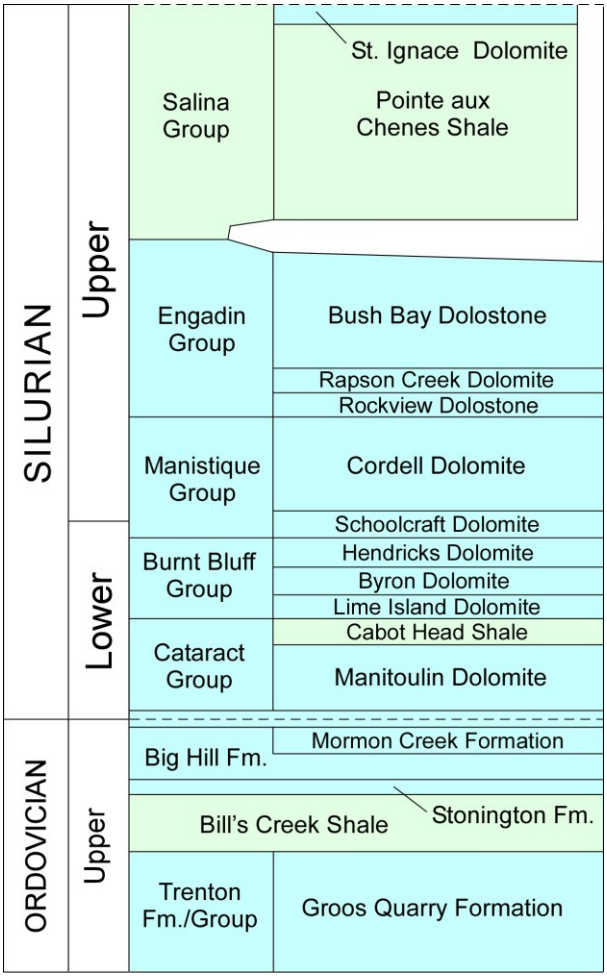
The Michigan's largest spring, Kitch-iti-kipi (translated from the native American Ojibwe (Chippewa) language to meanings like "Big Cold Water" or "Big Cold Spring") is in Palms Book State Park, Schoolcraft County, located by the northwest end of Indian Lake in Michigan's Upper Peninsula (Fig. 3.8.13). The park is 11 miles from Manistique, small town on the shores of Lake Michigan, via County Roads 442 and 455, and State Highway M149 (which changes to Sawmill Drive). Wayfinding signage refers to the park as Palms Book; however, locals also refer to the park as The Big Spring and Kitch-iti-kipi. The state of Michigan received the spring with accompanying land as a gift from the federal government in 1926, under the stipulation that it be turned into an official public park. It is now one of the most visited state parks in Michigan. Kitch-iti-kipi site is recognized by the state as "Traditional Cultural Property significant to the heritage of the Great Lakes Anishinaabek."

The Precambrian surface upon which the Paleozoic sedimentary rocks of Schoolcraft County are deposited generally slopes southeastward toward the center of the Michigan basin. The Paleozoic rocks of the basin were deposited in nearly horizontal layers, but gradual subsidence and compaction of the beds, which was contemporaneous with deposition and greatest in the center of the basin, produced a bowl-shaped structure (see Chapter 2.9 for more details). The youngest beds are exposed at the surface in the central part of this structure in central Michigan, and the older formations crop out in roughly concentric bands. Schoolcraft County is near the northern edge of the basin, where older sedimentary rocks are exposed. The regional dip of these formations in the county is to the southeast at about 40 feet per mile. The formations tend to become thicker in that direction.



Kitch-iti-kipi is fed by groundwater from the Early Silurian Burnt Bluff formation which discharges through a thin layer of glacial sand and gravel deposits at the bottom of the spring basin, forming powerful sand boils.

Figure 3.8.13 General map of Michigan's Upper Peninsula with principal geologic periods of bedrock rocks and the location of Kitch-iti-kipi (Big Spring), the largest spring in Michigan. shaded relief base from USGS. Geologic boundaries simplified from https://www.michigan.gov/egle/maps-data/geowebface



The Burnt Bluff formation is composed of about 250 feet of thin-bedded to massive light-gray to buff dolomite and dolomitic limestone. Some of the upper strata of the formation are beds of high calcium limestone, which is especially soluble. Caves have been formed by solution of portions of the high-calcium limestone members in adjacent areas in Mackinac County. The water in the formation is generally confined under sufficient artesian pressure such that a number of wells tapping this formation flow at the land surface. The Burnt Bluff formation is the most important aquifer in the county, as it is the source of water to most of the wells in the populated areas (Sinclair, 1959).

Figure 3.8.14 Stratigraphy of the Silurian carbonate aquifer of Michigan's Upper Peninsula. From Swezey, 2008.

Most of the wells in the city of Manistique are finished in this aquifer at a depth of about 200 feet. The water produced from some of these wells, however, contains large quantities of calcium and sulfate which is not characteristic of the Burnt Bluff formation. Apparently, water yielded by these wells is contaminated by upward leakage of calcium sulfate waters from the underlying Cataract formation. Some of the wells may be bottomed in the upper part of the Cataract formation and derive water from both aquifers. The Cataract formation is composed of buff to gray, dense, shaly dolomite with layers of gypsum and greenish-gray shale. This formation ranges in thickness from about 150 to 200 feet. The Cataract is permeable as a result of solution openings which have been formed by leaching of the gypsum beds. This aquifer has been tapped for water at Blaney Park, at Manistique, and in the vicinity of Indian Lake, but in these areas, it generally yields saline water. Generally, fresh water cannot be obtained from this aquifer anywhere in Schoolcraft County because of the soluble gypsum present in the formation (Sinclair, 1959). Chemical analyses of Big Spring water show somewhat elevated concentration of sulfates, higher alkalinity and specific conductance, indicating that the aquifer feeding the spring includes both the Burnt Bluff and the Cataract formations.

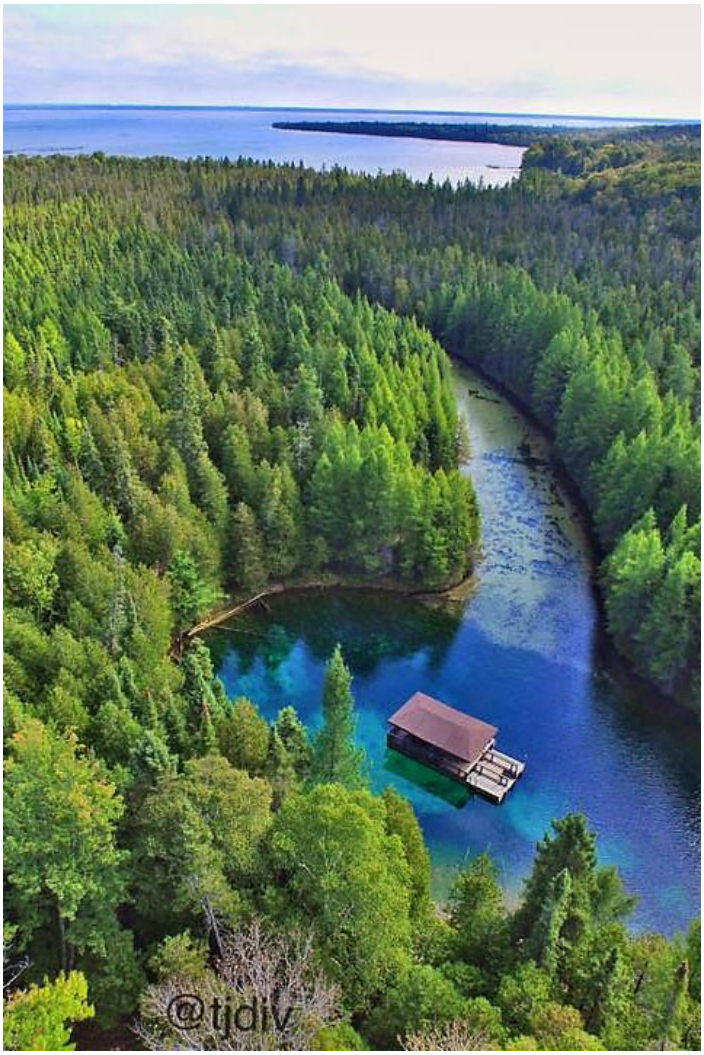
A considerable amount of water is discharged from the Burnt Bluff and Manistique Group aquifer by springs. Several springs are located along the west shore of Indian Lake. Kitchi ti-ki-pi is the largest and best known, with an average flow rate exceeding 10,000 gallons per minute (gpm). For comparison, another nearby spring has a discharge rate of nearly 1,400 gpm and provides enough cool water to operate the Thompson State Fish Hatchery that produces Steelhead, Chinook Salmon, Walleye and Muskellunge annually (Bauman, 2023).

Due to high groundwater inputs, the chemical and physical characteristics of the Big Spring vary little from the bottom to the surface of the water column. For example, year-round bottom to surface temperature ranges from 45 to 46 degrees Fahrenheit respectively. Dissolved oxygen concentrations also have a narrow range from the bottom to the surface, and values tend to be slightly hypoxic. Water transparency in the Big Spring is exceptionally high from 28 to 39 feet deep, indicating low nutrient levels (Bauman, 2023).





Figure 3.8.15 Kitch-iti-kipi or "The Big Spring". The spring is roughly an oval pool 300 by 175 feet with sides which slope abruptly to a depth of 40 feet. Water enters the pool under strong artesian pressure, roiling the sandy bottom to such an extent that a boiling effect is seen from the surface. The beautiful emerald green of the water when viewed from the shores could be due to the luxuriant growth of aquatic vegetation draping the submerged trunks and sides of the pool. A self-operated raft is the most popular means to explore the spring. Photographs courtesy of Michigan Department of Natural Resources.



http://www.visitmanistique.com/bigspring.shtml

During the 1950s there was interest on behalf of several agencies to quantify the discharge of the Big Spring. Present day signage suggests that the total discharge of the Big Spring is more than 10,000 gallons per minute. Michigan Department of Natural Resources' interoffice communications, archived from the 1950s, suggest that the discharge of the Big Spring is often higher than 10,000 gallons per minute. Discharge was quantified nine times from 1950 to 1952. The average discharge from these nine measurements was 14,833 gpm with a range in discharge from 12,200 to 16,700 gpm. Discharge reportedly was higher during the spring and summer months compared to the fall. These discharge estimates would suggest that, based on an estimated volume of 8 to 14 acre-feet, the Big Spring has a complete turnover rate of approximately once every three to six hours (Bauman, 2023).

Figure 3.8.16 Westerly view over Kitch-iti-kipi towards Green Lake. Photo by Thomas J Dolaskie IV.

How Kitchi-iti-kipi Came to Be
https://www.michigan.gov/-/media/Project/Websites/egle/Documents/Programs/GRMD/Catalog/04/GIMDL-GGKITCH.PDF?rev=4ee5117903ef474b8269922ee52fa4bb

"....But Nature, never satisfied with what she builds, was busily at work destroying the limestone she had made, to fashion something else. Pure limestone is readily soluble and some of the Silurian limestones are pure lime. Water entered the cracks, dissolved the stone, made funnel-shaped depressions as it went dissolving down the joints, and made caves and caverns as it dissolved its way along the bedding planes. It made one cave above another with a thin cracked layer of stone between and the tops or roofs of the upper caves fell in, making funnel-shaped surface pits and hour-glass fashion carrying down the sands of the glacial drift through the cracks into the cave below. Two such caves united below the surface, and their unsupported roofs collapsed making one oval depression loading to a crack in the floor. Waters from the lower cave bubbled up through the crack bringing a fountain of sands--lime, sand, shell fragments, and exquisitely formed microscopic shells of ostracods. The water-filled depression became a pool and vegetation gave its emerald color in reflection to the clear waters. Big Spring was born. Dying and dead trees fell into the pool. Nature draped them with verdant mosses--making them eerie things of beauty in their watery graves. The Indians built their village Osawanimiki on the southeastern shore of Indian Lake and dreamed the legends of the wonder spring."



Figure 3.8.17 Kitch-iti-kipi. Photo courtesy of Village of Garden.

https://www.villageofgarden.com/kitchitip

The Anishinaabe tribes have carried their legends of the spring from one generation to the next through storytelling, rarely writing them down. A variety of versions of the Kitch-iti-kipi legend have been shared through the centuries. In 2020, Carole Lynn Hare published a book titled, "*The Legend of Kitch-iti-kipi*," which tells her tribe's version of the story of the spring:

"Many folk tales exist about this natural wonder found deep in the woods of the Upper Peninsula. Many of those stories were admittedly made up by John I. Bellair, a local businessman in the 1920s, in an effort to attract more visitors to the area. But what is the authentic Native American legend of Kitch-iti-kipi? The story in this book has been passed down orally in the author's Native American family for more than one hundred years. Although Natives seldom write down their folklore, the author's great-great-aunt actually did write this legend in a published booklet which was found when she passed away in 1969. That booklet was the author's inspiration for this book. The Legend of Kitch-iti-kipi reads like a Native version of Romeo and Juliet. In it, the deep love between a handsome brave and a young maiden drives a powerful chief to act out his jealousy. The results are tragic for all three."

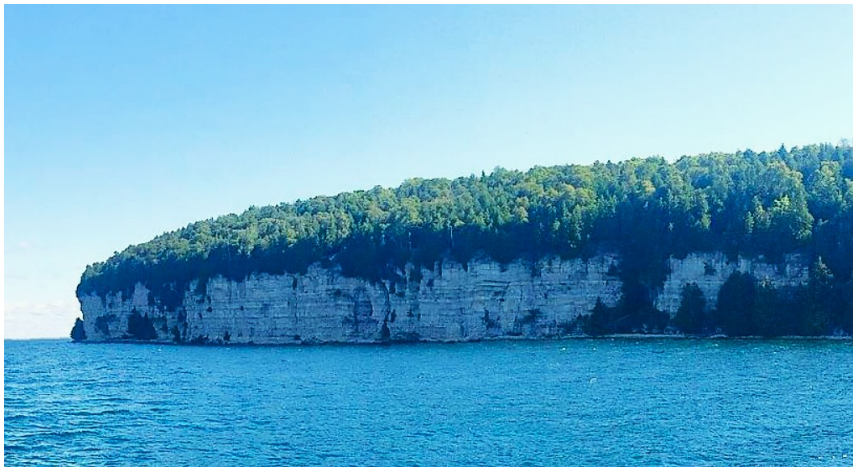


Figure 3.8.18 The Burnt Bluff, and the overlying Manistique group carbonates and Engadine dolostone (dolomite), are a key part of the Niagara Escarpment in Michigan, forming spectacular vertical cliffs which are most visible on the Garden Peninsula, in Fayette Historic State Park at Snail Shell Harbor shown here. Photo by Aaron Joe Pick. Creative Commons Attribution–ShareAlike 4.0. Via Wikimedia Commons.

Great Sulphur Spring

Great Sulphur Spring in Monroe County is located within the Erie Marsh preserve, managed by The Nature Conservancy and the Erie Shooting and Fishing Club. The spring was first described and named by Sherzer in 1900 for its sulfide-rich water:

"*A similar but much larger spring occurs in the Erie marsh near Vienna, N. E. ¼, Sec. 22. It is reached by boat by punting one-half mile through an artificial channel. At a distance the site of the spring is marked by a spot of the most livid green contrasting strangely with the dark waters of the marsh. The general form of the spring is circular as shown in Fig. 8* (Fig. 3.8.19 in this book) *and its diameter is about 160 feet. The bottom is funnel shaped, a marginal strip of 25 to 40 feet shelving off slowly and then making a sudden drop. The bottom where visible through the clear water is covered with a mat of Chara, encrusted with a deposit of sulphur. The sounding line struck bottom at 30 to 33 feet, although it is quite probable that fissures occur leading to greater depths. The outlet from the spring is a small sized river having considerable velocity, a breadth of 33 feet and a depth of one to two feet. The current is so swift that there is some difficulty experienced in entering the spring with a boat. The channel soon becomes broader and proportionately more shallow, and over the bottom of it much sulphur has been deposited. Fragments of tufa are found about the margin of the spring showing that the water is probably rich in calcium carbonate*."

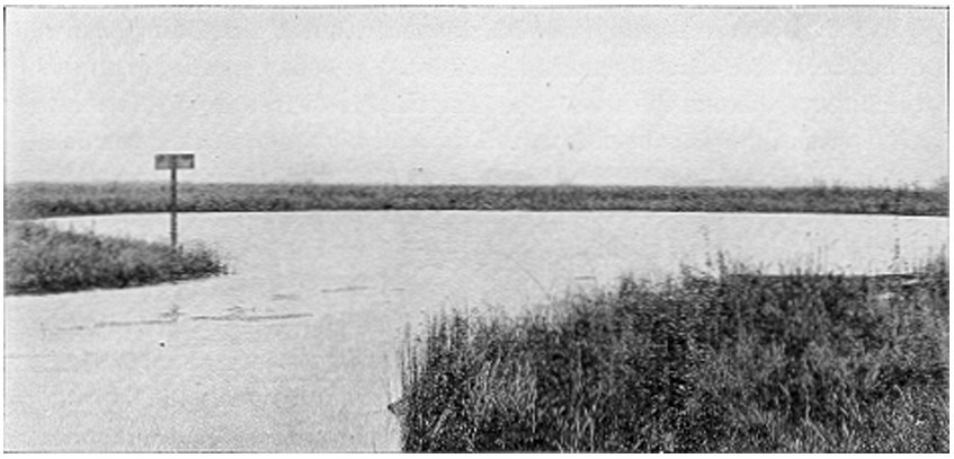


Figure 3.8.19 *Great sulphur spring in the lake marsh near Erie. The figure gives also a view of the inundated margin*. (Fig. 8 from Sherzer, 1900)

Great Sulphur Spring arises from the Salina Group (part of the Silurian-Devonian karst aquifer) within 0.5 km of the Lake Erie shoreline, near the Ohio-Michigan border. The discharge from the spring pond outlet was recorded four times between August 2003 and February 2004 by the USGS: 3.76 cfs, 5.16 cfs, 5.46 cfs; there was no measurable flow on October 1, 2003 (monitoring station USGS-04176632).

Quaternary clay-rich glaciolacustrine sediments are about 35 to 45 feet thick to the west of the marsh and overlie the bedrock which consists of, in ascending order, the Salina Group, the Bass Islands Dolomite, the Detroit River Dolomite, the Dundee Formation, the Traverse Group, and the Antrim Shale (see map in Fig. 3.8.20). In Monroe County, the Detroit River Dolomite has been informally subdivided into the basal Sylvania Sandstone and the upper Detroit River Dolomite. Because the county is on the southeastern rim of the Michigan Basin, the bedrock subcrops are oldest in the southeast part of the county and youngest in the northwest. Bedrock units strike northeast-southwest and dip about 50 ft/mi to the northwest and the center of the Michigan Basin. All of the bedrock units are fractured, and sinkholes are present in the southwestern part of the county (Nicholas et al., 1996).

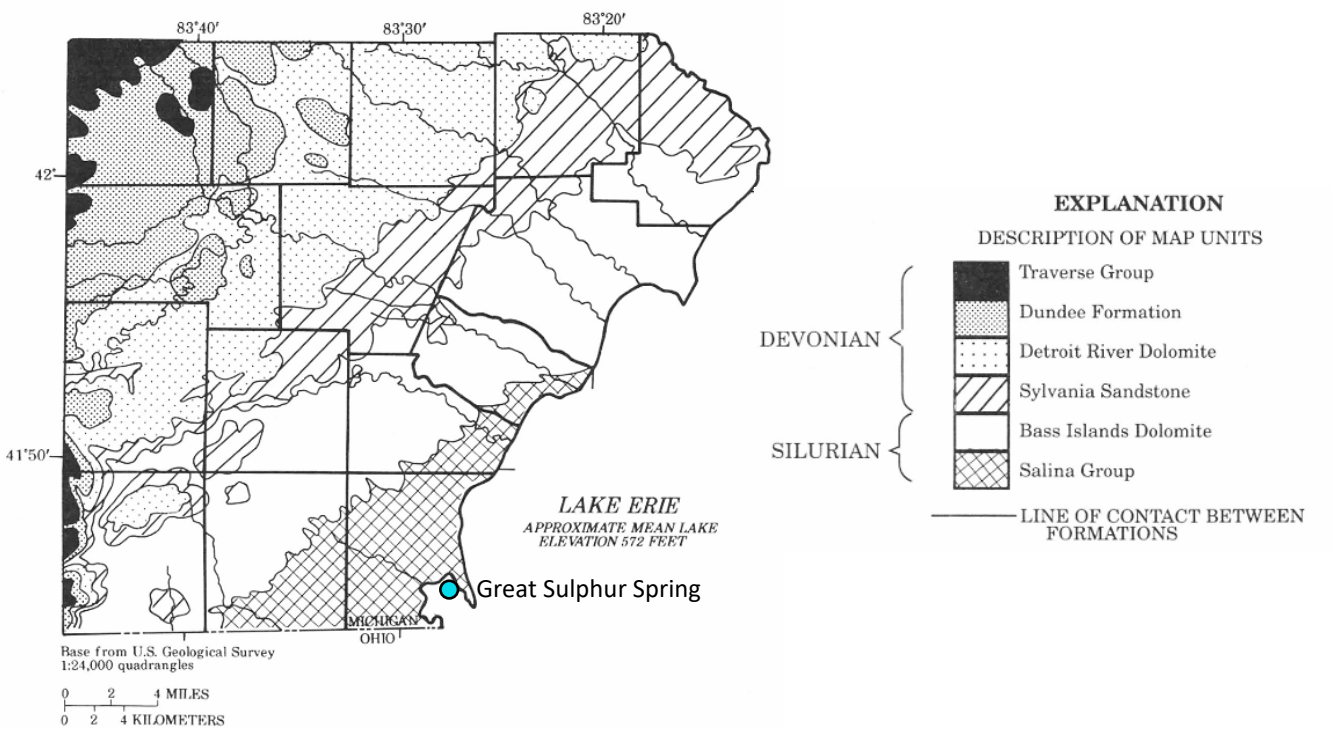


Figure 3.8.20 Bedrock geology of Monroe County, Michigan. From Nicholas et al., 1996.

The karstic and confined nature of the Silurian-Devonian carbonate aquifer were vividly described by Sherzer in 1900 as illustrated with the following excerpts and photograph in Fig. 3.8.21:

It is very probable that there exists throughout this region a series of underground galleries and chambers, but probably of no great dimensions. The openings are not known to show air currents, which would probably be the case if they communicated with extensive caverns. These subterranean channels seem to extend below the level of the lake and hence must have been created when the land stood at a higher level. The water which enters these sinks supplies the great springs, which are found to the east along the lake shore. Some of it also very probably reappears in the artesian wells of the region and some of it may reach the lake directly, without coming to the surface.

There is evidence that these underground waters are inhabited by a special fauna similar to that found in caverns. Sometime in the 70's there were pumped from a well in the Summerfield township, three small fish which are said to have showed no trace of eyes. The well was eighty to ninety feet deep and stands upon land now belonging to J. Cosgroy. The fish were seen swimming in a pail by Mr. Ezra Lockwood, who gives the length of the largest as about one and one-half inches, dark brown in color, slender in form, "shaped like a mullet" but with much enlarged paired fins and the dorsal fin extending to the tail. A similar find was reported from Utica, north of Detroit, in Macomb county, some two years ago. Two larger fish were obtained from a well, neither of which according to the owner, showed any trace of eyes. Very unfortunately a cat made a meal of them before they could be secured by the writer. Similar discoveries may be expected in the future and it is to be hoped that the material will fall into friendly hands.

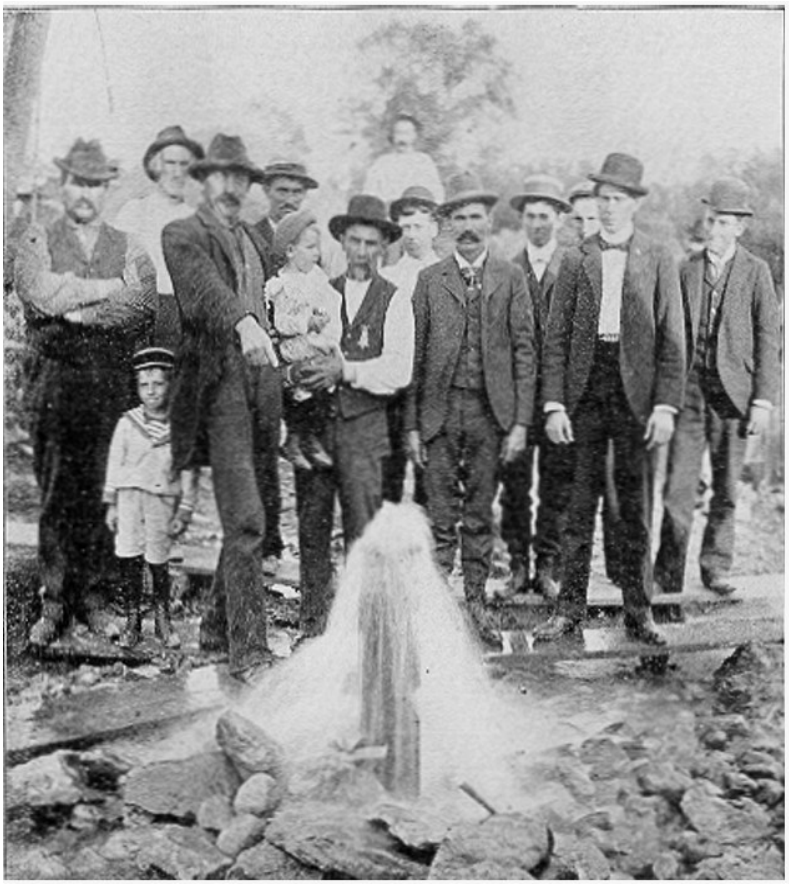


Figure 3.8.21 Text and photograph from Sherzer (1900): "*Gushing well struck upon Otter Creek, September, 1899. Residence of Edward Sharkey, 2 miles southeast of LaSalle. The heaviest flow of artesian water known in this part of the state was struck in September, 1899, upon the south bank of Otter Creek, about 2½ miles southeast of La Salle, at the residence of Edward Sharkey. The well is 51 feet deep and 6 to 8 feet in the rock, having been drilled with a three inch drill. September 4th water was struck and began to flow, slowly at first and then with greater strength. A wooden pump-stock was driven into the hole and projected four feet above ground. Through this the water was thrown with force and to a considerable distance, latterly through 10 one and one-half inch holes bored in the side. Thus prevented from escaping freely the water began to rise through the clay around the mouth of the well and at a distance of 25 to 30 feet from it, forming a shallow lake about 100 feet across. By this time Mr. Sharkey became alarmed for the safety of his house and the pump-stock was withdrawn, an eight inch pipe being inserted in its place. At the time of the writer's visit a very rapid stream two feet broad and four inches deep was flowing from the well to Otter Creek.*"



Figure 3.8.22 The water of Great Sulphur Spring is anoxic (lacks oxygen) and typically maintains a cold temperature around 54 degrees Fahrenheit and pH of 6.3. It hosts rare lineages of *Archaea* and *Epsilonproteobacteria* that thrive on sulfur metabolism, creating the turquoise-colored water and white-crusted "calcified algae" formations. Photo by Patrick Hall, Mar 31, 2026. Courtesy of AllTrails. Available at https://www.alltrails.com/poi/us/michigan/erie/great-sulphur-spring

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