# U.S. Geological Survey Karst Interest Group Proceedings, St. Petersburg, Florida, February 13-16, 2001


**Water-Resources Investigations Report 01-4011**

**Department of the Interior**
**U.S. Geological Survey**

## PREFACE

Karst aquifer systems occur throughout the United States and it’s territories. The complex depositional environments that form carbonate rocks combined with post-depositional tectonic events and the varied climate in which these rocks occur, result in unique systems. The dissolution of calcium carbonate and the subsequent development of distinct and beautiful landscapes, caverns, and springs has resulted in some karst areas of the United States being designated as National or State Parks and even commercial caverns. Karst aquifers and landscapes that form in tropical areas, such as the north coast of Puerto Rico differ greatly from karst areas in more arid climates, such as central Texas or South Dakota. Many of these public and private lands contain unique flora and fauna associated with the water. Thus, multiple Federal, state, and local agencies have an interest in the study of karst areas.

Carbonate sediments and rocks are composed of greater than 50 percent carbonate, $CO_3$, and the predominant carbonate mineral is calcium carbonate or limestone, $CaCO_3$. Unlike terrigenous clastic sedimentation, the depositional processes that produce carbonate rocks is complex as it involves both biological and physical processes. These depositional processes have a major impact on the development of permeability of the sediments. Additionally, carbonate minerals readily dissolve and precipitate depending on the chemistry of the water flowing through the rock, thus the study of both marine and meteoric diagenesis of carbonate sediments is multidisciplinary. Once the depositional environment and the subsequent diagenesis is understood, the dual porosity nature of karst aquifers presents challenges to scientists attempting to study ground-water flow and contaminant transport.

Many of the major springs and aquifers in the United States occur in carbonate rocks and karst areas. These aquifers and springs serve as major water supply sources and as unique habitats. Commonly, there is competition for the water resources of karst aquifers, and urban development in karst areas can impact the ecosystem and water quality of these aquifers.

During the November 1999, National Ground-Water Meeting of the U.S. Geological Survey Water Resources Division, the idea for developing a Karst Interest Group evolved. As a result, he Karst Interest Group was formed in 2000. The Karst Interest Group is a loose-knit organization of U.S. Geological Survey employees devoted to fostering better communication among scientists working on, or interested in, karst hydrology studies.

The mission of the Karst Interest Group is to encourage and support interdisciplinary collaboration and technology transfer among U.S. Geological Survey scientists working in karst areas. Additionally, the Karst Interest Group encourages cooperative studies between the Water Resources Program Districts and National Research Program Offices, between the Water Resources Program Districts and other U.S. Geological Survey Biological Program Science Centers, Mapping Centers, or Geologic Program Research Centers, and between the U.S. Geological Survey and other Department of Interior Agencies, and University researchers.

These proceedings result from the first effort to bring together U.S. Geological Survey scientists with other Department of Interior scientists and managers and University researchers interested in karst hydrology. The presentations cover: karst ecosystems, natural resource development in karst areas, the geologic framework of karst sytems, aquifer hydraulics in karst systems, programs in the Department of Interior that involve karst, numerical modeling in karst, cave and spring species and habitats, geochemistry of karst systems, geophysical methods in karst, contaminant transport in karst, and tracers in karst.

The planning committee for this workshop was: Zelda C. Bailey, Alan W. Burns, Norman Grannemann, Eve L. Kuniansky, Randall C. Orndorff, Albert T. Rutledge, Ann B. Tihansky, Patrick Tucci, Peter W. Swarzenski, and Stephen J. Walsh. We sincerely hope that this workshop promotes future collaboration among scientists of varied educational backgrounds and improves of our understanding of karst systems in the United States and it’s territories.

The extended abstracts of U.S. Geological Survey authors were reviewed and approved for publication by the U.S. Geological Survey. Articles submitted by University researchers and other Department of Interior Agencies did not go through the U.S. Geological Survey review process, and therefore may not adhere to our editorial standards or stratigraphic nomenclature. The use of trade names in any article does not constitute endorsement by the U.S. Geological Survey.

The workshop and proceedings were sponsored by Bonnie A. McGregor, Eastern Regional Director, and co-hosted by Wanda C. Meeks, Southeastern Regional Hydrologist, and Lisa L. Robbins, Director of the Center for Coastal Geology, all of the U.S. Geological Survey. The assistance of Carolyn Casteel, U.S. Geological Survey, in the final proceedings layout and printing was invaluable.

Eve L. Kuniansky
Karst Interest Group Coordinator

## PREFACE

## CONTENTS

## INTRODUCTION

The first U.S. Geological Survey, Karst Interest Group (KIG) Workshop was held in St. Petersburg, Florida, on February 12-16, 2001.Technical sessions were conducted on February 13 and 14, and a field trip to local area karst features took place on February 15, 2001. Scientists and researchers from the U.S. Geological Survey, universities, as well as other Federal, state, and local agencies met to learn and discuss topical subjects related to karst systems and hydrology. Among the topics covered during this important first workshop were discussions about the geologic framework and hydrogeologic characterization of karst; various geophysical techniques used in karst landscapes; geochemical characteristics of karst; the ecological role of karst wetlands and the various biological communities typical of karst habitats; employing numerical modeling to simulate flow as well as transport in karst; monitoring contaminant and tracer movement in karst; and the protection and management of karst resources.

This workshop was particularly important because it formally established the Karst Interest Group within the U.S. Geological Survey, and it represented the first organized effort to provide scientists a forum to communicate and share their experiences in studying and investigating karst systems. We hope the workshop will serve as a springboard promoting future collaboration among scientists and researchers to improve our understanding of karst systems in the United States and its territories.

This proceedings volume is a compilation of abstracts and papers from oral and poster presentations given at the workshop. The last paper included in the volume is the field trip guide. The papers contributed to these proceedings reflect the efforts of over 60 scientists working in karst systems.

## ACKNOWLEDGEMENTS

The planning committee for the workshop would like to thank all those authors who contributed to this proceedings volume, and to the technical session moderators for their time and participation.

The planning committee would like to extend our appreciation to Bonnie A. McGregor, Eastern Regional Director, U.S. Geological Survey, who sponsored the workshop and this proceedings volume. Appreciation is also extended to Wanda C. Meeks, Southeastern Regional Hydrologist, and Lisa L. Robbins, Director of the Center for Coastal Geology, both of the U.S. Geological Survey, for co-hosting the workshop. This publication, which will serve as an excellent reference for all workshop participants, is a direct result of their sponsorship and support.

Finally, the planning committee would like to thank the following individuals for their assistance: Ann B. Tihansky, Tampa, Florida Subdistrict, U.S. Geological Survey, who planned the karst field trip and supplied the artwork for the KIG logo. Erica J. Heeg and Rita S. Bowker, St. Petersburg Office, U.S. Geological Survey, who helped with registration and meeting logistics

# Karst Landscapes and the Importance of Three Dimensional Data in Protection of Cave and Karst Resources


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## Abstract

Karst and similar landscapes are found in a wide range of biogeographic classes. In the U.S. for example, Everglades, Mammoth Cave, and Hawaii Volcanoes National Parks have little in common - except karst or pseudokarst, and a cultural past (even though these are very different). This diversity of geologic settings makes karst difficult to categorize and work with when designing a national program such as the recent NPS-USGS Geo-Indicators effort. A GIS-based approach with multiple datalayers is the only sane way to understand and convey the many relationships, in X, Y, and Z axes, between component ecosystems and cultural resources within karst and pseudokarst landscapes. Obviously, karst and cultural landscapes cross modern political as well as biogeographic boundaries. Here again, three-dimensional data are the foundation for understanding similar to that in anatomy and physiology: structure and function. In understanding where the most vulnerable "pressure points" exist within karst landscapes, we can target landscape-scale ecosystem management to greatest effect. USGS and the National Cave and Karst research Institute could play an extremely significant role in cave and karst management on a national scale beyond NPS or other agency boundaries via cooperative management of three-dimensional karst datasets analogous to programs in several states.


## Karst, Pseudokarst, and Non-Karst Caves

Karst landscapes have characteristic morphology and underground drainage including caves created largely by chemical solution. This is typical in carbonate, sulfate, and chloride rocks, but can also occur in less soluble rocks such as quartz diorite (Jennings 1971) or quartzite (White et al 1966). Karst accounts for approximately 15 % of the earth's land surface, and that in carbonate rocks is most prevalent (White and White 1989). Pseudokarst landscapes may have morphology similar to karst, and often have subterranean drainage and caves, but porosity is created via completely different means. Pseudokarst typically develops in lava, unconsolidated sediments or volcanic ash, talus, ice, and permafrost (Kempe and Halliday 1997). Pseudokarst is most highly developed in basalt flows with lava tubes (Halliday 1960). Ironically, water may or may not penetrate lava tubes, but surface runoff often quickly sinks to resurge at springs via crack systems in the lava fields. Despite the differences in origin, volcanic and solutional caves have environmental similarities that have led to analogous faunas (Peck 1973).

Not surprisingly, ecological similarity means that some of the same threats apply, such as impacts from road construction (Halliday 1996), disturbance to cave fauna by visitors, sinkhole dumps (Kambesis 2000), raw sewage (Halliday 2000), pharmaceuticals, household pesticides (Lao and Gooding 2000) and other sources of contamination. Caves in talus can also be important habitat as explained in Bat Conservation International's "Bats of Eastern Woodlands" report to the U.S. Fish and Wildlife Service (Keeley et al 2001). As an example, cave visitation was recently limited in deference to a maternity colony of Townsend's Big-Eared Bats at Pinnacles National Monument, California. Caves in talus may be part of a pseudokarst landscape, or may be isolated features more appropriately classed as non-karst caves along with erosional sea caves. Non-karst caves, like their counterparts in karst and pseudokarst landscapes, can also be significant geological features important as wildlife habitat or cultural sites.

## NPS-USGS Geoindicator Effort

Beginning in March of 2000, the National Park Service (NPS) and U.S. Geological Survey (USGS) began a joint effort to inventory geologic processes potentially influenced by


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human activities in national parks (Higgins and Wood 2001). The foundation selected for this work is a set of 27 earth processes known as geoindicators that may exhibit important changes within a century, and which could be significant in assessments of environmental stability and ecosystem integrity. This excellent set of geoindicators was developed by the International Union of Geological Science's Commission on Geological Sciences for Environmental Planning, and is detailed in Berger and Iams (1996). One of the geoindicators is Karst Activity, which focuses on carbonate and other soluble bedrock terrain. Rather than focus this geoindicator exclusively on solutional karst features, the NPS will broaden the scope to encompass pseudokarst landscapes and isolated non-karst caves. In a similar vein, the national karst atlas proposed by Jack Epstein and Randall Ordnorff of USGS will include pseudokarst landscapes, building on the work of Davies et al (1984). These are important steps toward inclusion of caves and landscapes that function in similar ways despite different origins. Though almost all speleologists were introduced to the field via carbonate caves, general acceptance of a broader array of landscapes and cave types appears imminent. In 1997 the International Union for the Conservation of Nature decided with some trepidation to not include lava caves and other forms of pseudokarst in its publication titled "Guidelines for Cave and Karst Protection." However, the possibility of future inclusion was discussed (Hamilton-Smith 1997). Though the Karst Waters Institute (KWI) does not discuss caves and karst of non-solutional origin in its basic explanation of karst and karst ecology, the Koloa Lava Tube System in Hawaii was included in KWI's 2nd Annual Top Ten List of Endangered Karst Ecosystems (Belson 1999). As well, the excellent map titled "Subterranean Biodiversity of Karst in the United States" by KWI's Dave Culver includes areas with lava tube biota (Culver 1999). Finally, a worldwide web search on pseudokarst reveals that from Texas to Tasmania, pseudokarst is considered significant enough to map alongside classic karst areas.

Returning to the Karst Activity Geoindicator, there are other relevant areas in need of amplification and inclusion besides the expansion beyond solutional karst described above. Perhaps due to the geological orientation of the originators, emphasis is given to water-rock interactions. It is important to note however that even caves of solutional origin long ago may be dry, and preserve extremely significant archaelogical resources (Watson et. al. 1969, 1974). As an indication of how rich cave passages can be in cultural resources, more than 9000 items have been recorded over the past 8 years in approximately 2.5 miles of upper level trunk in Mammoth Cave (Crothers, Swedlund, and Ward 2000). Among pseudokarst caves, lava tubes have a particularly high frequency of cultural artifacts (Ron Kerbo pers. comm.). The NPS /USGS inventory of geologic processes in national parks will consider subterranean cultural artifacts as both resources to be protected, and as indicators of distortion to the cave environment. Decay of in situ ancient organic artifacts may indicate changes in the cave environment caused by entrance manipulation. Also preserved very well in caves are paleontological and paleobotanical resources. These natural and culturally deposited remains can be used for ecological restoration in caves and on the surface (Toomey et. al. 1998, Olson 1998). Equally important along with cultural resources are cave biota. For example, terrestrial cave ecosystems historically provided refuge for literally billions of bats across the continent. Given the appetite bats have for insects (Whitaker 1993), many species had, and one day could again, have a significant controlling effect on insect populations affecting croplands, orchards, forests, savannas, and prairies. The aquatic cave ecosystem species, like all others, are important in their own right, but community level monitoring data are also an important component of and complement to direct water quality monitoring. Modification of dry cave entrances, creation of open artificial entrances, drilling through passages, damming of base level surface streams, and land-use related impacts in recharge areas can severely affect habitat conditions and therefore biota in cave ecosystems. These effects can ramify into neighboring ecosystems in the karst landscape.

One final point I want to make regarding the Karst Activity Geoindicator is that exemplary caves within karst landscapes are economically significant, especially in terms of tourism. Caves, as perhaps our favorite part of the karst landscape, are regional economic engines. For example, Mammoth Cave National Park had a statewide economic impact of over $116 million in 1994 (Atwood 1995). This economic value, including tax revenues generated, helps to make karst landscapes *worth protecting* to a broader range of people. Even though ecosystems tend to run themselves, with the ever- increasing human population and associated total impact, we must carry out compensatory ecosystem management. And as we all know, ecosystem management is funding dependent. This interdependence of our economy upon intact ecosystems, and ecosystem management upon a strong economy is little appreciated by business and resource managers respectively. Arguably, broad realization of this relationship between ecosystems and economies is most vital in karst and pseudokarst areas. In most landscapes, key resources can be specified with a set of XYZ coordinates, with little possibility for confusion or overlooking key resources. However, in karst landscapes it is possible to have natural or cultural surface features at a given set of XYZ coordinates, a terrestrial cave community or an archaeological site at the same XY but different location on the Z axis, and then cave aquatic habitat with historical objects such as lanterns and bottles at a still lower point along the same Z axis. This is an extreme example, and I use it to make a point, but it is also important to bear in mind that superposition is not a precondition for potentially impacting subterranean resources. Sites far removed but down gradient can be profoundly impacted. The best analogy I can draw is that karst and pseudokarst landscapes are like playing three-dimensional chess instead of the usual mind boggling two-dimensional game. In moving from general to specific, I will use Mammoth Cave National Park as an example.

## The Mammoth Cave Case

Due to high biological diversity, and the most extensive known cave system in the world, a significant portion of South Central Kentucky's world-class karst landscape with its component ecosystems was designated as the Mammoth Cave Area Biosphere Reserve (MCABR) in 1990 by the United Nations Educational and Scientific Organization. Subsequently, an expansion of the reserve was approved in 1996. The central goals of MCABR programs are cooperative conservation of biodiversity, development of a sustainable economy, and environmental education. The first two of these goals are interdependent and require a thorough understanding of the complexity and limitations of both the natural and human ecosystems in the area. The third goal, environmental education, is an essential tool in successfully accomplishing the first two goals.

The conduit aquifer of the South Central Kentucky Karst is a great economic asset to the region because of the caves that are the foundation of the tourism industry. At the same time, this karst landscape is an economic hindrance because of its vulnerability to groundwater pollution, and the greater cost of infrastructure due to unstable bedrock. The path to a sustainable economy in such a complex, vulnerable and multi-layered landscape must be based upon an understanding of how surface and subsurface ecosystem~~s~~ relate to each other, and how these interrelationships may affect the human environment. The first map of Mammoth Cave intended to clarify surface-cave relationships was completed in 1908 by Max Kaemper of Berlin,Germany. Because this project was driven by property issues, the Kaemper map was kept confidential until discovered in the 1970s, but it is still used by researchers today. Subsequently, five scientifically-driven cartographic efforts spanning five decades have contributed toward understanding this complex landscape:

1. A map set titled "The Flint Ridge Cave System", was produced by Roger Brucker and Denver Burns in 1964. This folio cave map included surface topography, and incorporated text on history and geology.

2. A series known as the "Map Card" was first published by the Cave Research Foundation in 1973, and displayed 169 miles of the Flint-Mammoth Cave System within the park. A topographic overlay adapted from USGS quadrangles allowed understanding of surface-cave relationships. The most recent in the series was published in 1993, and shows 342 miles of the Mammoth Cave System in and near Mammoth Cave National Park with a similar topographic overlay.

3. The next step toward a higher level of understanding the regional karst landscape was a map titled "Groundwater Basins in the Mammoth Cave Region", published by Jim Quinlan and Joe Ray in 1981. This monumental work based upon dye tracing and measuring water well depths helped park managers to realize the importance of working with neighbors to help protect park resources. On this map, drainage basin boundaries, groundwater level contours, and probable underground flow routes in and near the park were displayed along with line plots of cave passages in and beyond park boundaries. This

map has supported protection of the endangered Kentucky Cave Shrimp by justifying creation of a regional sewage treatment facility, negotiation of runoff filtration and spill retention structures along Interstate Highway 65, and cleanup of sinkhole dumps in the Pike Spring Basin (Olson 1996, Olson et al 1999).

4. Cartographic delineation of the cave systems within the Biosphere Reserve is a work in progress that none of us alive today will likely see completed. To date, over 535 miles of cave passages have been surveyed within the MCABR, and these were assembled on a single map titled "Caves of the Dripping Springs Escarpment" by Don Coons in 1994. The simultaneous display of surface topography along with line plots of cave passages is a significant contribution toward our understanding of the three-dimensional landscape within the MCABR.

5. An ongoing effort focussed on hydrology is the Karst Atlas Series of maps being developed in Kentucky by Joe Ray and Jim Currens. The first maps in the series became available in 1998, and are maintained as GIS datalayers in order to facilitate updates. The maps in this series display drainage basin boundaries, surface topography, and probable underground flow routes. Line plots of cave passages are not displayed, but sites with known cave streams are indicated with a symbol.

All of the above works have helped to set the stage for what I will call Karst Landscape Ecosystem Management (KLEM) within the MCABR. This is based upon the concept of landscape ecology, which considers the interplay between component ecosystems on a landscape scale. Examples within karst or pseudokarst landscapes include the relationship between forest, savanna, prairie, or desert and terrestrial cave ecosystems, and the connection between aquatic cave ecosystems and surface counterparts. At a minimum, one platform on which cave survey and resource data, drainage basins, and surface topography can be displayed is needed. Additional relevant information useful in KLEM would include, but not be limited to, vegetation and other cover types indicating land use, habitat types, plus an array of GIS-linked inventory and monitoring datasets. Identification of cave passages with intermittent or perennial streams would also be very useful in evaluating potential risks to aquatic ecosystems. The GIS approach taken with the Karst Atlas Series in Kentucky and the proposed National Karst Atlas is the only rational approach since there are limits to how much information can be displayed at one time without confusion. With such three-dimensional data available for display in the combinations needed, we have the foundation for understanding similar to that in anatomy and physiology: structure and function. In understanding where the most vulnerable "pressure points" exist within the MCABR, we can target ecosystem management to greatest effect.

In the process of assembling information useful for KLEM, it is crucial to manage it in a way that minimizes opportunities for misuse of this information. A sensitive information protection strategy must be a deliberate part of any plan for what and how information will be made available in order to reduce risk to the very resources we seek to protect. Most obviously, cave entrances should not be indicated except in the most general way as on the Karst Atlas Series, and line plots of caves short enough that the entrance location is easily deduced should not be included even on datalayers with (theoretically) restricted circulation. This information must be managed in the same secure manner as archaeological sites or rare species locations. One minimum risk approach that would provide basic information relevant to developers and the public is the Geographic Exploration System (GES) in which images composed of different datalayers can be acquired, but the cartographic data from which it was derived cannot. Such basic map sets would be a boon to environmental educators. Hard copy output from this type of graphic display could be (and maybe should be) used for place mats in area restaurants. This could be part of environmental education efforts within the MCABR that would not place any resource in jeopardy.

As well, GES could become an important tool to use in the economic development decision-making process. Kentucky Geological Survey Division of Oil and Gas Inspectors need to be aware of where cave passages exist both for the protection of caves, and to help well drillers avoid loss of drill bits, or other expensive situations. In the permit review process of the Kentucky Pollution Discharge Elimination System, it is crucial that cave and karst features are known and considered because of the rapid

transit of water and contaminants in karst aquifers. Similarly, quarry operators are better off avoiding underground voids in order to avoid the environmental and economic costs associated with the impact of these operations on endangered species habitat, local water supplies or archaeological sites.
Research grade three-dimensional GIS coverage of the MCABR would have both greater capabilities and vulnerabilities. The ability to view cave passages from any angle is certainly crucial to understanding passage development, but we must insure that these same data cannot be used to acquire and misuse sensitive resource information. Perhaps most crucial, cave surveys and digital elevation models for surface topography may someday be precise enough to identify current or potential entrances. Cave Geographic Information Systems (CGIS) are certainly useful for all branches of research, but information on all vulnerable resources must be kept secure. One recent development designed to facilitate dissemination of cave survey data to those who need it while providing adequate security is the establishment of the Kentucky Speleological Survey (KSS) (Florea 2000). In partnership with the Kentucky Geological Survey (KGS), certain types of cave survey data will be maintained, and made available to the public and government agencies with concurrence of the KSS. Because the data are the property of KSS and not the KGS, they are exempt from Freedom of Information Act requests. This type of arrangement, already established in Virginia, Illinois, and Missouri, could lead to a national program with USGS and the National Cave and Karst Research Institute. This could be accomplished via affiliations between state geological surveys and USGS, or more directly through cooperation with the National Speleological Society and Cave Research Foundation. Recognition of the importance of three-dimensional datasets for KLEM on a national scale by USGS could take cave, karst, and pseudokarst management to a new level. This would be highly significant both within agency-managed units and beyond.

## CONCLUSION

The public in general and planners in particular need to be aware of critical points of wildlife habitat and other natural or cultural resource vulnerability in karst and pseudokarst landscapes plus areas with non-karst caves. Using this approach will reduce the frustration economic developers face when dealing with cave and karst environments. Overall, the potential benefits of GIS supported Karst Landscape Ecosystem Management on local and national scales would far outweigh any potential damage caused by misuse of three-dimensional datasets. This discussion is relevant to all agency-managed areas with karst, pseudokarst, or non-karst caves, and to the far greater lands beyond. USGS is ideally prepared to assist the land management agencies in sustainable stewardship of caves, karst and pseudokarst because its mission is truly national in scope rather than being limited to certain parcels scattered across the country.

Even if the only gain from GIS supported Karst Landscape Ecosystem Management was increased awareness and understanding of regional drainage and cave resources, the effort would be worthwhile. Such increased awareness and understanding could help with matching the highest impact land uses with the least vulnerable sites. With such an approach, we can maximize our chances of realizing an ecologically sustainable economy. This is not yet an urgent political issue, but it will become so with time.

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# The Ecological Role of the Karst Wetlands of Southern Florida in Relation to System Restoration

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## INTRODUCTION

With the recent funding of the Comprehensive Everglades Restoration Plan (CERP), the largest ecosystem restoration program ever attempted, there is a pressing need to be able to detect changes in natural habitats as a result of restoration actions. Human activities, particularly the construction of canals and levees that can either drain or flood wetlands, have affected the natural variability of environmental conditions (Gunderson and Loftus 1993). CERP intends to restore natural hydropatterns to areas that have been damaged by water management. Baseline data on constituent aquatic communities and their ecology are needed before, during, and after the restoration activities commence.

Freshwater fishes and invertebrates are important ecosystem components in the Everglades/Big Cypress system. They operate at several trophic levels in the wetlands, from primary consumers of plant material and detritus to carnivores and scavengers. Factors that influence fish and invertebrate numbers, biomass, and composition therefore affect energy flow through the wetlands. The ecology and life histories of these animals are intimately tied to the hydrology of the wetlands, which is determined mainly by rainfall, but increasingly by water-management practices. Because of the hydrological changes wrought by drainage and impoundment, and the loss of spatial extent and functioning of former wetlands to development (Gunderson and Loftus 1993), there is little doubt that standing crops and overall numbers have declined. Changes to the original ecosystem have also altered the timing and the areas of prey availability to predators. Non-native fishes have colonized natural and disturbed habitats during the past three decades. Non-native fishes have affected native animals through predation, nest-site competition, and habitat disturbance (Loftus 1988) and may divert food-web energy into biomass unavailable to top-level predators.

Aquatic animals in southern Florida wetlands have a variety of ways to cope with environmental variability. These include movements to find refuge from drying habitats in winter and spring, and dispersal away from those refuges with the onset of the wet season (Kushlan 1974, Loftus and Kushlan 1987). This pattern of movements among habitats with fluctuating water depths is common to seasonal wetlands in the tropics (Lowe-McConnell 1987, Machado-Allison 1993). The major natural refuge habitat most-studied by scientists in southern Florida is the alligator hole (Craighead 1968, Kushlan 1974, Nelson and Loftus 1996). Canals and ditches offer a relatively recent but spatially extensive form of artificial refuge for aquatic animals on the landscape (Loftus and Kushlan 1987). In this study, we are studying the function of other types of aquatic refuges in the Everglades.

The Rocky Glades, or Rockland, habitat is a karstic wetland unique to Everglades National Park (ENP) in southern Florida (Figure 1), although similar habitats exist elsewhere in Yucatan, Cuba, and the Bahamas. Approximately half of the original area of this habitat occurs outside of ENP where agricultural and urban development has forever altered its geological structure and ecological function. This region is a high priority for restoration in CERP because it is the largest remnant, short-hydroperiod wetland in the eastern Everglades. That habitat has been disproportionately lost from the ecosystem. Unfortunately, the habitat remaining in ENP has been degraded by water management (Loftus et al. 1992).

In Eve L. Kuniansky, editor, 2001, U.S. Geological Survey Karst Interest Group Proceedings, Water-Resources Investigations Report 01-4011, p. 8-15

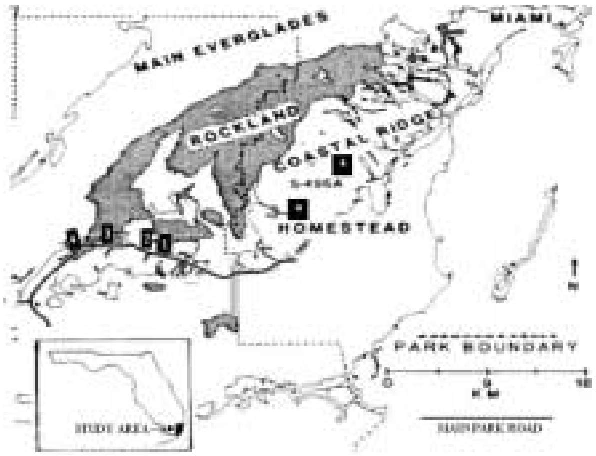



Figure 1. Locations of the study sites within the Rocky Glades and Atlantic Coastal Ridge in southern Florida. The numbers indicate the drift-fence arrays on the main park road, and the stars on the coastal ridge are the well sites with Miami cave crayfish.

The highly eroded karst structure of the Rocky Glades appears to be responsible for the persistence of aquatic-animal communities by offering dry-season refuge in thousands of solution holes of varying depths, (Loftus et al. 1992). Their work was the first to indicate a tight relationship among the biological, geological, and hydrologic components of this region. Loftus et al. (1992) also found evidence that aquatic animals disperse, feed, and reproduce on the wetland surface during the short flooding period, then retreat below ground for periods of months to years. They also reported that several introduced species, particularly the pike killifish (*Belonesox belizanus*), walking catfish (*Clarias batrachus*), Mayan cichlid (*Cichlasoma urophthalmus*), and black acara *(Cichlasoma bimaculatum*) were common in the Rocky Glades (Loftus et al. 1992). Unfortunately, their study was interrupted by Hurricane Andrew and not continued.

In this paper, we report the rationale and results of the first year of a new study in which the primary goal is to define the interactions of the aquatic-animal community with the geologic structure and hydrologic conditions of the Rocky Glades. We are addressing questions that have arisen from past work there. How do composition, size-structure, and recruitment of aquatic animals change during the flooding period? Are the dispersal patterns of animals related to water flow? Are the animals dispersing from the main sloughs to recolonize the Rocky Glades, or is the Rocky Glades a source of animal colonists for the sloughs? Do roadways act as barriers to movement? The objectives of this study segment are:

- Collect baseline ecological data on the epigean aquatic communities in the karst landscape of the Rocky Glades.
- Quantify the direction and degree of dispersal by fishes and invertebrates during the wet season.
- Document the seasonal changes in species composition, size structure, and reproductive patterns of animals on the wetland surface.
- Survey the topography of representative areas of the Rocky Glades, particularly around the sampling sites, to provide depth-distribution data for the simulation model of the region.
- Develop a visual survey method for sampling fish communities in open, rugged terrain to follow community dynamics in the Rocky Glades in the wet season.
- Identify the extent of near-surface voids.

The Atlantic Coastal Ridge is another area affected by urbanization and changing hydrologic management (Figure 1). Aquatic habitats, such as the transverse glades that cut through the Ridge, have been replaced by canals and will not be restored. Ground-water habitats and animal communities may have been less affected. As in karst areas elsewhere, deeper geological formations (>5 m) beneath the Rocky Glades and the Atlantic Coastal Ridge have voids of various dimensions known to house truly subterranean aquatic species (Radice and Loftus 1995, Bruno et al., this volume). These include the Miami Cave Crayfish *(Procambarus milleri)*, known only from a few wells in southern Florida (Hobbs 1971). The composition, distribution, and abundance of other hypogean animals are poorly known. Ground-water withdrawal and saltwater intrusion (Leach et al. 1972), limestone mining, and pollution may threaten these communities before they have been fully catalogued. Elsewhere in the world, such communities are known to be very sensitive to changes in their delicately balanced physical environment. The second goal of this project is to identify the composition, distribution by depth and space, and ecological relations of this subterranean fauna. The objectives of the second study element include:

- Develop effective traps to capture invertebrates and possibly fishes from subterranean habitats.

In Eve L. Kuniansky, editor, 2001, U.S. Geological Survey Karst Interest Group Proceedings, Water-Resources Investigations Report 01-4011, p. 8-15

- Inventory hypogean communities and relate the composition and distribution to environmental factors.

- Collect life-history data for the Miami cave crayfish from a large captive population.

## METHODS

This first project year has been a pilot study to test designs and methods. The study is divided into two elements with several components each.

Element 1 – In the Rocky Glades, we selected four sites along the ENP main road (Figure 1) to test the use of drift-fence arrays to describe directional animal dispersal and community successional patterns in the wet season. The four X-shaped arrays had 12-m wings made of black plastic ground cloth (Figure 2) to direct animals into one of 3 traps that faced east, north, and west, based on the direction that they were moving (Figure 2). The road shoulder formed a barrier to the south of each array. The 3-mm mesh minnow traps were fished overnight for 24 h to provide data on fish relative abundances, movements, and catch per unit effort (CPUE).

Figure 2. Top panel: Array 3 in the Rocky Glades; Bottom panel: three minnow traps facing west, north, and east at the center of the array.

When the wetlands reflooded in June, we collected samples daily for the first two weeks, then reduced the frequency to twice weekly for the next two weeks, and finally made collections once a week until the marshes dried. All animals were identified, weighed, and measured in the lab, and the numbers of animals in each trap on a particular day was compared to the water flow and depth to assess directional movement. We processed samples of fishes and crayfish for stable-isotope analysis as they appeared on the surface to compare with isotope signatures after several weeks and several months aboveground. These data may show whether trophic patterns change when animals begin to forage on the wetland surface. We also saved fishes for analysis of reproductive status to learn whether they are ready to spawn upon emergence onto the surface.

To complement data from the arrays, which are activity traps, we used visual sampling (Loftus et al. 1992; Frederick and Loftus 1993) to estimate fish composition and density on the surface of the Rocky Glades. We set up 24 survey plots, six at each array, that were scanned by binoculars each week in the wet season. Each 4-$m^2$ plot was scanned for 2 minutes, and all individuals seen were counted, and identified to species and size-class.

We began to survey the micro-scale topography of representative areas of the Rocky Glades. The physical characteristics of the sampling sites are required in the simulation model. The surface extent and depth dimensions of solution holes and surrounding marsh surface are measured by standard surveying techniques. Those physical characteristics will be correlated with biological measures of species composition, survival, and density. We have begun to use ground-penetrating radar (GPR) to try to estimate belowground extent of deep solution holes. We hypothesize that the survival of fishes reported by Loftus et al. (1992) in holes in which no standing water was visible, is related to the presence of hidden subterranean cavities connected to the holes.

Element 2: To inventory the hypogean fauna beneath Rocky Glades and the Atlantic Coastal Ridge, we selected a series of existing wells along four east-west transect lines from Miami to Homestead in which to sample routinely. Borehole videography is helping us to select the best wells and depths for sampling. We are using a combination of pumping and filtering ground water from wells to collect copepod crustaceans (Bruno et. al., this volume). We tested several designs for traps to collect larger invertebrates and possibly fishes in wells. We also used GPR to locate areas of high porosity in which hypogean animals might be likely to occur. Drilling of new wells to access subterranean cavities will begin in February 2001, in which a combination of videography and trapping will be used to capture and record animals for study. Any fishes or invertebrates collected will be identified, and then sent to specialists for confirmation.

We used YSI-6000 continuous recorders to measure water-quality in ground water to characterize the environment of hypogean organisms. We collect parameters such as dissolved oxygen, pH and temperature at the surface, middle and bottom of the wells. We will attempt to correlate the environmental variables with species distributions.

We are collating data on the distribution of the Miami Cave Crayfish from wells, and are trapping for it in existing and new wells. Because it is difficult to obtain enough wild-caught animals on which to base a life-history study, we have gained access to a captive population at a local fish farm where we perform monthly assessments of the proportion of males, females, gravid females, and juveniles in the population, their size distributions, size at maturity, fecundity, egg size, and other important life-history parameters.

## RESULTS

The pilot study of the drift-fence arrays provided important inventory and baseline ecological data for the aquatic fauna. The method was successful in meeting the element's objectives. Arrays 3 & 4, west of the Pineland Trail in ENP, flooded in early June 2000. Array 3 had surface water until November, while Array 4 stayed wet until the end of December. Water flow at those arrays was generally east to west, towards Shark River Slough. Arrays 1 & 2, east of the Pineland Trail, did not flood until mid-July, 2000, and dried by November. Water flow was generally west to east at Arrays 1 & 2, towards Taylor Slough.

Animals appeared rapidly on the surface as the wetlands around the arrays reflooded. Fishes and crayfish reappeared in the traps on the same day that the wetlands reflooded, demonstrating the existence of local subterranean refuges. Large catches of several species occurred within a few days of reflooding (Figure 3). The fishes exhibited mass directional dispersal as the wetlands flooded. Although flow velocities were relatively slow in these shallow wetlands, the animals appeared to orient to the flow. Most individuals appeared to follow the flow of water, although a few species, particularly the Everglades crayfish (*Procambarus alleni*) moved mainly against the flow. Subsequent sampling provided data on community-succession patterns as new species appeared in the traps and relative abundances changed. The majority of species appeared at each array within one week of flooding (Figure 4). Non-native and larger-bodied native fishes were slower to appear at the arrays, indicating dispersal from distant refuges.

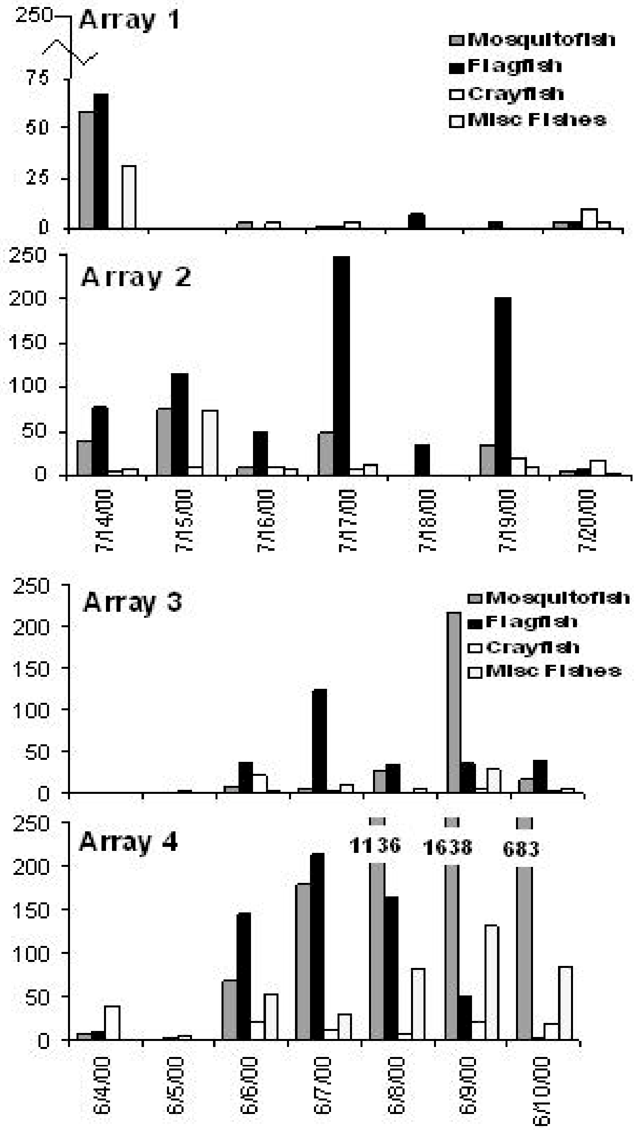


Figure 3. Appearance of fish and crayfish at the arrays shortly following marsh flooding.

We documented the onset of recruitment using the size-structure data and the visual-sampling data. All fishes emerging onto the surface were adults that began reproducing with one or two weeks. Small juveniles appeared in the wetlands within a month of reflooding (Figure 5). In the visual plots, mosquitofish (*Gambusia* *holbrooki*) were most visible, probably because they are in constant motion. Sedentary, cryptic species were more difficult to observe.

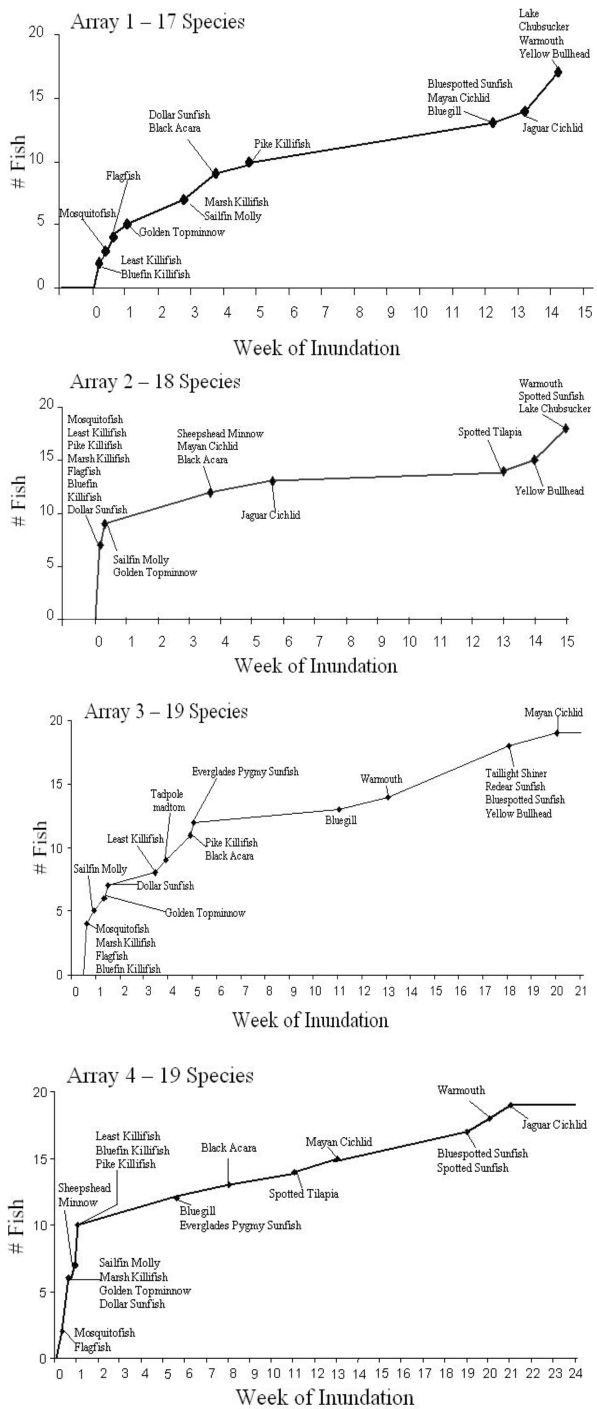



Figure 4: Succession of fish species at the arrays. Cichlids, tilapia, and pike killifish are non-native species.

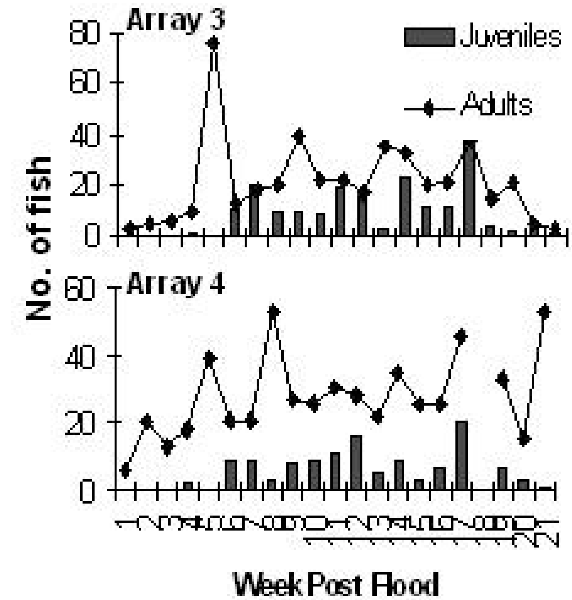
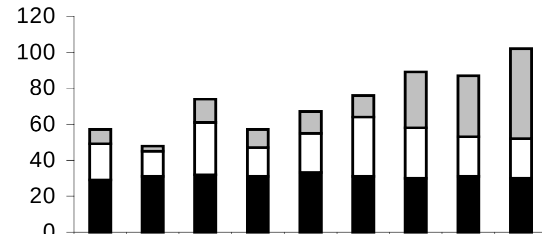



Figure 5: Visual-sample data from arrays 3 & 4 showing the pattern of juvenile recruitment.

In the second element of the pilot study, we have tested the most effective trap designs for capturing animals from groundwater in wells. We selected a 0.5-L plastic bottle trap with an inverted funnel to capture large animals, and a 3.8-cm diameter, 30-cm long perforated PVC tube with a removable core of plastic filter material to serve as an artificial substrate for smaller-bodied animals. These are our standard sampling units for the wells.

Preliminary sampling of wells by pumping and filtering in ENP resulted in more than 10 species of copepods (Bruno et al., this volume) and other crustaceans. Samples from wells on the coastal ridge near Homestead and Miami have produced records from several new locations for the Miami Cave Crayfish (Radice and Loftus 1995), as well as an amphipod that may be new to science. We continue to receive anecdotal reports from local residents of blind white shrimp and fish in wells, but have been unable to confirm those reports.

We have collected nine months of data on the size-structure, sex ratios, sexual status, fecundity, and size of maturity for the Miami Cave Crayfish from the captive population. The proportion of juveniles increased in late spring and again in late autumn because of recruitment in earlier months (Figure 6). The mean size of males and females in this species did not differ, unlike many other crayfish species that exhibit strong sexual dimorphism.




| | Feb | Mar | Apr | May | Jun | Jul | Aug | Sep | Oct |
|---|---|---|---|---|---|---|---|---|---|
| JUVENILE | 8 | 3 | 13 | 10 | 12 | 12 | 31 | 34 | 50 |
| MALE | 20 | 14 | 29 | 16 | 22 | 33 | 28 | 22 | 22 |
| FEMALE | 29 | 31 | 32 | 31 | 33 | 31 | 30 | 31 | 30 |

Figure 6: Proportion of the captive Miami Cave Crayfish population comprised by juveniles, females and males. N of ~30 for females is the sample size sought during each month's random sample, after which that month's sampling ceases.

As might be expected for a species that lives in a fairly constant environment, the data indicate that this crayfish is capable of year-round breeding (Figure 7). We are examining data on the number of eggs produced by variously sized female, the length of time until she releases them into the environment, and their growth in constant temperature in captivity. Those data are being compared to species of surface-dwelling crayfishes in southern Florida.

Figure 7: Gravid female Miami Cave Crayfish with newly extruded egg mass.

The ground-water environment is somewhat variable, as seen in this five-day plot during the dry season of 2000 (Figure 8). Water temperature changed on a diel basis, as did pH, and to a lesser extent, dissolved oxygen. The peak in specific conductance and corresponding rise in pH resulted from a rain event that probably washed soil into the well. Note that the % saturation of dissolved oxygen was very low so any animal in this environment must be adapted to low oxygen tensions.

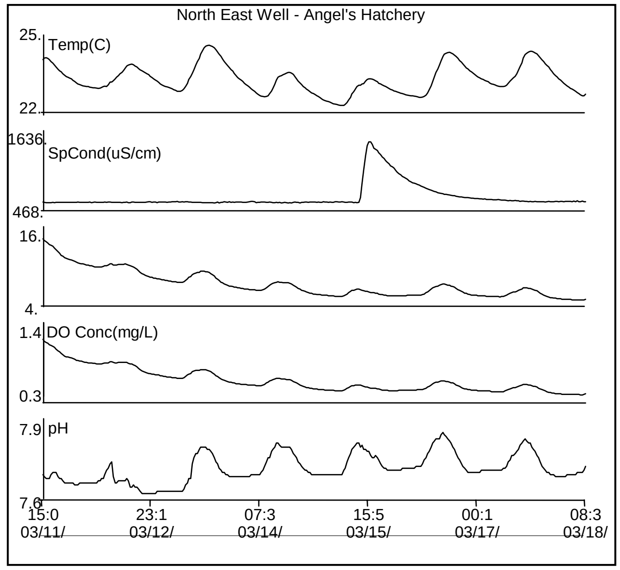



Figure 8: Five-day springtime plot from a well on the Atlantic Coastal Ridge north of Homestead of water temperature ($^{o}$ C), specific conductance ($\mu S/cm^3$), % $O_2$ saturation, dissolved oxygen level (mg/L) and pH.

## RESEARCH NEEDS / APPLICATIONS

The use of drift-fence arrays and visual sampling are new sampling techniques for shallow-water marshes with open, rugged terrain, where other methods fare poorly. We will expand our testing of the sampling characteristics of these methods, and continue to follow animal-community patterns by visual survey and by trapping. Based on the results of further testing, these methods are likely to be accepted by other scientists studying aquatic animals in shallow wetlands. We are now evaluating the results of the pilot project to design and implement a more spatially expansive study for this next year.

The project elements combine to answer questions about the ecological interrelations of surface and subterranean habitats to address how management has adversely affected this region and describe the benefits that restoration will produce. The Rocky Glades and other short-hydroperiod wetlands have been implicated in the decline of nesting wading birds in the Everglades (Fleming et al. 1994). The loss of naturally short-hydroperiod wetlands may have affected the availability of important late wet season/early dry season feeding habitat for wading birds and other predators when wet prairie and slough habitats were too deep for feeding. This project also addresses questions important to understanding the ecology of freshwater communities in southern Florida, and the data will provide more confidence in tools like the ATLSS fish-simulation model to be used in CERP assessments. Data from the present study will provide the information needed to predict the responses of aquatic communities to restored hydrological patterns and to an increase in the spatial extent of the system. The data, and the models that incorporate them, should also help define the reasons behind wading-bird decline as relates to prey availability and abundance.

In addition to the application of these data to modeling, the data collected during these companion studies represent new information about the composition and adaptations of the surface-water and ground-water communities. The baseline data from this study will be essential to future monitoring of the effects of the restoration effort. Preliminary collections from the wells have produced several first records for species for the United States, and potentially new species for science. The data are under review by USFWS, which is considering one species for candidacy for listing under the Endangered Species Act. The interactions of groundwater and surface-water habitats demonstrate the critical and delicate ecosystem linkages that occur on this karstic landscape. The relationships we describe, and the information we collect may help managers of other karstic wetlands, as in Mexico, Belize, and the Bahamas, better protect their resources.

There are several areas for collaboration with other USGS researchers. The occurrence of holes on the Rocky Glades landscape appears to vary spatially, and the availability of holes for refuge is an important factor to include in the spatially explicit ATLSS fish model. We hope to work with experts from the USGS Cartography Division to examine whether remote sensing will allow the estimation of the frequency and distribution of solution holes. We will continue our collaboration with geologists from USGS to further explore the relationship of karst structure to aquatic animal ecology.

## Acknowledgements

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# Geologic Framework of Karst Features in Western Buffalo National River, Northern Arkansas

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## Abstract

Caves, sinkholes, and springs are all common karst features developed in carbonate rocks in and adjacent to the western Buffalo River watershed in northern Arkansas. Recent geologic mapping at 1:24,000 scale, combined with data from cave and spring inventories and dye-tracer studies, provide insight into the geologic framework of the karst features and ground water pathways. These data provide a scientific basis for resource management at Buffalo National River (Mott and others, 1999), a 130-mile-long, river-corridor park established in 1972.

Physiographically, the western Buffalo River flows eastward near the junction of plateau surfaces of the Boston Mountains and Springfield Plateau, where it has eroded a valley network 130-400 m deep. Stratigraphically, the watershed for the western Buffalo River exposes an aggregate sequence of sedimentary rocks that is nearly 500 m thick. These rocks include alternating carbonate, sandstone, and shale formations of Pennsylvanian, Mississippian, and Ordovician age (Hudson, 1998). Limestone and dolomite, typical karst hosts, are important components of five of the eight major map formations exposed within the watershed and are most common in the lower half of the stratigraphic sequence. The most extensive carbonate unit is the Mississippian Boone Formation, a 120-m-thick cherty limestone. Chert content of the Boone Formation in the western Buffalo River watershed is variable but it is commonly less than the >50 percent chert that is typical elsewhere in northern Arkansas, enhancing continuity of karst features throughout the formation. Structurally, rocks the western Buffalo River region lie on the southern flank of the Ozark Dome. They are mostly gently dipping (<5°) but are broken by a series of faults and monoclinal folds that formed during late Paleozoic time (Hudson, 2000). In combination, these structures produce relief of as much as 300 m of a marker datum across the region. Faulting is complex and includes normal, strike-slip, and less common reverse structures. Maximum vertical offset across individual faults ranges from 30 to 120 m. Monoclinal folds that formed over buried faults typically have vertical relief of 20 to 40 m and contain strata that dip 10°-25°, and locally greater. Three main joint sets formed in the region with strike directions of N-S, NE-SW, and WNW-ESE, in order of decreasing frequency.

The stratigraphic distribution of karst can be inferred from an inventory of caves located within the boundaries of western Buffalo National River, although this inventory may exclude caves in upper formations that lie within the watershed but outside the park corridor. Of 96 inventoried caves, 78 percent are within limestone of the Boone Formation, 17 percent are in limestone or dolomite intervals within the Ordovician Everton Formation, and the remaining are in limestone of the lower part of the Pennsylvanian Bloyd Formation. Caves within Boone Formation are distributed throughout its thickness, but entrances are slightly more common within 12 m of either its upper or lower contact. The upper Boone contact is overlain by the 2- to 12-m-thick Batesville Sandstone that is commonly slumped into solution cavities within Boone limestone. These sinkholes serve to concentrate surface waters whose acidity is likely enhanced by oxidation of pyrite that is common within both the Batesville Sandstone and overlying Fayetteville Shale. The basal Boone limestone unconformably overlies sandstone of the Ordovician Everton Formation and hosts the greatest density of springs within the western Buffalo River watershed. This relation illustrates that the Boone Formation is the main karst aquifer for the region. Dye-tracer studies document that some large springs gather recharge from far beyond the watershed boundaries (Mott and others, 1999). Erosion of the Buffalo River valley has left most karst aquifers perched above the current river level and, consequently, their local base-level elevations are controlled by relief across structures. Down-dropped blocks of Boone Formation host both the largest springs and the most extensive cave systems known within this part of the watershed. Boone Formation within one graben lies below river level and here leakage into the Boone is sufficient to completely drain the river into the subsurface during low-flow periods. Increased fracturing associated with structures has probably further enhanced karst dissolution. As an example, several vertical shafts, including the deepest (48 m) shaft known in the park, developed within the fractured limb of one fault-cored fold.

In Eve L. Kuniansky, editor, 2001, U.S. Geological Survey Karst Interest Group Proceedings, Water-Resources Investigations Report 01-4011, p. 16-17

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# Geologic Framework of the Ozarks of South-Central Missouri—Contributions to a Conceptual Model of Karst

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## Abstract

A geologic framework is required to understand the environmental impact of proposed mining of lead and zinc on large springs in the karst area of south-central Missouri. Information about lithologies, faults, joints, and karst features (sinkholes, caves, and springs) contributes to the development of a conceptual model of karst hydrogeology. Conduits and caves along bedding planes and joints provide avenues for ground-water recharge, movement, and discharge. The trend of joints was studied to determine if they controlled the orientation of cave passages and conduits. The data show that cave passages are curvilinear and do no correlate well with measured joint trends. Instead, stratigraphy, bedding-plane dip, and local base level affect conduit and cave development. The majority of caves in south-central Missouri have developed within stromatolitic dolomite horizons beneath sandstone beds. It is thought that the sandstone beds act as confining units allowing artesian conditions and mixing to occur beneath them, thus, enhancing dissolution. Joints and the high primary porosity of the stromatolitic dolomite beds form openings in the bedrock that initiate solution. Where a solution-widened joint intersects a bedding plane, lateral movement of ground water is controlled by the bedding plane.

## INTRODUCTION

Within bedrock regional aquifer systems in karst, the geologic framework provides critical information on the geologic boundary conditions. This paper presents results and observations related to the karst system of south-central Missouri acquired by geologic mapping, fracture analysis, cave and conduit mapping, and other related studies in and around the Ozark National Scenic Riverways, south-central Missouri. These observations and results provide the basis for a conceptual model of this complex karst system.

Geologic mapping at scales of 1:24,000 and 1:100,000 is being done in a part of the Ozark Plateaus in south-central Missouri, USA, to understand karst hydrology, especially relating to water quality and land-use issues on public lands. The study area is located in the Current River and Eleven Point River drainage basins and includes parts of the Ozark and Eleven Point National Scenic Riverways, the Mark Twain National Forest, several state forests, and some private lands (fig. 1). This area is characterized by many large springs, losing and disappearing streams, caves, and sinkholes. The terrain consists of steep-sided rolling hills and valleys, and entrenched, meandering streams; altitudes range from 135 to 400 m and the average relief is 120 to 150 m. The rocks in the study area are Upper Cambrian and Lower Ordovician dolomite,

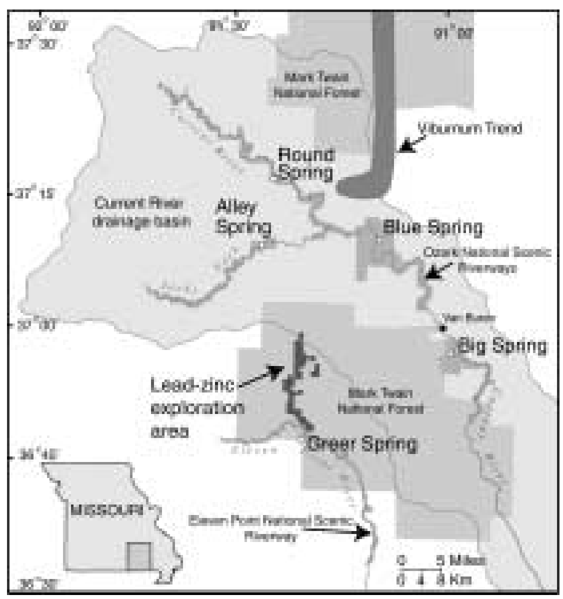


Figure 1. Index map of south-central Missouri showing Federal lands, mineral exploration area, large springs, and major rivers.

In Eve L. Kuniansky, editor, 2001, U.S. Geological Survey Karst Interest Group Proceedings, Water-Resources Investigations Report 01-4011, p. 18-24

sandstone, limestone, shale, and chert, which overlie Middle Proterozoic rhyolite and granite.

The world's largest lead-zinc mining district, the Viburnum Trend, lies on the northern fringe of the study area. Exploration for similar deposits has been carried out in the Mark Twain National Forest near Greer Spring (fig. 1). Applications for permits to conduct mineral exploration in public lands have been requested by private industry in the past few years. Federal and State agencies are concerned about the environmental impact of exploration and potential mining activities on natural and recreational resources. These competing interests have generated a need for detailed geologic and hydrogeologic studies in order to provide data for informed land-management decisions. Geologic studies that identify karst, stratigraphy, and structural features contribute to the understanding of how ground water is transported.

## GEOLOGIC AND HYDROGEOLOGIC SETTING

About 750 to 900 m of gently dipping Upper Cambrian and Lower Ordovician dolomite, sandstone, limestone, shale, and chert unconformably overlie Middle Proterozoic rhyolite and granite (fig. 2). Dolomite is the dominant rock type. Of the Upper Cambrian and Lower Ordovician rocks, only the Potosi Dolomite and younger units are exposed in the study area. Middle Proterozoic basement rocks are exposed as knobs that protrude into the Paleozoic section as high as the Gasconade Dolomite. Caves investigated for this study occur in the Eminence and Gasconade Dolomites, and the Roubidoux Formation. The Eminence Dolomite is a massive to thick-bedded, medium- to coarse-grained, light-gray, locally cherty dolomite with an intercalated sandstone bed in the eastern part of the study area. The Gasconade Dolomite contains a basal interbedded sandstone and dolomite unit, the Gunter Sandstone Member, overlain by medium- to thick-bedded, fine- to coarse-grained, light-gray dolomite with several cherty horizons. The Roubidoux Formation is interbedded fine- to coarse-grained, poorly sorted sandstone, thin- to medium-bedded, fine- to medium-grained dolomite, and chert. All three formations contain stromatolitic dolomite.

The Ozark Plateaus Province is a large structural dome. In southeastern Missouri, strata generally dip gently to the southeast toward the Mississippi embayment. Locally, strata dip steeply

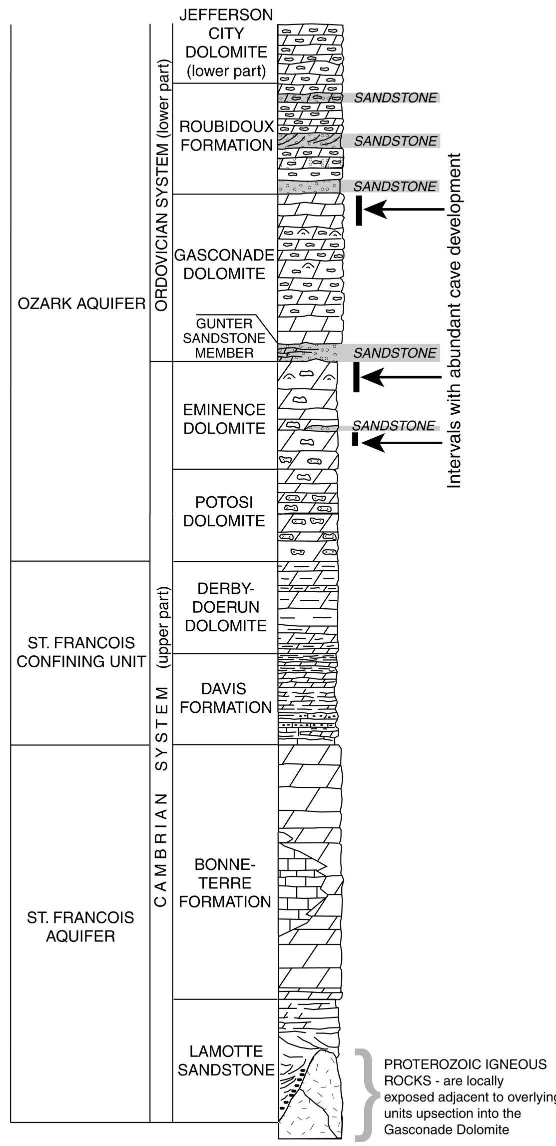


Figure 2. Stratigraphic and hydrogeologic units of south-central Missouri showing stratigraphic intervals with abundant cave development and highlighting sandstone horizons.

away from Middle Proterozoic knobs and near fault zones. Faults are generally steep and most have a northwest or northeast trend (fig. 3). Many faults in the Paleozoic rocks are aligned with Middle Proterozoic basement faults indicating that these faults may be reactivated. Faults with probable strike-slip motion have been identified in the study area by stratigraphic offset and the occurrence of fault breccia (McDowell, 1998; McDowell and Harrison, 2000; Orndorff and others, 1999; Weems, in press). Vertical joints in the Upper Cambrian and Lower Ordovician rocks occur in two dominant sets, $340^{o}$-$0^{o}$ and $70^{o}$-$90^{o}$. The general trends of faults do not parallel these regional joint sets.

Upper Cambrian and Lower Ordovician strata form three geohydrologic units; two aquifers separated by a confining unit (Imes, 1990) (fig. 2). The lower aquifer, the St. Francois, is 30 to 180 m thick and consists of the Lamotte Sandstone and Bonneterre Formation. Overlying the St. Francois aquifer is the St. Francois confining unit (90-110 m thick) formed by shale, dolomite, and limestone of the Davis Formation and Derby-Doerun Dolomite. The upper aquifer, the Ozark (as much as 300 m thick), consists of the Potosi, Eminence, and Gasconade Dolomites, the Roubidoux Formation, and the Jefferson City Dolomite. The Ozark aquifer is the primary source for springs and streams and is used for domestic water supply.

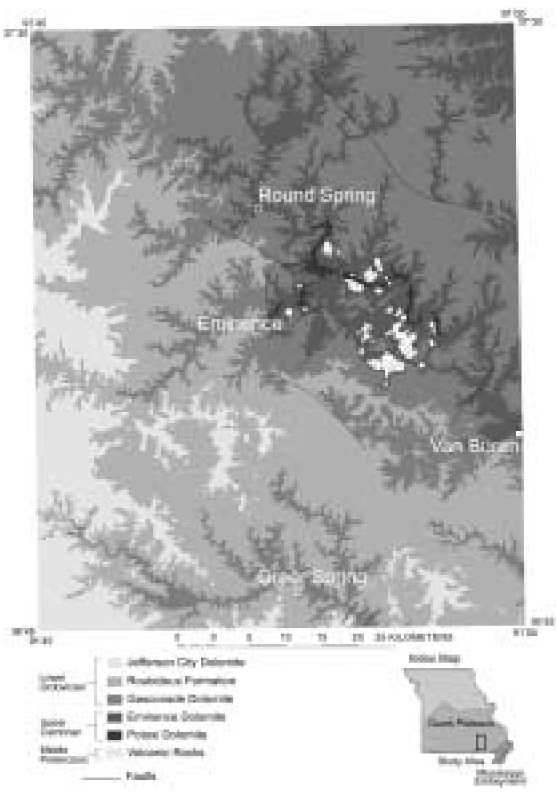


Figure 3. Generalized geologic map of south-central Missouri.

Distinctive karst features including underground drainage are abundant. Some of the largest springs in the United States are found in the area, including the two largest springs in Missouri, Big Spring (average flow 12 $m^3$/sec or 282 million gallons per day) and Greer Spring (8 $m^3$/sec or 183 million gallons per day) (fig. 1) (Vineyard and Feder, 1982). The discharge of all Ozark springs fluctuates with amount of precipitation. The prevalence of underground drainage is indicated by losing and disappearing streams as well as extensive cave and conduit systems. Dye traces show that subsurface drainage crosses surface drainage divides (Aley and Aley, 1987). For example, dye introduced in the Eleven Point River drainage basin surfaced at Big Spring along the Current River (Aley, 1975).

## GEOLOGY OF THE CAVE AND CONDUIT SYSTEM

Thirty nine caves were visited and studied in order to understand the geologic controls on the development of the Ozark cave and conduit system. Geologic mapping of 19 of these cave systems included stratigraphy, passage orientations, fracture measurements, and passage morphology. In addition, 42 maps of other caves in the study area were examined for passage orientations and morphology. Sebela and others (1999) discuss the geology of four caves in detail, including stratigraphy, structure, hydrology, and cave development. For this paper, a cave is defined as a natural underground opening large enough for a human to enter; a conduit is smaller in size.

### The Role of Joints in Cave and Conduit Development

To determine the role of joints in cave and conduit development, orientations of cave passages were compared to regional and local joint trends. This was done in two ways: first, by comparing passage orientations with joints measured in several caves, and second, by comparing the passage orientations of all of the caves in the study area to the regional joint trends. In this area of the Ozarks, joints in the Cambrian and Ordovician rocks are vertical to subvertical and have a bimodal distribution (fig. 4a).

A detailed study of four caves showed that although there were some passage orientations parallel to joints measured in the caves (fig. 5c and 5d), the passages are branching and meandering and have scattered orientations (fig. 5) (Sebela and others, 1999). Cave passage orientations of 58 caves (14,260 m of passage) in the study area were compared to joint trends measured during bedrock geologic mapping on the surface (fig. 4). Joint trends in Cambrian and Ordovician dolomite are

In Eve L. Kuniansky, editor, 2001, U.S. Geological Survey Karst Interest Group Proceedings, Water-Resources Investigations Report 01-4011, p. 18-24

bimodal with the trends of 340°-0° and 70°-90°. Compass-rose diagrams of cave passage orientations overall show much variation (fig. 4b). It is apparent that there is no correlation between joint orientations and trends of cave passages.

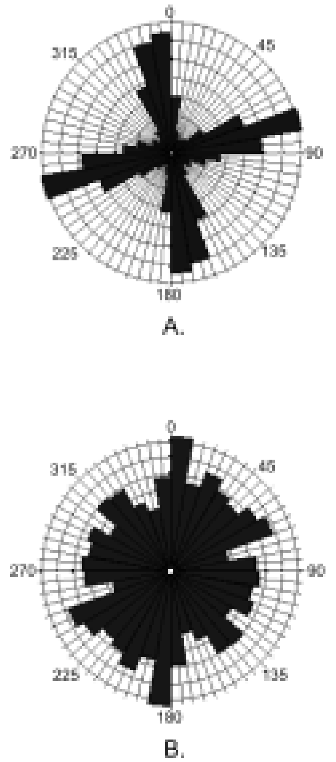


Figure 4. Compass-rose diagrams showing (A) orientation of joints measured in the study area, n=5,285, each circle represents 2 percent of total, maximum 14 percent, and (B) orientation of cave passages, n=14,260 m, each circle represents 1 percent of total, maximum 8 percent.

## The Role of Bedding in Cave and Conduit Development

Evidence for bedding control on cave and conduit development in the Ozarks includes a relationship between cave horizons and stratigraphic position, the branching morphologies of the caves, caves parallel or subparallel to bedding horizons, and the lack of correlation between cave passages and joint trends. As discussed previously, joint trends do not correlate with cave passage trends suggesting other geologic controls on the development of the cave and conduit system. Palmer (1991) showed that branching cave systems are indicative of bedding control where curvilinear passages of branchwork morphology are controlled by bedding parting porosity.

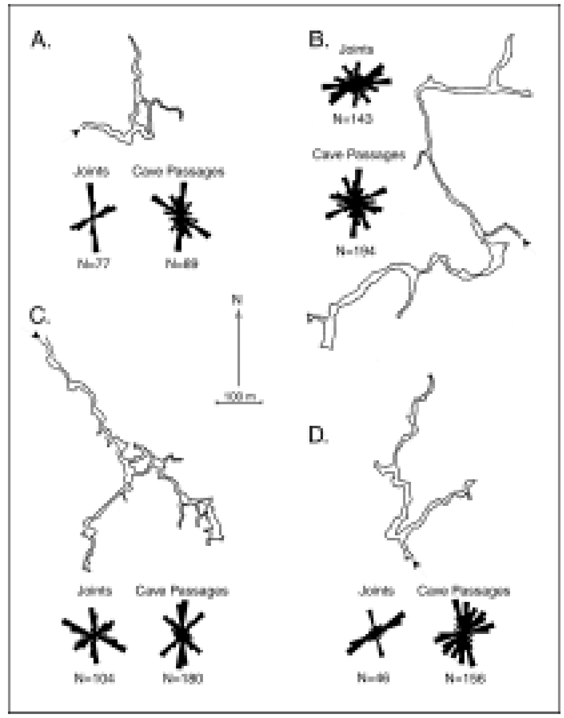


Figure 5. Cave plans and compass-rose diagrams of joints in caves and cave passages for (A) Branson Cave, (B) Round Spring Cavern, (C) Wind Cave, and (D) New Liberty Cave. Cave plans reduced from maps produced by the Cave Research Foundation. Arrows indicate cave entrances. Cave passages were weighted with respect to length.

Cave horizons almost exclusively occur immediately below sandstone horizons in the stratigraphic section (fig. 2). Sandstone horizons occur within the Eminence Dolomite in the eastern part of the study area, at the base of the Gasconade Dolomite as the Gunter Sandstone Member, and at the base and within the Roubidoux Formation. Ninety-four percent of all caves examined in this study are directly below sandstone. The sandstone within the Eminence Dolomite pinches out in the western part of the study area, and where it does not exist, the correlative dolomite horizon does not contain caves. Although it is difficult to determine the role of the sandstone in the development of caves, its role as a confining unit is evident because in this region it has a relatively low primary porosity. Several possibilities for dissolution below sandstone include a change in the chemistry of water as it migrates through fractures within the sandstone, concentrated recharge beneath the sandstone cap (Palmer, 2000), or aggressive water due to increased pressure and mixing below the confining sandstone horizons (artesian conditions). Arakaki and Mucci (1995) showed that increasing the partial pressure of $CO_2$ decreases the pH of the solution. In confining conditions beneath the sandstone, this drop in pH can make the ground water more aggressive and therefore enhance dissolution within these horizons. Also, Shi and Zhang (1992) modeled dissolution in confining conditions in gently dipping strata and found that caves developed near the top of the confined aquifer. It is possible that the sandstone was a confining unit in the past and the increased pressure on the system contributed to solution in the dolomite beneath the sandstone (fig. 6). Since ground water in confined conditions would have to travel long distances and are not close to recharge areas, epigenic water sources dependent on soil $CO_2$ are limited (Klimchouk, 2000).

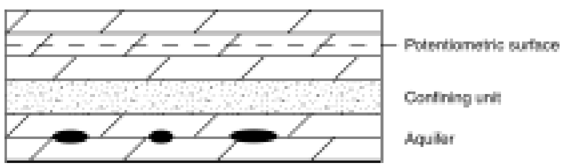


Figure 6. Diagrammatic cross section showing sandstone as a confining layer and the potentiometric surface above the confining unit. Black ovals represent caves. Water in the aquifer is under pressure and enhances dissolution because of decrease in pH.

Deep-seated sources of aggressive water in the Ozarks include organics in the stratigraphic section and sulfates from mineralization in this area. Palmer (2000) noted that interaction between sulfates and carbonates greatly increases the solubility of dolomite. The upper part of the aquifer under pressure would be a likely place for enhanced dissolution from the mixing of meteoric water with the deep-seated waters. It is generally accepted that solutional aggressiveness can be enhanced by mixing water of differing chemistry (Klimchouk, 2000). The branchwork type of cave morphology is also consistent with cave development under confining conditions (Palmer, 1991; 2000). Ford and others (2000) showed that increasing pressure gradients allow fluid to travel greater distances before attaining saturation and create a larger incidence of branching channels.

Most cave horizons occur in stromatolitic dolomite. Vugs in stromatolitic dolomite have a higher primary porosity than other dolomite horizons. Since stromatolitic dolomite occurs throughout the stratigraphic section and not just beneath sandstone horizons, it is presumed that the sandstone plays a more important role in the development of the cave and conduit system. The vugs within the dolomite provide the surface area for dissolution to occur.

Although most cave passages have been modified in the vadose zone by cave streams (canyon cutting) and ceiling breakdown, some still show their phreatic origin as tubes. These cave segments show how bedding plane partings are the control on development as joints are not seen in many of the passages (fig. 7). These bedding plane partings may be open from gentle folding allowing for dissolution to occur along them.


Figure 7. Photograph of passage in Camp Yarn Cave. Note phreatic tube-like morphology and lack of joints in ceiling. Phreatic tube developed along a bedding plane parting.

## CONCEPTUAL MODEL

Figure 8 is a diagram of the karst hydrologic system for the Ozarks of south-central Missouri. Precipitation that does not evaporate or runoff into surface streams enters the karst system through diffuse infiltration and sinkholes. Some component of ground water also enters from losing streams. In the Ozark aquifer, the cave and conduit system along with fractures in the vadose zone transports water both laterally to springs and seepage areas and vertically to the water table. Active dissolution occurs near the top of the phreatic zone where carbonic acid in meteoric waters reacts with the carbonate rock. Ground water from the phreatic conduits is then discharged to the Current River and Jacks Fork through large springs such as Big Spring, Greer Spring, Blue Spring, Alley Spring, and Round Spring (fig. 1). Dissolution may also be occurring near the top of the confined St. Francois aquifer by methods described above. Active dissolution in the St. Francois aquifer would require a source of aggressive water produced by mixing near the top of the aquifer. Most water in this aquifer is stored in the fracture and pore system.

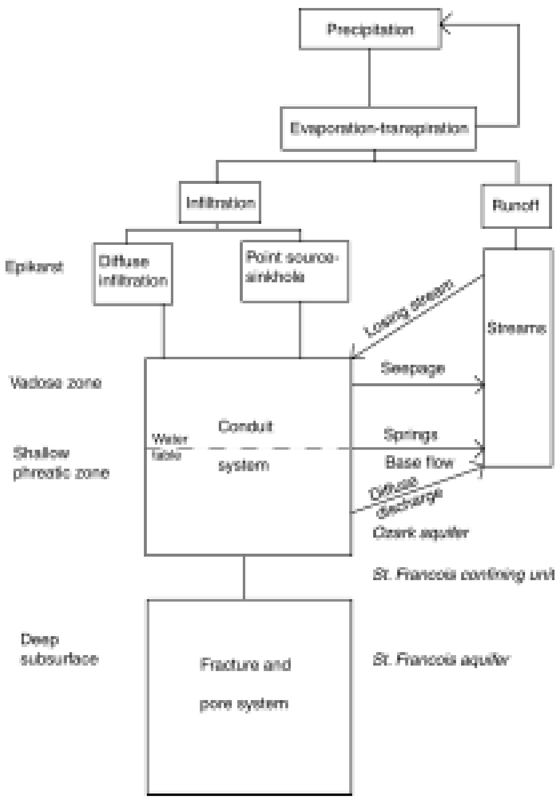


Figure 8. Diagram showing the karst hydrologic system of the Ozarks of south-central Missouri.

## CONCLUSIONS

The dominant geologic controls on cave and conduit development in the Ozarks of south-central Missouri are bedding and stratigraphy. Primary porosity as vugs in the stromatolitic dolomite and secondary porosity as joints are the pre-solutional openings where dissolution is initiated. After a certain conduit width is achieved, bedding planes then control the lateral movement of ground water. These bedding planes are more readily used for ground-water movement than joints because they are more continuous. Preferential horizons for cave and conduit development are beneath sandstones. These sandstones are hypothesized to act as confining units where hydraulic pressure beneath them and mixing of water with differing chemistry increases dissolution in the dolomite beds underlying them.

By understanding these geologic controls on karst development along with hydrogeologic properties of aquifers, land-use managers can have a better understanding of the water resources in this area of potential lead and zinc mineralization. The dip of bedding planes as part of subtle structures, as delimited on geologic maps, may help to understand the direction of ground-water flow. This would then help to determine if mining activities in the National Forest in certain areas would environmentally effect the ground-water and spring system in the Ozark National Scenic Riverways.

## ACKNOWLEDGEMENTS


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# The Relation Between Structure and Saltwater Intrusion in the Floridan Aquifer System, Northeastern Florida

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## Abstract

Saltwater intrusion is a potential threat to the quality of ground water in northeastern Florida. Elevated chloride concentrations have been observed in more than 70 wells tapping the Upper Floridan and the upper zone of the Lower Floridan aquifers. In Duval and northern St. Johns County, increased chloride concentrations in water from some wells along the coast and up to 14 miles inland indicate that saline water is gradually intruding into the freshwater zones of the Floridan aquifer system. Several mechanisms may explain this intrusion of saline water and the consequent increase in concentrations of chloride in northeastern Florida. The most plausible explanation is the upward movement of saline water along joints, fractures, collapse features, faults, or other structural anomalies. Land-based seismic reflection and marine seismic reflection profiles along the St. Johns River and off the coast of northeastern Florida show the presence of widely scattered solution collapse features in the Floridan aquifer system and overlying sediments. These features can create conduits of relatively high vertical conductivity, providing a hydraulic connection between freshwater zones and deeper, more saline zones. Lower heads caused by pumping from the shallower freshwater zones of the aquifer can result in an increased potential for upward movement of saline water through nearly vertical zones of preferential permeability. Saline water then can move laterally through the porous aquifer matrix or along horizontal fractures or solution zones within the aquifer toward well fields or other areas of lower hydraulic head.

## INTRODUCTION

The Floridan aquifer system is the major source of water supply in northeastern Florida (fig. 1). Ground-water withdrawals in the area increased from about 183 to 235 million gallons per day from 1965 to 1995 (Marella, 1995; 1999). Approximately 90 percent of the total withdrawal is from the Floridan aquifer system, resulting in long-term declines in the potentiometric surface of the Upper Floridan aquifer of about one-third to one-half foot per year. Associated with this decline in water levels has been an increased potential for saltwater intrusion into the freshwater zones of the aquifer. Gradual, but continual, increases in the chloride concentrations in water from the aquifer system have been observed in several inland and coastal areas (fig. 2). The potential for saltwater intrusion is expected to increase as population growth and potentiometric surface declines continue in northeastern Florida.

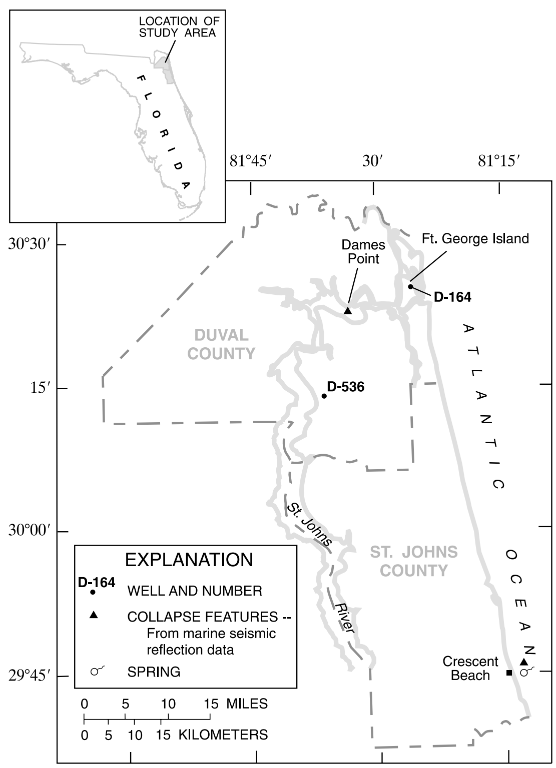


**Figure 1. Location of the study area.**

## HYDROGEOLOGIC FRAMEWORK

Major stratigraphic and corresponding hydrogeologic units of northeastern Florida (fig. 3) include a thick sequence of marine sedimentary rocks that overlie a basement complex of metamorphic strata. The sedimentary sequences are primarily carbonates that contain interbedded evaporites in the deeper units and siliclastic material in the shallower units. Rocks of

In Eve L. Kuniansky, editor,2001, U.S. Geological Survey Karst Interest Group Proceedings, Water-Resources Investigations Report 01-4011, p. 25-29

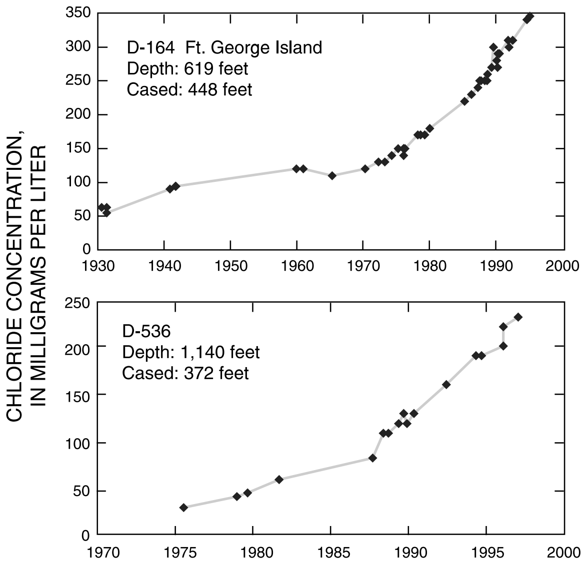


**Figure 2. Chloride concentrations in water from selected wells tapping the Floridan aquifer system (locations of wells shown in figure 1).**

the Cedar Keys Formation of late Paleocene age underlie all of northeastern Florida. They are overlain, in ascending order, by the Oldsmar Formation of early Eocene age, the Avon Park Formation of middle Eocene age, the Ocala Limestone of late Eocene age, the Hawthorn Group of Miocene age, and the undifferentiated deposits of late Miocene to Holocene age.

The principal water-bearing units are the surficial and Floridan aquifer systems. The two aquifer systems are separated by the intermediate confining unit, which is composed of clays, silts, and sands. The intermediate confining unit contains beds of lower permeability that confine the water in the Floridan aquifer system. The Floridan aquifer system is divided into the Upper and Lower Floridan aquifers, which are separated by a zone of lower permeability (fig. 3). The surface of the Upper Floridan is a paleokarst plain that exhibits erosional and collapse features that developed before the deposition of the overlying Hawthorn Group. Two major water-bearing zones exist within the Lower Floridan aquifer: the upper zone and the Fernandina permeable zone (fig. 3). These zones are separated by a less-permeable semiconfining unit that restricts the vertical movement of water (Brown, 1984). Water in the Fernandina permeable zone varies from fresh to highly saline in northeastern Florida.

| Series | Stratigraphic unit | General lithology | Hydrogeologic unit | | | Hydrogeologic properties |
|---|---|---|---|---|---|---|
| Holocene to Upper Miocene | Undifferentiated surficial deposits | Discontinuous sand, clay, shell beds, and limestone | Surficial aquifer system | | | Sand, shell, limestone, and coquina deposits provide local water supplies. |
| Miocene | Hawthorn Group | Interbedded phosphatic sand, clay, limestone, and dolomite | Intermediate confining unit | | | Sand, shell, and carbonate deposits provide limited local water supplies. Low permeability clays serve as the principle confining beds for the Floridan aquifer system below. |
| Eocene | Ocala Limestone | Massive fossiliferous chalky to granular marine limestone | Floridan aquifer system | Upper Floridan aquifer | | Principal source of ground water. High permeability overall. Water from some wells shows increasing salinity. |
| | Avon Park Formation | Alternating beds of massive granular and chalky limestone, and dense dolomite | | Middle semiconfining unit | | Low permeability limestone and dolomite. |
| | | | | Lower Floridan aquifer | Upper zone | Principal source of ground water. Water from some wells shows increasing salinity. |
| | Oldsmar Formation | | | | Semiconfining unit | Low permeability limestone and dolomite. |
| | | | | | Fernandina permeable zone | High permeability; salinity increases with depth. |
| Paleocene | Cedar Keys Formation | Uppermost appearance of evaporites; dense limestones | Sub-Floridan confining unit | | | Low permability; contains highly saline water. |

**Figure 3. General geology and hydrogeology of northeastern Florida (modified from Spechler, 1994).**

## MECHANISMS OF INTRUSION

There are several possible mechanisms, some more plausible than others, that can explain the processes and pathways of saltwater movement within the Floridan aquifer system (Spechler, 1994). They are: (1) the movement of unflushed pockets of relict seawater within the aquifer system; (2) landward movement of the freshwater-saltwater interface; (3) regional upconing of saltwater below pumped wells; and (4) the upward leakage of saltwater from deeper, saline water-bearing zones of the aquifer system through semiconfining units that are thin or are breached by joints, fractures, collapse features, or other structural anomalies.

During the Pleistocene epoch, sea level stood at a much higher level than it does today and the Floridan aquifer system was invaded with seawater. Some of this water may not have been completely flushed from the aquifer. Generally, pockets of unflushed relict seawater are found in strata of relatively low permeability. Geophysical logs of many wells in Duval County, however, indicate that the more permeable zones are the source of saline water to the well, implying that unflushed relict seawater is a minor source of saline water in the study area.

Lateral encroachment of recent seawater is an unlikely explanation for the increase in chloride concentrations in the Floridan aquifer system in Duval County. If seawater were moving laterally through the Upper Floridan aquifer from outcrops in the Atlantic Ocean, the saltwater would first be detected in wells nearest the coast. However, many of the public supply and domestic wells along the coast have chloride concentrations less than 30 milligrams per liter (mg/L), whereas some wells as much as 14 miles inland of the coast have chloride concentrations exceeding 250 mg/L. In addition, data from abandoned oil wells and exploratory wells drilled off the coast indicate that the position of the freshwater-saltwater interface at the top of the Floridan aquifer is miles offshore (Johnston, 1983; Johnston and others, 1982; Wait and Leve, 1967).

Data from geophysical logging indicate that regional upconing of saline water apparently is not occurring in Duval County. If upconing were occurring, elevated chloride concentrations in water would be areally distributed under cones of depression caused by pumping. Also, the transition zone would be moving upward and chloride concentrations would be expected to increase with depth. However, fluid resistivity logs and chloride samples collected from several wells with elevated chloride concentrations indicate alternating zones of fresh and saline water, and that less-mineralized water generally underlies the shallower higher chloride zone (Phelps and Spechler, 1997).

Because of the areal and vertical variability of chloride concentrations, the most plausible mechanism for the movement of saline water into the freshwater zones of the Floridan aquifer system is the upward movement of saline water along solution-enlarged joints or fractures, and subsequently formed collapse features, combined with horizontal movement in fractures or solutionally enhanced flow zones (fig. 4). Marine seismic reflection profiles show that the Continental Shelf off the northeastern coast of Florida is underlain by solution-deformed limestone of late Cretaceous to Eocene age (Meisburger and Field, 1976; Popenoe and others, 1984; Kindinger and others, 2000). Dissolution and collapse features are widely scattered throughout the area and are expressed as: (1) sinkholes that presently breach the sea floor (Spechler and Wilson, 1997); (2) sinkholes that breached the sea floor in the past and are now filled with sand; and (3) dissolution-collapse features (fig. 5) that originated deep within the geologic section, deforming the overlying units (Popenoe and others, 1984). The deep dissolution-collapse features are believed to originate in the Upper Cretaceous and Paleocene rocks (Popenoe and others, 1984). Additional seismic- reflection investigations along the St. Johns River and in lakes in northeastern Florida by Snyder and others (1989), Spechler (1994), and Kindinger and others (2000) also revealed a number of buried collapse features (fig. 6) and other karstic features similar to those observed by Meisburger and Field (1976), and Popenoe and others (1984).

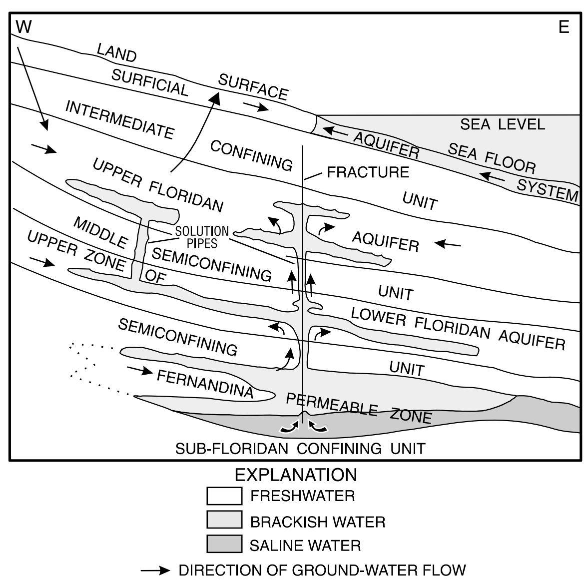


**Figure 4. Simplified section of the Floridan aquifer system (modified from Spechler, 1994).**

In Eve L. Kuniansky, editor,2001, U.S. Geological Survey Karst Interest Group Proceedings, Water-Resources Investigations Report 01-4011, p. 25-29

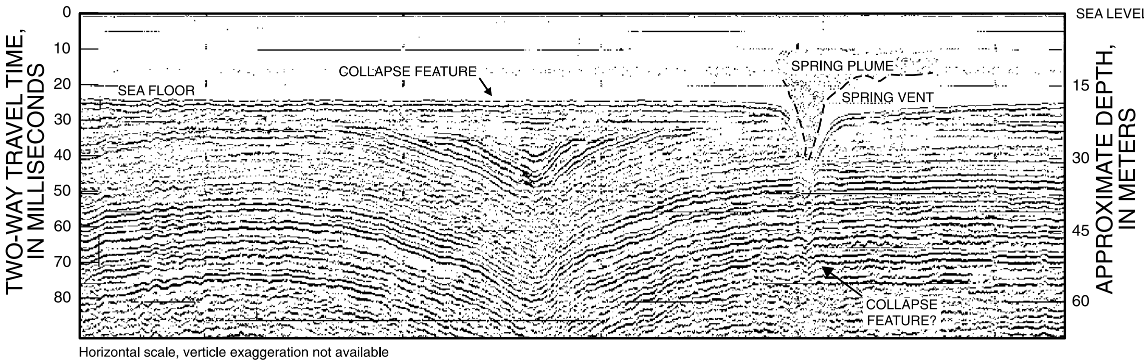


**Figure 5. Seismic record showing collapse features, offshore Crescent Beach, Florida (modified from Kindinger and others, 2000).**

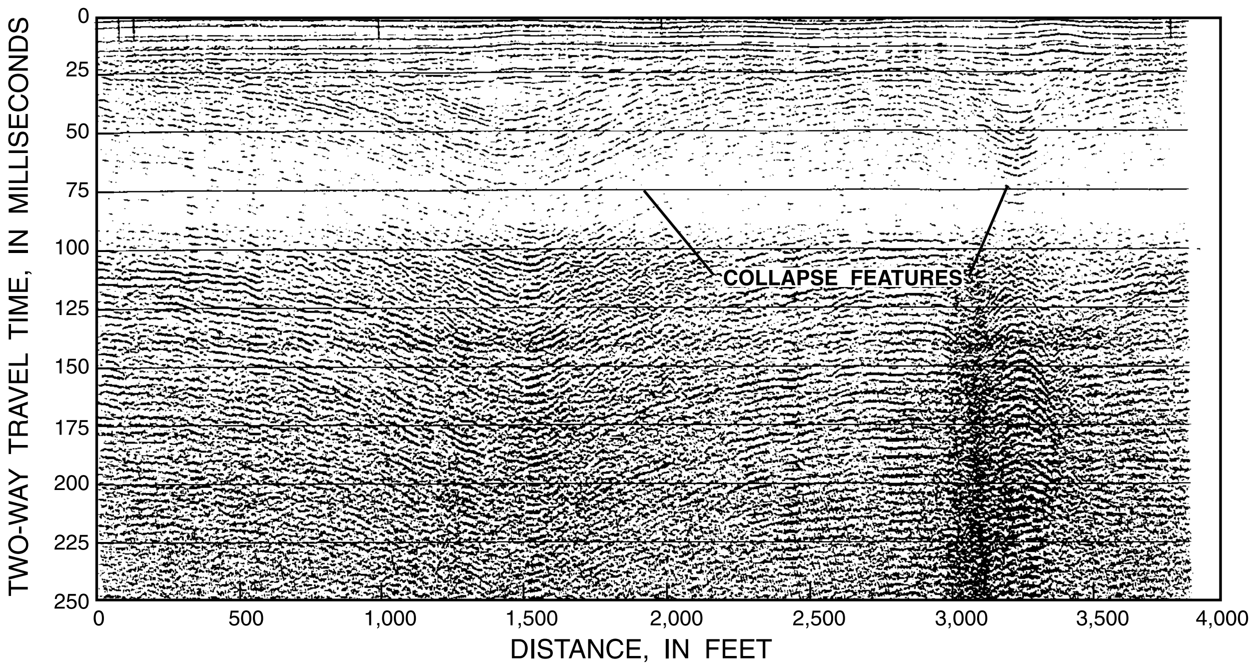


**Figure 6. Seismic record showing collapse features along an approximate 4,000-foot section of the St. Johns River near Dames Point in Duval County (from Spechler, 1994).**

Recent land-based seismic reflection surveys at Ft. George Island, Florida (fig. 1) show a large solution feature in the northeastern part of the island (Odum and others, 1999). Reflectors within the upper Hawthorn Group show evidence of downwarping and displacement (approximately 65 feet of vertical subsidence) as the interbedded carbonates, clay, and sand strata deform plastically downward over a deeper solution pipe. The seismic profiles show that a karst solution feature has likely breached the middle semiconfining unit within the Floridan aquifer system and possibly the semiconfining unit that separates the upper zone of the Lower Floridan aquifer from the Fernandina permeable zone. This feature may have created zones of relatively high vertical hydraulic conductivity through rocks of otherwise low vertical hydraulic conductivity, thereby providing a hydraulic connection between freshwater zones and deeper saline zones (fig. 4). Chloride concentrations in water from nearby wells (Spechler, 1994) are highest near the interpreted limits of the solution pipe (fig. 2), indicating that saline ground-water movement is controlled by these features.

## CONCLUSION

The areal and vertical distribution of chloride concentrations in Duval County indicates that structural anomalies are the most likely cause for increased chloride concentrations in the Floridan aquifer system. These features can create zones of relatively high vertical hydraulic conductivity, thereby providing a hydraulic connection between freshwater zones and deeper, more saline zones. Lower heads caused by pumping from the shallower freshwater zones of the aquifer can result in an increased potential for upward movement of saline water through nearly vertical zones of preferential permeability. As saline water enters the freshwater zones, it can mix and move through the porous matrix of the aquifer or along horizontal fractures or solution zones.

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# Hydrology, Hazards, and Geomorphic Development of Gypsum Karst in the Northern Black Hills, South Dakota and Wyoming

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## Abstract

Dissolution of gypsum and anhydrite in four stratigraphic units in the Black Hills, South Dakota and Wyoming, has resulted in development of sinkholes and has affected formational hydrologic characteristics. Subsidence has caused damage to houses and water and sewage retention sites. Substratal anhydrite dissolution in the Minnelusa Formation (Pennsylvanian and Permian) has produced breccia pipes and pinnacles, a regional collapse breccia, sinkholes, and extensive disruption of bedding. Anhydrite removal in the Minnelusa probably dates back to the early Tertiary when the Black Hills was uplifted and continues today. Evidence of recent collapse includes fresh scarps surrounding shallow depressions, sinkholes more than 60 feet deep, and sediment disruption and contamination in water wells and springs. Proof of sinkhole development to 26,000 years ago includes the Vore Buffalo Jump, near Sundance, WY, and the Mammoth Site in Hot Springs, SD. Several sinkholes in the Spearfish Formation west of Spearfish, SD, which support fish hatcheries and are used for local agricultural water supply, probably originated 500 feet below in the Minnelusa Formation. As the anhydrite dissolution front in the subsurface Minnelusa moves down dip and radially away from the center of the Black Hills uplift, these resurgent springs will dry up and new ones will form as the geomorphology of the Black Hills evolves. Abandoned sinkholes and breccia pipes, preserved in cross section on canyon walls, attest to the former position of the dissolution front. The Spearfish Formation, mostly comprising red shale and siltstone, is generally considered to be a confining layer. However, secondary fracture porosity has developed in the lower Spearfish due to considerable expansion during the hydration of anhydrite to gypsum. Thus, the lower Spearfish yields water to wells and springs making it a respectable aquifer. Processes involved in the formation of gypsum karst should be considered in land use planning in this increasingly developed part of the northern Black Hills.

## INTRODUCTION

The Black Hills of western South Dakota (fig. 1) is experiencing increased urban development requiring an assessment of ground-water contamination potential. Detailed bedrock and surficial geologic mapping, in cooperation with the Lawrence County Planning Commission and the City of Spearfish, SD, will be useful for assessing aquifer-contamination potential by describing major lithologic characteristics, delineating surface recharge areas, and characterizing subsurface structural configuration. The maps will also be useful for depicting areas of potential landsliding, soil erodability, and subsidence due to solution of underground gypsum and anhydrite.

The Black Hills comprise an asymmetric uplift, about 130 miles long and 60 miles wide. Erosion has exposed a core of Precambrian metamorphic rocks which are rimmed by shallow marine to nearshore-terrestrial sediments of Paleozoic and Mesozoic age which dip away from the center of the domal uplift (fig. 1). The homoclinal dips are locally interrupted by monoclines, structural terraces, low-amplitude folds, faults, and Tertiary igneous intrusions. One fold, the LaFlamme anticline, is a prominent structure west of Spearfish (see fig. 9). It plunges to the northwest, it is at least 10 miles long, about 8 miles wide, has a structural relief in places of more than 600 ft (260 m), and the dips on its flanks are as much as $20^{o}$.

More than 300 ft (91 m) of gypsum and anhydrite were deposited at various times in evaporite basins. Rocks of the Madison Limestone (Pahasapa of other reports), Minnelusa Formation and older sediments form the “limestone plateau” that rims the central Precambrian metamorphic core. Erosion of weak red siltstones and shales of the Spearfish Formation has formed the "Red Valley" (fig. 1), the main area of present and proposed future development. Resistant sandstone forms the hogback that encircles the Black Hills and defines its outer physiographic perimeter.

In Eve L. Kuniansky, editor, 2001, U.S. Geological Survey Karst Interest Group Proceedings, Water-Resources Investigations Report 01-4011, p. 30-37

## STRATIGRAPHY OF CALCIUM SULPHATE-BEARING ROCKS

Whereas karstic features in limestone and dolomite, such as caves, sinkholes, and underground drainage, are abundant in the Black Hills, similar solution features are also abundant in gypsum ($CaSO_4 \cdot 2H_20$) and its anhydrous counterpart anhydrite ($CaSO_4$). Calcium sulphate rocks are much more soluble than carbonate rocks, especially where they are associated with dolomite undergoing dedolomitization, a process which results in groundwater that is continuously undersaturated with respect to gypsum (Raines and Dewers, 1997).

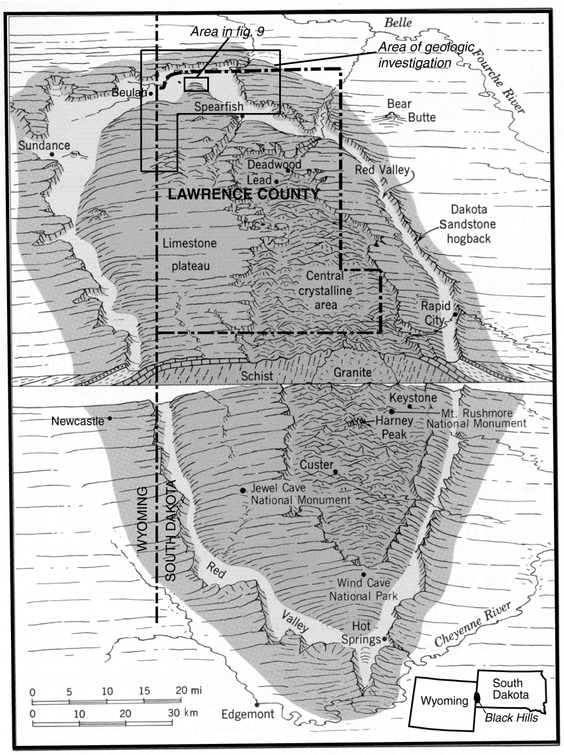


Fig. 1. Generalized diagram showing the geology and geomorphology of the Black Hills. Most of the urban development and karst features in Lawrence County are in the Red Valley, underlain by Triassic red beds (where gypsum karst is becoming a growing concern) and in the limestone plateau, underlain by a variety of Pennsylvanian and Permian rocks. Modified from Strahler and Strahler, 1987, with permission.

Gypsum and anhydrite are conspicuous evaporite deposits in four sedimentary rock units in the Black Hills (fig 2). They comprise about 30 percent of the Minnelusa Formation (generally present only in the subsurface), less than 5 percent of the Opeche and Spearfish Formations, and about half of the Gypsum Spring Formation.

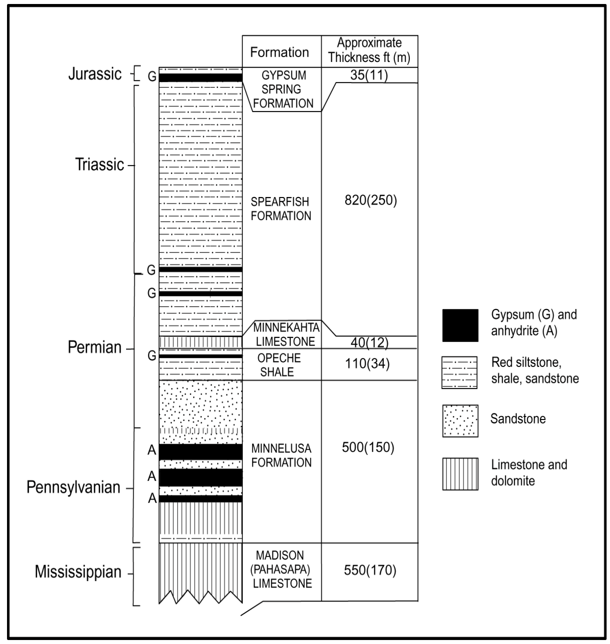


Fig. 2: Stratigraphic column showing distribution of gypsum and anhydrite in the northern Black Hills.

The Minnelusa Formation in the northern Black Hills consists of approximately 500 feet (150 m) of dolomite, sandstone, and shale with anhydrite prevalent in the middle. The anhydrite is mostly absent in surface outcrops, having been removed by solution in the subsurface. The solution of anhydrite and consequent formation of voids in the Minnelusa at depth resulted in foundering and fragmentation of overlying rocks, producing extensive disruption of bedding, a regional collapse breccia, many sinkholes, and breccia pipes and pinnacles (e.g., Epstein, 1958a,b; Brobst and Epstein, 1963; Bowles and Braddock, 1963)(Figs. 3,4,5). Some sinkholes and resistant calcite-cemented pinnacles extend upward more than 1,000 ft (300 m) into overlying strata (Bowles and Braddock, 1963). The collapse breccia consists of angular clasts of limestone, dolomite, and sandstone in a sandy matrix that is generally cemented with calcium carbonate. It has a vuggy secondary porosity, which, along with the porous sandstone, makes the upper half of the Minnelusa an important aquifer in the Black Hills.

Gypsum is not abundant in the 110 ft (34 m) of poorly exposed red shale, siltstone, and fine-grained

In Eve L. Kuniansky, editor, 2001, U.S. Geological Survey Karst Interest Group Proceedings, Water-Resources Investigations Report 01-4011, p. 30-37

sandstone of the Opeche Formation, a confining unit between the Minnelusa Formation and Minnekahta Limestone.

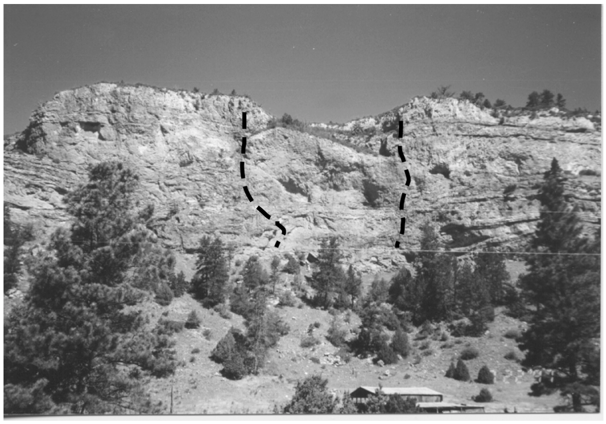

Figure 3. Sinkhole (outlined) in the Minnelusa Formation exposed on 400-foot-high cliff face in Redbird Canyon, about 10 miles east of Newcastle, WY, in Custer County, SD. The collapse resulted from removal of anhydrite by ground water prior to fluvial erosion, which exposed the sinkhole on the canyon wall.

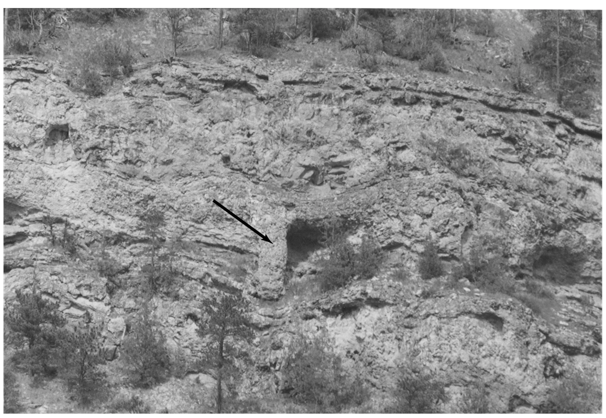

Figure. 4. Disrupted bedding and breccia pipe (arrow) in the Minnelusa Formation. Collapse and disruption were due to dissolution of anhydrite at depth. Cliff in Cold Brook Canyon, just north of Hot Springs, SD.

The Spearfish Formation consists of about 820 ft (250 m) of fine red beds with several layers of gypsum in the lower 200 ft (60 m). Anhydrite, which probably was the original form of calcium sulphate to be deposited in the Spearfish, undergoes about a 40 percent expansion when hydrated to form gypsum.

Figure 5. Erosion of resistant breccia pipe that is cemented by calcium carbonate forms a pinnacle that is common within the brecciated upper part of the Minnelusa Formation. Red Bird Canyon, about 10 miles east of Newcastle, WY.

As a result, beds of gypsum in the Spearfish Formation are commonly highly folded. When gypsum dissolves, it becomes mobile and is injected downward as thin veinlets into fractures in the confining red beds (fig. 6). These veinlets are generally less than ½ inch (1 cm) wide, they occur along a multitude of variably oriented fractures beneath the parent gypsum bed, and they contain gypsum fibers lying perpendicular to the fracture walls. Thus, the lower 200 ft (60 m) or so of the Spearfish has developed a secondary fracture porosity. This part of the formation has supplied water to wells, many sinkholes have developed in it, and resurgent springs are numerous. Ground water flows

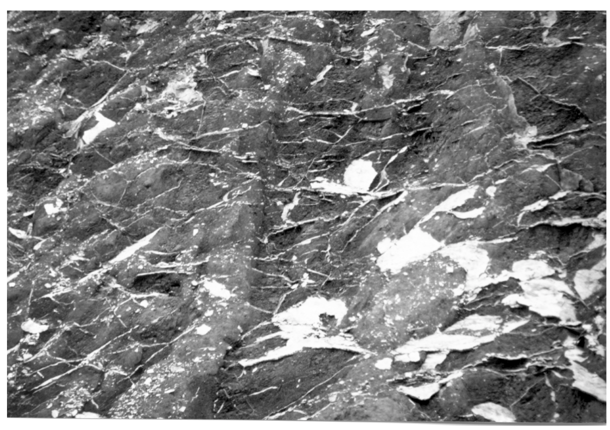

Figure 6. Thin gypsum veinlets extending down from parent gypsum bed (not shown) and filling a multitude of fractures in the lower part of the Spearfish Formation near Cascade Springs, along State Highway 71, 13 miles southwest of Hot Springs, SD.

through the fractures and solution cavities in the gypsum. Although many hydrologists consider the entire Spearfish Formation to be a confining unit, the lower 200 ft (60 m) of the Spearfish is an aquifer, in the northern Black Hills, at least. This is not surprising since high ground-water flow has been reported in gypsum in many areas of the United States (Thordarson, 1989).

The upper part of the Spearfish, about 600 ft (180 m) thick, consists of red siltstone, shale, and very fine-grained sandstone. Gypsum beds are lacking. Bedding is regular and the unit lacks the fractures seen in the lower part of the formation. This part of the Spearfish is a confining layer.

The Gypsum Spring Formation consists of about 35 ft (11 m) equally distributed between ledge-forming white gypsum at the base and shaly siltstone with thin gypsum at the top. Many sinkholes have developed in the Gypsum Spring.

## DISSOLUTION FRONT IN THE MINNELUSA FORMATION

The upper half of the Minnelusa Formation contains abundant anhydrite in the subsurface, and except for a few areas near Beulah and Sundance, Wyoming (Brady, 1931), and in Hell Canyon in the southwestern Black Hills (Braddock, 1963), no anhydrite or gypsum crops out. A log of the upper part of the Minnelusa from Hell Canyon contains 235 ft (72 m) of anhydrite and gypsum (Brobst and Epstein, 1963). Where anhydrite is present in the Minnelusa, its rocks are not brecciated. Where the rocks are brecciated in outcrop, anhydrite is absent. Clearly, the brecciation is the result of collapse following subsurface dissolution of anhydrite

The Madison and Minnelusa are the major aquifers in the Black Hills. They are recharged by rainfall on and by streams flowing across their up-dip outcrop area. In the Minnelusa, removal of anhydrite progresses downdip with continued dissolution of the anhydrite (fig. 7), collapse breccia is formed, breccia pipes extend upwards, and resurgent springs develop at the sites of sinkholes. Cox Lake (fig. 8), Mud and Mirror Lakes, and McNenny springs, are near the position of the dissolution front (fig. 9). As the Black Hills is slowly lowered by erosion, the anhydrite dissolution front in the subsurface Minnelusa moves downdip and radially away from the center of the uplift. The resurgent springs will dry up and new ones will form down dip as the geomorphology of the Black Hills evolves. Abandoned sinkholes on canyon walls (fig. 3) attest to the former position of the dissolution front.

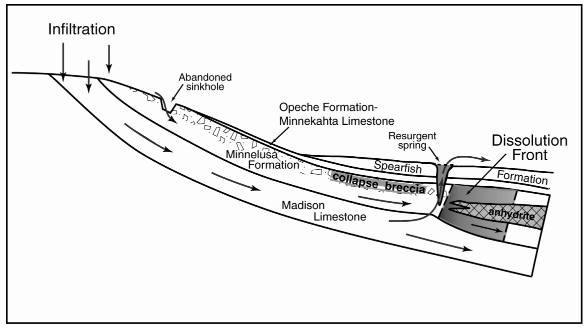


Figure 7. Dissolution of anhydrite in the Minnelusa Formation and down-dip migration of the dissolution front.

Because ground water has dissolved the anhydrite in the Minnelusa in most areas of exposure, and because anhydrite is present in the subsurface, a transition zone should be present where dissolution of anhydrite is currently taking place. A model of this zone has been presented by Brobst and Epstein (1963, p. 335) and Gott and others (1974, p. 45) and is shown here in figure 7. Consequences of this model include (1) the updip part of the Minnelusa is thinner than the downdip part because of removal of significant thickness of anhydrite, (2) the upper part of the Minnelusa should be continually collapsing, even today, and (3) the properties of the water in this transition zone may be different than elsewhere.

If this process is correct, then present resurgent springs should be eventually abandoned and new springs should develop down the regional hydraulic gradient of the Black Hills. One example might be along Crow Creek where a cloud of sediment from an upwelling spring lies 1,000 ft (300 m) north of McNenny Springs (fig. 9). This circular area, about 200 ft (60 m) across, might eventually replace McNenny Springs.

Solution of anhydrite in the Minnelusa probably began soon after the Black Hills was uplifted in the early Tertiary and continues today. Recent subsidence is evidenced by sinkholes more than 60 ft (18 m) deep opening up within the last 20 years (fig. 10), collapse in water wells and natural springs resulting in sediment disruption and contamination (Hayes, 1996), and fresh circular scarps surrounding shallow depressions.

Figure 8. Cox Lake, a resurgent (artesian) spring with a flow of nearly 5 cubic feet (0.5 cu m) per second in the Spearfish Formation in the northern Black Hills. It occupies a sinkhole that is more than 60 ft (18 m) deep (outlined by the darker water just beyond the edge of the dock). The chemical signature of the water indicates that the Minnelusa Formation and underlying Madison Limestone are the contributing aquifers (Klemp, 1995). The lake is near the anhydrite dissolution front shown in fig. 7.

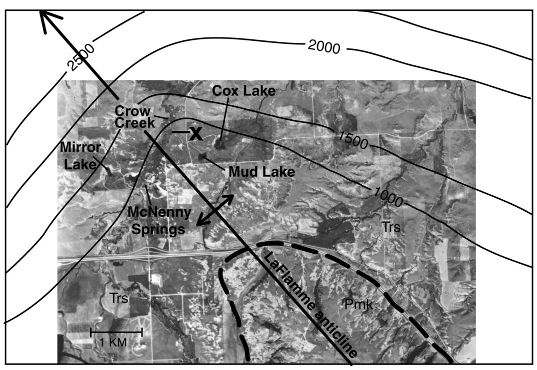


Figure 9. Air photograph showing location of resurgent springs in the Spearfish Formation adjacent to the LaFlamme anticline. X marks site of new resurgent spring along Crow Creek. Pmk, Minnekahta Limestone; Trs, Spearfish Formation. Specific conductance in the Minnelusa aquifer (contours in microseimens per second) from Klemp (1995).

The Plains Indians that inhabited the area 300 years ago trapped and slaughtered thousands of buffalo for their primary food by stampeding the animals over the steep rim of one of the large sinkholes near Beulah, WY (the Vore Buffalo Jump). The site is undergoing archeological excavation by the University of Wyoming. The Hot Springs Mammoth Site in Hot Springs, SD., is another large sinkhole in the Spearfish Formation that was the site of a breccia pipe extending down into the Minnelusa Formation. It was an active trap for large mammals at least 26,000 years ago (Laury, 1980; Agenbroad and Mead, 1994).

Figure 10. Sinkhole in the red beds of the Spearfish Formation that formed about 1950 near Beulah, WY. It developed in a larger and shallower 1000-foot wide depression.

A series of springs that apparently occupy sinkholes, as well as dry sinkholes, occur in the lower half of the Spearfish Formation, generally within 200 ft (60 m) of the base of the formation, and at or near where several beds of gypsum are exposed. Several lines of reasoning suggest that these sinkholes are not the result of solution of gypsum in the Spearfish, even though there is some collapse due to dissolution of gypsum in that formation: (1) The gypsum beds exposed in the lower Spearfish aggregate no more than about 25 ft (8 m) in thickness, whereas the sinkholes are more than 50 ft (15 m) deep in places; (2) Several of the sinkholes lie below many of the gypsum beds, and (3) the waters of some of the lakes occupying the sinkholes are derived from underlying formations (Klemp, 1995).

## HYDROLOGIC IMPLICATIONS OF SINKHOLES IN THE LOWER PART OF THE SPEARFISH FORMATION;

The Spearfish Formation comprises red shale, siltstone, and fine sandstone with scattered beds of gypsum. In the area of Spearfish, SD, and west to the Wyoming-South Dakota border, the lower part of the formation has different structural and lithologic characteristics that affect its hydrologic behavior. The lower 200 ft (60 m) or so is an aquifer. The overlying rocks are a confining layer. The lower part yields abundant water to wells and springs. It contains gypsum beds, many of which are contorted, and gypsum veinlets that are intruded into fractures at variable angles. Local ranchers examined the sinkhole shown in figure 10 and running water was heard beneath. A cavern extending horizontally beyond the limits of their flashlight beam was seen (Ted Vore, oral commun., 1999). Cox (1962, p. 11) reported a well 2.8 miles (4.5 km) ENE of Cox Lake that bottomed in a cavern in the Spearfish. A working hypothesis is that the lower Spearfish contains abundant, interconnected caves and solution fractures along which water flows rapidly and supplies the wells and Cox Lake and surrounding resurgent springs. This zone is a fractured rock aquifer in which ground water travels by conduit flow. Where the potentiometric surface lies above the ground surface in this zone, the sinkholes are sites of resurgent springs. Where the potentiometric surface lies below the ground surface within this zone, dry sinkholes have developed (fig. 11). It is possible that the aquifer rocks are more intensely fractured in the LaFlamme anticlinal area, allowing for greater secondary porosity and permeability, and accounting for the location of these springs (fig. 8). An analysis of intensity of fracturing is under investigation.

Many of the sinkholes in the Spearfish Formation are too large to be accounted for by solution of the relatively thin gypsum beds within that formation. They were more likely produced by the removal of much thicker gypsum in the Minnelusa Formation, approximately 500 ft (150 m) below. Many sinkholes that extend down through the Spearfish into the Minnelusa are sites of resurgent springs (fig. 7), resulting from fairly recent dissolution of anhydrite in the Minnelusa. These springs are important for recharging to surface streams in the Black Hills. Rainfall recharges aquifers higher up near the core of the Black Hills. The ground water discharges along the flanks of the hills, mainly in or near the upper Spearfish aquiclude, where the potentiometric surface is higher than the ground surface.

## ENLARGEMENT OF MIRROR LAKE VIA "HEADWARD COLLAPSE"

Mirror Lake (fig. 12), located in the South Dakota State Wildlife Management Area, has a dogleg shape. The eastward-trending section is artificial. The northwest-trending, 900-foot (275 m)-long alcove is cut into a 50-foot-high ridge of the

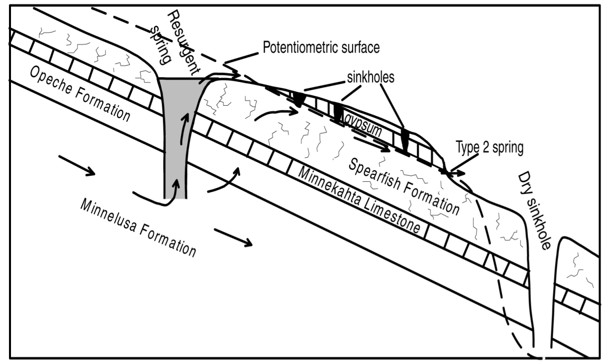


Figure 11. Generalized diagram showing hypothetical flow path of ground water in resurgent spring and in spring emerging from beneath gypsum in Spearfish Formation (type 2 spring of Rahn and Greis, 1973) and sinkholes developed by solution of anhydrite in the Minnelusa Formation. Dry sinkhole lies above potentiometric surface. Gypsum veinlets in fractured rocks of the lower part of the Spearfish shown by jagged lines.

Spearfish Formation (fig. 12). The lake, similar to other lakes in the area, occupies a depression formed by dissolution of calcium sulphate at depth, probably in the Minnelusa, although gypsum underlies the lake as shown by outcrops nearby. A deposit of calcareous tufa, as much as four ft (1.2 m) thick, consisting of light-brown, porous limestone with abundant plant impressions (“moss rock” of local ranchers) is found 1,000 ft (300 m) southeast of the lake. The deposit dips gently to the east, away from Mirror Lake and presumably was deposited earlier by spring water that emerged from the lake. Numerous sinkholes, several feet deep, are found at the north end of the alcove. These presently are active and indicate that the lake is expanding to the northwest by continued collapse due to solution of gypsum at depth. The fine sediment derived from the Spearfish is presumably removed by the emerging spring water. Presumably the lake was once higher at the time the tufa was deposited (fig. 12). Continued downcutting and northwest migration of the headwall has produced the present landform, a pocket valley that has been termed a “steephead” (Jennings, 1971). Dating the sediments in the bottom of the lake may yield the rate of headward erosion of the steephead.

## ENVIRONMENTAL CONCERNS

Karstic collapse due to dissolution of gypsum and anhydrite is an active process in the northern Black Hills. Dissolution of gypsum in the Spearfish and Gypsum Spring has resulted in collapse and formation

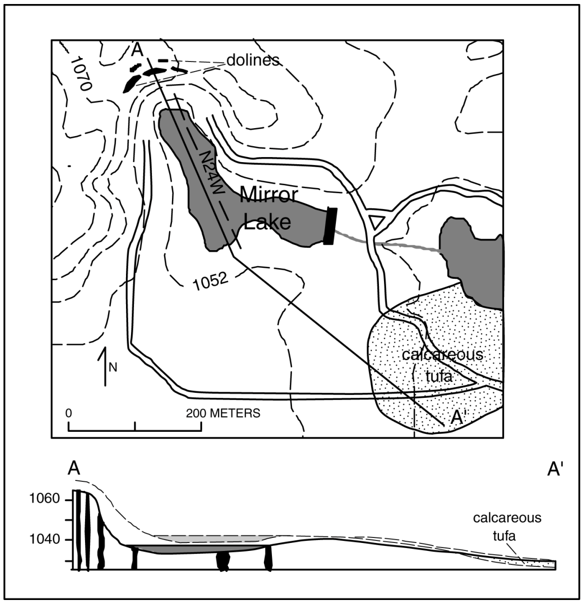


Fig. 12. Map and cross section showing sinkholes at the northwest end of Mirror Lake and the inferred earlier topography on which calcareous tufa was deposited. Contour interval 6 meters.

of many sinkholes in several areas that are presently undergoing development between Rapid City and Spearfish, SD (Rahn and Davis, 1996; Davis and Rahn, 1997). In 1972, the City of Spearfish constructed a sewage lagoon on the Gypsum Spring Formation. The lagoon leaked into sinkholes and the lagoon was abandoned in favor of an expensive water-treatment plant. Recently, the city entertained plans to convert the lagoon site into a recreation area with construction of buildings and light towers. The Public Works Administrator requested the USGS for a judgment on the potential for subsidence at the site. A geologic map was prepared, similar to one prepared by Davis (1979, fig. 3) showing that at least ten sinkholes, one of which is about 1,000 ft (300 m) long, had developed in the gypsum. This information was subsequently used by the city planners in their decision to abandon the project.

The lower part of the Spearfish Formation is also undergoing active collapse and is considered to be an aquifer. This zone is more susceptible to rapid infiltration of contaminants than the upper part of the Spearfish, a fact that should be considered in future land-use planning. Collapse due to dissolution of soluble rocks can be exacerbated by removal of ground water by pumping. If areas in the Red Valley of the northern Black Hills are extensively developed and water supplies derived from pumping, then a possible concern might be the increase in frequency of such collapse in the Spearfish Formation.

The Minnelusa Formation is a heterogeneous unit. The upper part, which is highly brecciated and contains numerous breccia pipes, has a greater fracture porosity than the lower part. Care should be taken in not considering the entire Minnelusa as a unified aquifer, especially within and updip of the dissolution front.

## CONCLUSION

Dissolution of gypsum and anhydrite in the Minnelusa and Spearfish Formations in the northern Black Hills has led to subsidence and collapse resulting in the development of disrupted bedding, breccia pipes and pinnacles in the Minnelusa, and sinkholes and breccia pipes extending up into the Spearfish and higher formations. Many of these sinkholes in the Spearfish are sites of resurgent springs where the potentiometric surface is above the land surface. Dry steep-walled sinkholes are located where the potentiometric surface lies below the bottom of the sinkhole. The largest sinkholes are the result of dissolution of the thick anhydrite in the subsurface Minnelusa and consequent stoping to the surface. As the dissolution front of the Minnelusa anhydrite moves radially outward from the center of the Black Hills, and as the potentiometric surface falls to lower stratigraphic levels while the land surface is lowered by erosion, the present springs will dry up and new ones will develop down the regional dip. Abandoned sinkholes attest to the former position of these springs. The down-dip migration of these springs is exemplified by Mirror Lake where headward spring migration resulted from continued sinkhole collapse of the headwall of this steephead valley. The location of many of the sinkholes within the lower part of the Spearfish may be related to the greater fracturing of that part of the Spearfish due to downward intrusion of gypsum veinlets that developed highly fractured bedrock. The springs emerge at or near a prominent gypsum horizon. Above that horizon siltstone and shale are highly impermeable restricting the upward movement of ground water. Appreciation of the processes involved in the formation of gypsum karst should be considered in land use planning in this increasingly developed part of the northern Black Hills.

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# Exchange of Matrix and Conduit Water with Examples from the Floridan Aquifer

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## Abstract

Rapid infiltration of surface water and contaminants occurs in karst aquifers because of extensive conduit development, but contamination of ground water supplies requires loss of conduit water to the matrix. This process is also important for ground water management and for dissolution and diagenetic reactions. Many factors control exchange between conduits and matrix including the head gradient between matrix and conduits, the permeability of the matrix, the gradients of the regional water table and the conduits, and the relative elevation of the conduits and regional water table. The Floridan Aquifer, which is characterized by high matrix porosity and permeability, provides several examples.

## INTRODUCTION

Dissolution of soluble minerals and development of conduits within karst aquifers results in high permeability and allows rapid and extensive mixing between surface and ground water (e.g. Kincaid, 1997; 1998). In places, particularly where the aquifers are unconfined, ground and surface water may constitute a single body of water (Katz et al., 1997). Consequently, contamined surface water can readily infiltrate ground water supplies (Field, 1988; 1993). Additionally, surface water is commonly limited in areas of unconfined karst, causing ground water to be the major source of water supplies. Protection of the ground water can be complicated, however, by the lack of correspondence to surface water drainage divides and limited information on subsurface flow paths and rates.

Karst aquifers are characterized by three types of porosity: intergranular matrix porosity, fracture porosity, and large cavernous conduits (e.g. White, 1969; 1977; Smart and Hobbs, 1986). These different types of porosity lead to heterogeneous distribution of permeability and consequently flow rates depend on whether the flow path is through matrix, fractures, conduits, or a combination. Early work on karst systems showed that variations in discharge, temperature, chemical composition, and the saturation state of calcite of spring water could be used to separate flow paths into diffuse versus conduit systems. Diffuse flow systems occur predominately within intergranular and fracture porosity, while conduit flow occurs within conduits (Pitty, 1968; Shuster and White, 1971; 1972; Paterson, 1979). Subsequent work showed that karst aquifers can not be separated simply into purely diffuse or conduit flow but were rather a combination of these two types of flow (Newson, 1971; Ternan, 1972; Atkinson, 1977a;b). This view of karst aquifers suggests that they constitute two component systems, in which a majority of the storage occurs within matrix porosity and fractures, while a majority of the transport occurs in the large dissolution conduits (Atkinson, 1977a). Matrix flow is likely to be laminar, whereas conduit flow will likely be turbulent. Except where noted in special cases, in this paper we will lump the intergranular and fracture porosity within matrix porosity as being distinct from large conduit porosity, essentially separating the two components into laminar and turbulent flow.

Contaminant distribution and flow rates within karst aquifers are clearly influenced by the relative proportions of laminar flow within matrix and turbulent flow within conduits. Surface contaminants will rapidly enter the subsurface conduits through openings such as sinkholes and swallets (Newson, 1971), but if they flow through the aquifers within conduits, they will rapidly be discharged at springs (e.g. Meiman et al., 1988; Ryan and Meiman, 1996; Mahler and Lynch, 1999). By this mechanism, contaminants will affect the surface water quality of the spring runs, but will be rapidly flushed from the ground-water reservoirs with little long term degradation to ground-water supplies. If matrix and conduit water mix, however, contaminants could infiltrate into the matrix, resulting in long residence times within primary karst ground-water reservoirs (e.g. Katz et al., 1999). Consequently, understanding the mechanisms and rates of exchange of conduit and matrix water is vital for karst hydrogeology and water resources of these areas.

## EXCHANGE OF CONDUIT AND MATRIX WATER

The importance of conduits for flow through many karst aquifers has focused much research on

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characterizing the conduit plumbing system within the aquifer (e.g. Fig. 1). For example, variations in the chemical composition of spring water provides constraints for mathematical predictions of recharge rates and areas (Dreiss, 1989). This approach has been used successfully in Paleozoic carbonates of southeastern Missouri (e.g. Dreiss, 1989; Wicks and Hoke, 1999).

Variations in spring discharge, coupled with artificial dye tracing, have also been used to create detailed maps of the distribution of conduits and to estimate sizes of conduits and the relative contributions to flow of individual conduits within branched conduit networks (Smart and Ford, 1986; Meiman and Ryan, 1999). Such maps have been created in the Paleozoic limestones of Castleguard Meadows, Canada (Smart and Ford, 1986) and Mammoth Cave, Kentucky (Meiman and Ryan, 1999). In these dense and recrystallized rocks, intergranular matrix porosity is likely to be low, resulting in little matrix water exchanged with the conduits. Thus, the assumption of predominately conduit flow is reasonable for these well-studied Paleozoic systems.

In contrast, conceptual models that are based solely on conduit flow may be invalid in areas where carbonates have high porosity and permeability. Although there will be exceptions, high porosity and permeability are likely to be more common for young carbonates, such as Tertiary carbonate platforms of Florida, Yucatan and ocean islands. In these areas, extensive exchange may occur between matrix and conduit porosity, which would complicate a direct relationship between spring discharge and flow paths (e.g. Martin and Gordon, 2000). Areas that are characterized by significant exchange between matrix and conduits will require an understanding of the extent of the exchange, as well as the mechanisms that may control this exchange.

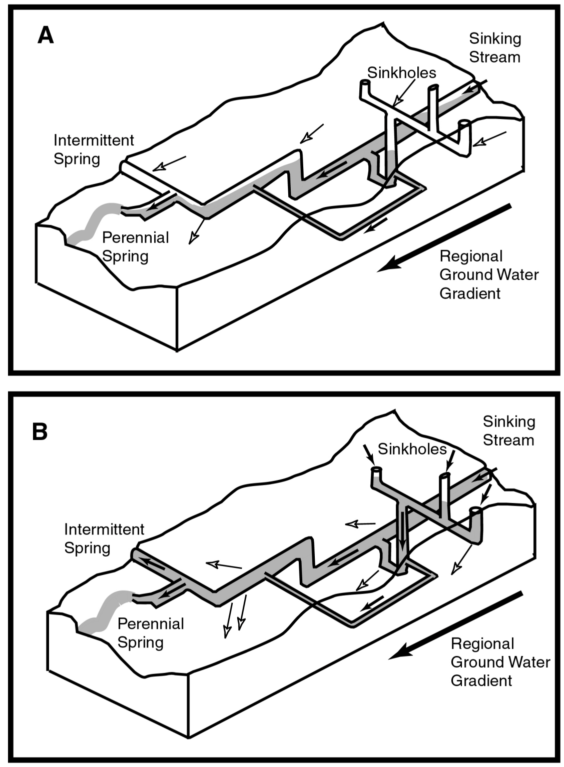


Figure 1 – Generalized diagram of the possible distribution of conduits in a karst region. The distribution of conduits is loosely based on results from Smart and Ford (1986). Solid arrows indicate direction of flow in conduits. **A**. Normal to low flow conditions when water enters conduits from matrix porosity and fractures. Some conduits may be only partially filled. Open arrows reflect flow from matrix to conduits except at constrictions where flow may be from conduit to matrix. **B**. Flood conditions when all conduits are filled from recharge into sinkholes and swallets. If head is sufficient, water would flow from conduits to the matrix, a flow path represented by open arrows. Depending on gradients, this water might become entrained in regional ground water flow.

## Possible Controls of Conduit-Matrix Exchange

Several factors are likely to control the loss and gain of water to conduits from the matrix. Most significant is the head gradient between the conduits and the surrounding matrix (White, 1999). At low flow conditions, the conduits will act as a drain from the surrounding matrix, providing a source for base flow from perennial springs (Fig. 2A). At flood conditions, the gradients may reverse, particularly where conduits are fed by allogenic recharge from sinking streams. In this case, the head within the conduit would be greater than the head of the surrounding matrix, causing water to flow from the conduits to the matrix (Fig. 2B). The water lost to the surrounding intergranular porosity and fractures may simply be stored until the head gradients are reversed. Depending on the orientation and magnitude of the regional ground water gradient and the matrix properties, flow of the water could become entrained in the slow laminar flow through the matrix (Fig. 2C).

In addition to temporal variations, the head gradient is likely to be variable along the route of flow within the conduit. These variations will depend in part on the orientation and structure of the conduits (e.g. Fig. 1).

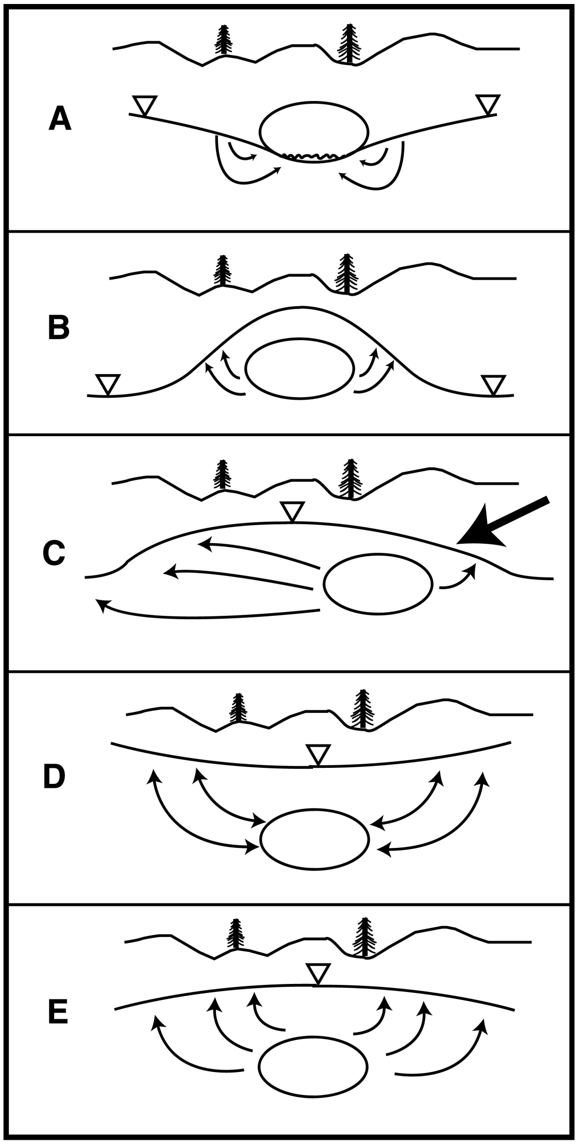


Figure 2 – Schematic and hypothetical examples of various potential controls on exchange of conduit and matrix water. Modified from White (1999). **A.** Base flow and **B.** flood with regional water table below conduit. **C.** Flood with external regional ground water gradient. **D.** Base flow and **E.** flood with water table above conduit level (i.e. permanently saturated).

For example, constrictions could increase the head within the conduits and cause water to flow into the matrix. Below the constriction, where the conduits widen, head in the conduits would be reduced, possibly allowing flow from matrix to conduits.

If conduits are located above the average elevation of the ground water table (i.e. vadose caves), then water is likely to be lost from the conduits during flooding under the force of gravity (Fig. 2A). Furthermore, this water will be permanently lost from the conduits as the flood recedes. Flood conditions within vadose caves are difficult to observe, however, because of complexities associated with fieldwork within flooded caves.

The coupling between the conduits and surrounding matrix is likely to be an important control of exchange between conduits and matrix (White, 1999). This coupling could be controlled by the hydraulic conductivity of the surrounding matrix, as well as the size of the conduits. There will be extensive exchange if the matrix is extensively fractured or dissolved, resulting in increased permeability (e.g. Wilson and Skiles, 1988). For example, as shown in Fig. 2D, water may alternately flow into or out of the conduits depending on changes in permeability of the matrix along the flow paths. Small anastomosing and branching conduits will increase the surface area of conduits relative to their volume, increasing the likelihood of exchange of water with the matrix. Large conduits are thus less likely to exchange water with the matrix than small conduits.

Other factors that could be important in the exchange include the relative elevation of the regional ground-water table and the conduits and slope of the ground-water table and the conduits (Fig. 2). The slope of the conduits and the ground-water table control the rate and direction of flow through the system. Direction of flow of the regional ground water would have to be non-parallel to the orientation of the conduits in order to entrain water lost to the matrix (Fig. 2C). Although regional ground water flow may follow relatively straight flow paths for many kilometers, the orientation of conduits is commonly curved over short distances (e.g. tens to hundreds of meters).

The physical coupling of the conduit-matrix system is clearly not static. Although the physical orientation of conduits and distribution of matrix permeability, which are invariant on short time scales, are important, the exchange of matrix and conduit water must vary with time and magnitude of recharge events. Consequently, observations of exchange between matrix and conduits must be made under widely varying conditions.

## EXAMPLES OF EXCHANGE WITH EMPHASIS ON THE FLORIDAN AQUIFER

### Flow from matrix to conduits

Some of the first studies to separate spring discharge into conduit and diffuse flow components focused on springs in the fractured limestones of Mendip Hills, England and other regions of Great Britain (Newson, 1971; Atkinson, 1977b). These studies showed that some springs cannot be classified into purely diffuse or conduit flow (e.g. Shuster and White, 1971). In a study of subsurface erosion, Newson (1971) found that water discharging from springs ranged from nearly all allogenic water recharged to swallets (e.g. "quick flow" which would represent conduit water) to nearly all water derived from the matrix porosity (referred to as "percolation water"). Although the fractions of these two water sources ranged widely through the group of springs being studied, they imply that the conduits gain water from the matrix feeding the springs. Similar results were obtained by Atkinson (1977b), who showed that water discharging from springs in the Mendip Hills is sourced approximately 50% from flow through conduits and 50% from slow percolation from the matrix.

Additional evidence for loss of water from the matrix to conduits comes from a study of environmental tracers in the Santa Fe River of north-central Florida (Martin and Dean, in press). Across north-central Florida, the Floridan Aquifer is separated into confined and unconfined portions with the semi-confined boundary referred to as the Cody Scarp (Fig. 3; Puri and Vernan, 1964). Discharge from the Santa Fe River averages ~10 $m^3$/sec once it emerges from an ~5 km passage underground where it flows across the Cody Scarp (at a first magnitude spring called the River Rise; Fig. 4). Discharge can be extremely variable through time, however, ranging from less than 1 $m^3$/sec to more than 100 $m^3$/sec. Furthermore, the discharge increases rapidly downstream from its resurgence point because of numerous springs that flow into the river. Because of the continuous flow of surface water into conduits at the River Sink (Fig. 4), as well as the large variations in recharge from base flow to flood conditions, the Santa Fe River provides an ideal field area to study the exchange of water between conduits and matrix through observations of changes in thermal and chemical compositions of the river water (Martin and Dean, 1999; in press).

At low flow conditions, changes in the natural chemical composition, temperature and discharge volume of the water as it flows through the subsurface towards the River Rise suggest that only 4% of the water is contributed from the river sink, while as much as 96% of the resurgent water comes from other sources (Martin and Dean, in press). The largest source is suspected to be a contributing conduit system, which has recently been mapped by cave divers (Fig. 4). This conduit system receives little direct recharge from the surface and consequently, most of the water in the conduit must derive from the matrix (Martin and Dean, in press). The fraction of river and other water depends strongly on the discharge of the river. At intermediate discharge, the fraction of other water drops to 27% with the remaining water originating from surface water flowing into the River Sink. The fractions of different water sources have not yet been measured at flood stage. These results support findings by Newson (1971) and Atkinson (1977b) that much spring water can originate from the matrix.

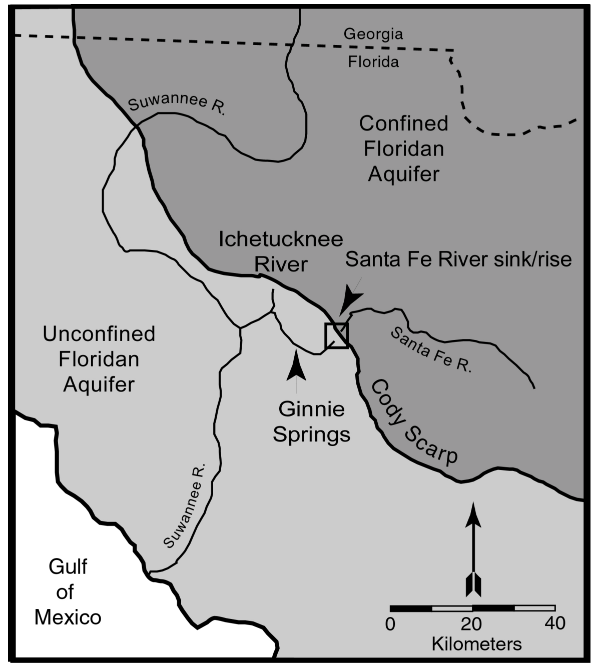


Figure 3 – Regional map of north-central Florida showing the location of three study areas – Santa Fe River sink/rise system, Ginnie Springs system, and Ichetucknee River. The darkly shaded region represents confined Floridan Aquifer and the lightly shaded region represents the unconfined Floridan Aquifer. The boundary is a semi-confined region referred to as the Cody Scarp.

These results also support findings of what appears to be substantial contributions by diffuse flow to springs discharging from the Floridan Aquifer (Martin and Gordon, 2000). The Ichetucknee Springs group discharges along the Cody Scarp to the Ichetucknee River (Fig. 3). Cave diving exploration and dye trace studies indicate that many of the springs in the group are connected to conduits. Annual and storm chemographs reflect little change in the composition of the spring water through time, however, as would be expected from purely conduit-fed springs (e.g. Pitty, 1968; Shuster and White, 1971; 1972). Consequently, the conduits that feed these springs are suspected to be predominantly sourced from the matrix (Martin and Gordon, 2000).

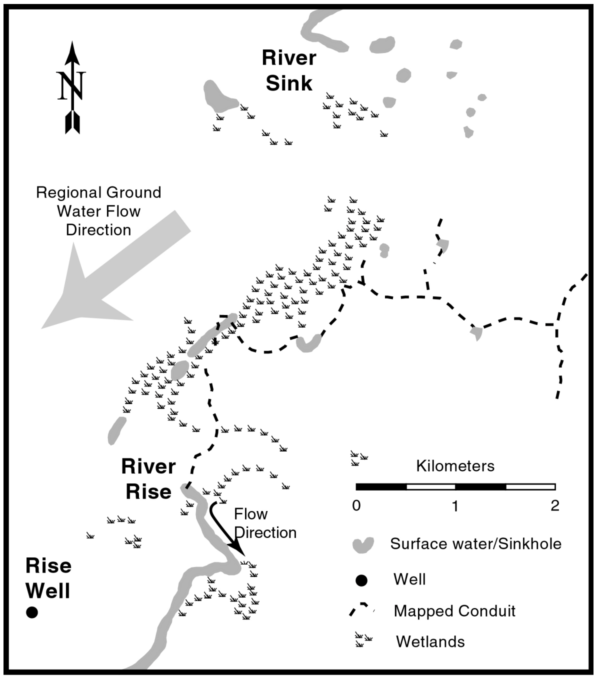


Figure 4 – Sketch map showing the location of Santa Fe River at the sink/rise system, surface water bodies that commonly represent sinkholes, distribution of wetlands and the extent of the mapped conduit system. The river sinks into the subsurface at the location labeled River Sink, and re-emerges at the location labeled River Rise. Modified from Martin and Dean (in press).

## Flow from conduits to matrix

The converse situation, where water is lost from conduits to the matrix is more difficult to study and document, but has important ramifications for regional water quality. Contaminants flowing into the conduits would enter the matrix porosity along with the water and subsequently would require long periods of time to be flushed from the system. One possible example of matrix pollution by $NO_3$ contamination is illustrated by the Floridan Aquifer in north-central Florida where $NO_3$ concentration of spring water have been increasing through time (Katz et al., 1999). The cause of the elevated concentrations is not clear, and could reflect in part an increasing number of sources. The increase could also reflect slow flushing of contaminated water from the matrix.

The regional ground water chemistry of the Floridan Aquifer provides another example of how surface and ground water may mix by loss of water from conduits to the matrix. Along the Cody Scarp, numerous streams flow into the subsurface through sinkholes and either disappear completely or re-emerge, similar to the Santa Fe River. This allogenic water greatly influences the chemical composition of the regional ground water (Lawrence and Upchurch, 1976; 1982; Upchurch and Lawrence, 1984). A detailed statistical study of the distribution of major and minor element concentration in ground water found that water chemistry is controlled by several types of fluid-solid reactions and sources. In particular, infiltration by surface water along the Cody Scarp leads to water that is undersaturated with respect to carbonate minerals and leads to dissolution reactions and karstification. Much of the recharged water along the Cody Scarp flows into sinkholes and subsequently into conduits. Samples collected from water supply wells typically pump from the matrix porosity, reflecting significant variations in chemical composition of the ground water. These variations in chemical composition across the Cody Scarp qualitatively suggest that water is lost from conduits to the matrix.

Wilson and Skiles (1988) have experimentally studied the question of the loss of water from conduits to matrix. In their study of Ginnie Springs Group in north-central Florida (Fig. 4), rhodamine WT dye was injected into several wells drilled through conduits that range in size from 0.3 to 1.8 m high. Average dye velocities to the discharge point at three springs within the group ranged from 7.7 to 32 m/hr, reflecting primarily conduit flow. Mapped large cave passages are limited in the region, however, suggesting that flow along the entire flow path was not restricted to conduits. In addition, the dye return curves exhibited tails that were 5 times longer than the time between initial return and peak return, further suggesting that some flow occurred through matrix rather than conduits. Wilson and Skiles (1988) suggested that this matrix flow occurred in “sponge-like” dissolutional openings, and concluded that the flow was darcian in character.

Martin and Dean (in press) found that water in a water supply well located down the regional gradient from the conduits at the Santa Fe River became increasingly dilute in the concentrations of conservative solutes following a major flood (Fig. 3). This dilution was interpreted to suggest that water was lost from the conduits during the flood. On the basis of the observed time lag, Martin and Dean (in press) estimated that the rate of flow through the matrix was on the order of 0.4 to 2.7 m/hr. Although this range is an order of magnitude slower than that observed by Wilson and Skiles (1988), some of the flow they measured must have occurred as rapid flow through conduits. It is also likely that the matrix permeability of the Floridan Aquifer varies greatly over short distances. The temporally and spatially variable loss of water from conduits to the matrix may be an important control on the distribution of chemical compositions across the region, as was observed by Upchurch and Lawrence (1984).

## SUMMARY

Although conduit-based models may be appropriate for low matrix permeability and porosity limestones, several studies of flow in karst aquifers suggest that significant volumes of water are exchanged between matrix and conduits. Quantifying this exchange is critical for developing conceptual and numerical models of contaminant transport and storage and the management practices in these aquifers. The number of studies is relatively limited, however, indicating the need for additional work. For example, the parameters that control the direction of water exchange, i.e. whether it is lost or gained from conduits, are currently poorly constrained. Controls include the permeability and distribution of fractures in the matrix, the gradient of conduits and regional water table, the variations in sizes of the conduits, the orientation of regional ground water flow relative to that of the conduits, and the extent of recharge into sinkholes.

Particularly important to assessing water quality in karst areas will be the ability to determine the volume of water lost from conduits to the matrix. This water has the greatest potential for dissolution of the matrix porosity, as well as contamination of the water in the matrix porosity, commonly the primary water supply. It is difficult to measure the volume of water lost to matrix because mixing within the matrix is slow, resulting in heterogeneous water composition, and environmental and the large volumes of water stored in the matrix porosity can dilute injected tracers. The Floridan Aquifer, with high matrix porosity and permeability provides an ideal location to determine the potential for the magnitude of this exchange.

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# Hydrogeologic Characterization of a “Transitional” Karst Aquifer, South-Central Louisville, Kentucky

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Carbonate aquifers typically exhibit a continuum in ground-water flow that ranges between quick flow through solution conduits and solution-enlarged fractures and slow flow through fine fractures and intergranular pores. Hydraulic properties and water quality often change at different locations in carbonate aquifers depending on the degree of solutional (karst) modification—that is, the relative proportion between quick flow and slow flow. Either end member can present special difficulties to hydrogeologic and contaminant transport characterization. The transition between a quick-flow dominated karst aquifer and a slow-flow dominated fractured-carbonate aquifer is examined in this study.

The aquifer system is composed of westward-dipping Silurian-Devonian limestones and dolostones in Jefferson County, Kentucky. Moldic porosity due to calcite-to-dolomite mineral replacement has developed in parts of the aquifer system, but most ground-water flow is conducted through solutionally-enlarged, horizontal fractures. The quick-flow dominated part of the aquifer system occurs to the east, where the rocks crop out near the surface, are more weathered, and are recharged by infiltration and runoff into sinkholes. The slow-flow dominated part of the aquifer system is to the west, where the carbonates are overlain by Devonian-age carbonaceous shale. Solutional modification of the carbonates diminishes toward the west, as the thickness of the shale confining unit and distance from the recharge area increases. A transitional zone can be delineated in the aquifer system where the influence of quick-flow dominated karst decreases and the hydrogeologic characteristics of the aquifer system revert to that of a low-permeability, fractured limestone. This transitional zone is marked by a significant change in dominant water-quality type, from calcium-magnesium-bicarbonate water to sodium-chloride water, changes in the solubility indices of dissolved mineral species, and in tritium and stable isotope values.

# PROPOSED NATIONAL ATLAS KARST MAP


By Randall C. Orndorff, Jack B. Epstein, and David J. Weary

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The U.S. Geological Survey (USGS), in cooperation with the National Park Service (NPS), is preparing a digitized 1:7,500,000-scale map showing the distribution of karst in the conterminous United States. The present “Engineering Aspects of Karst” map (Davies and others, 1984), “Karstlands” (Davies, 1970) and “Cavern Areas” (Davies, 1970) (fig. 1) in the National Atlas will be revised to better display surficial karstic features. The map will be combined with a map of NPS facilities to better address the needs of the NPS. The map could be hot-linked to references of detailed source maps showing karst within states and counties. A detailed map of karst for the “Atlas of Appalachia” will be prepared by the USGS in cooperation with by a consortium led by Morehead State University, Kentucky. Additionally, a generalized description of the karst within each park and the surrounding area that may have affect on management within the park could be prepared. This data could be also hot-linked to the map. Features that may be included are exposed carbonate and evaporite units, intrastratal karst, karst beneath surficial overburden, and percentage area covered by karst. A preliminary step in the preparation of the map will include evaluating geologic and karst maps of all states. A summary of the geology and karst features, along with an annotated bibliography, of each of the NPS facilities could be prepared, excluding those facilities that are “cave parks” and for which adequate information is already available.

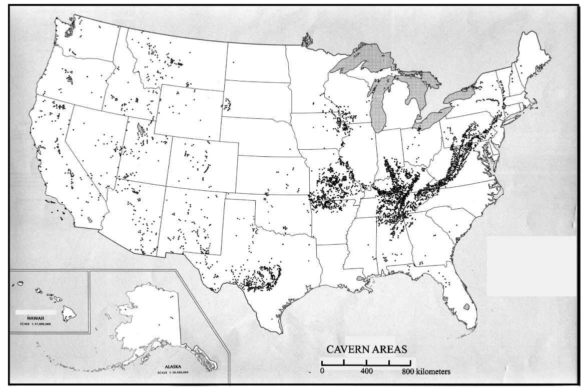


Figure 1. Cavern areas of the United States (from Davies, 1970).

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# Using Science to Change Management Perspectives at Carlsbad Caverns National Park

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## Abstract

Research at Carlsbad Caverns National Park (CCNP) has led to dramatic changes in the management of cave and karst resources in the park. Surface contaminants found in the pools of Carlsbad Cavern have led to a re-evaluation of the infrastructure above the cave and could lead to the removal of buildings, parking lots, and utility lines above the cave. The discovery of cave-adapted microbes in remote areas of some caves has changed how exploration and mapping is conducted. Lint accumulation in Carlsbad Cavern has been found to harm speleothems and may support a thriving non-native microbial population.

Multidisciplinary research has revealed many unforeseen human impacts on the cave resources of the park. The park is changing its management practices to protect these resources while still providing a quality experience for the half-million people who visit the park each year.

## INTRODUCTION

One of the goals of the Cave Resources Office of Carlsbad Caverns National Park is to protect the caves and cave resources of the park (Figure 1). The park must balance this goal with the need to provide over a half-million visitors per year with a memorable and safe experience in the park. In addition to visitation to Carlsbad Cavern, the park provides access to some caves for recreation and for scientific research. The management of the park caves must weigh the impacts of these activities against their benefits.

Cave management throughout the National Park Service (NPS) has traditionally focused on highly visible and easy to measure impacts such as broken speleothems, high-traffic areas, and litter. In karst areas characterized by sinkholes, springs, and sinking streams, the NPS has managed the surface to prevent impacts from fuel spills, industrial waste, and sewage leakage into the ground above the caves.

Recent research at CCNP has identified other issues that the park needs to manage. A study of the hydrology and hydrochemistry of Carlsbad Cavern has identified sources of contamination originating from the park facilities above the cave. Carlsbad Cavern is seen by more than 500,000 visitors a year, by far the most visited feature in the park. Park manangers have found that visitors leave behind tens of kilograms of lint in Carlsbad Cavern every year. These discoveries have changed the way the park manages the cave resources of Carlsbad Caverns National Park.

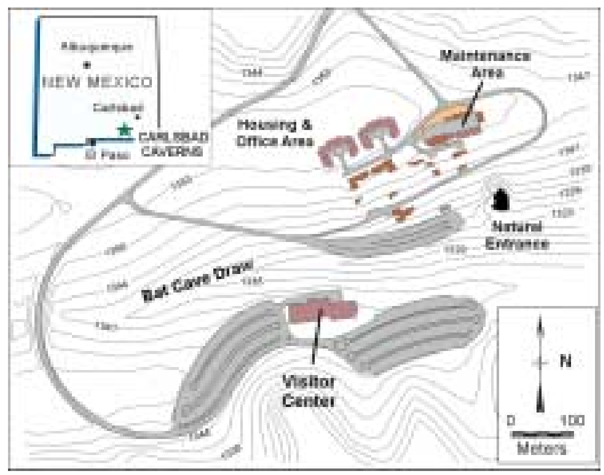


Figure 1—Location of Carlsbad Caverns National Park and major features of the developed area above the cave

Microbiological investigations in Lechuguilla Cave and Spider Cave have led to the discovery of previously unknown bacteria which may lead to cures for some human diseases, but which can be decimated by just a few contacts with people.

## INFILTRATION STUDY

An infiltration study was performed in the area around Carlsbad Cavern (Figure 2) as part of a Colorado School of Mines master's thesis (Brooke, 1996) and an investigation by the International Ground Water Modeling Center (van der Heijde et. al., 1997). This study identified areas in the cave threatened or already affected by contamination due to surface facilities, as well as probable pollution sources. The most affected

In Eve L. Kuniansky, editor, 2001, U.S. Geological Survey Karst Interest Group Proceedings, Water-Resources Investigations Report 01-4011, p. 47-51

areas of Carlsbad Cavern are: 1) Left Hand Tunnel, 2) Main Corridor and New Section, and 3) locations from the New Mexico Room to the Big Room. The study reported:

> Most of the unnaturally high concentrations of aluminum, zinc, total organic carbon, and nitrate... can be related to rather chronic, relatively low-level, releases at specific locations at the surface.. .A variety of accident, spill and leakage scenarios can threaten the water quality in the cavern, and even public health. Major potential sources identified in this study are: 1) leaks in the sewer lines; 2) spills and vehicle fires with subsequent contaminated runoff from the public parking lots and road segments in the western part of the [developed] area; and 3) spills, leaking tanks, fires and other accidental releases from the maintenance yard.

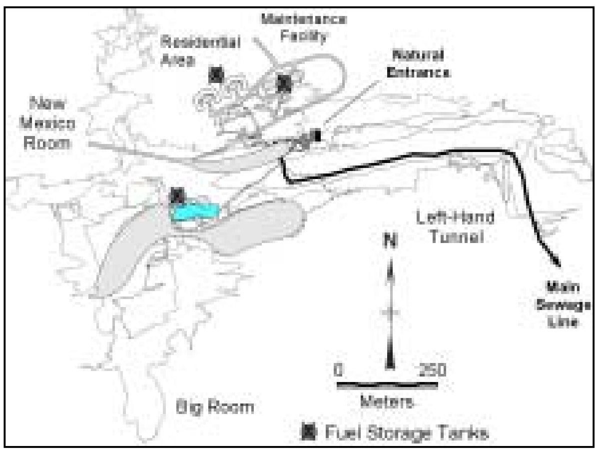


Figure 2—Location of surface features to Carlsbad Cavern.

Spills of hazardous materials, including oil and gasoline, into the subsurface could pose a potential threat to visitors. Such a danger could cause the cave to be closed to visitation until the danger was eliminated. It could also severely damage or destroy cave biota and associated ecosystems.

Total wastewater generated by visitors and staff in the park is 38 million liters per year. The sewer lines in the park vary in size, material, and condition. Some pipes are more than 65 years old and leak to a great extent. The collection system carries raw sewage from the residential and maintenance areas to a small lift station located in Bat Cave Draw near the natural entrance to Carlsbad Cavern. The lift station forces the sewage up to the ridge of the escarpment. Sewage flows by gravity from the visitor center along the escarpment ridge and drops off the ridge to sewage lagoons at the base of the escarpment. When the system backs up, sewage flows out of manholes and onto the ground. These leaks and spills supply raw sewage to the groundwater system and pose a threat to human health and water quality.

The main parking lots have the capacity for over 900 cars, 63 recreational vehicles, and approximately 152 meters of unmarked space that can be used for either recreational vehicles or cars. There is also an average of 25 cars parked in the residential housing area on the north side of Bat Cave Draw. These cars are generally parked all day and not for just the 4 hours a day the average visitor is parked. The remaining cars, however produce an effective visitation of 72 cars per day or 26,026 cars per year (Bremer 1988). This means that resident parking accounts for almost 14% of the parks yearly vehicle use.

The parking lots not only alter natural infiltration patterns, they collect and store hazardous materials generated by automobiles, maintenance operations, and residential activities. Between rainstorms, oil, fuel, antifreeze, and other contaminants accumulate on the pavement. A trash vehicle is usually parked behind the visitor center to collect waste for the concessionaire. Fluids from the waste leak and collect on the pavement. Kitchen floor mats and other items from the concessionaire restaurant kitchen are washed daily behind the visitor center. This contaminated water is allowed to drain into Bat Cave Draw. The maintenance facility makes wide use of solvents and other materials required for vehicle maintenance that may be spilled on the ground.

Rainwater carries contaminants off the pavement and into the fractured limestone above the cave. Most of the contaminants are carried by the first 13 mm of rainfall. The parking lots collect rainwater from a total of 42,550 $m^2$. This results in 540,000 liters of contaminated water entering the groundwater system in every 13 mm storm. There is an average of 10 storms per year that produce more than 13 mm of rain (Bremer 1998), making a total of 5.3 million liters of contaminated water entering the groundwater system every year. Some of the contaminants are absorbed by the thin soil and the rock, but unmitigated exposure to these materials will lead to contamination of the groundwater and the cave (Bremer 1998). Leaks from sewage lines and potential leaks from fuel storage tanks add to the amount of contaminants that may enter the cave from the developments above.

The park has proposed modification of the developed area above Carlsbad Cavern. The purpose of this is to: 1) protect the groundwater and cave resources from continuing chronic exposure to contamination; 2) protect the cave resources from potential catastrophic contamination; and 3) protect visitors to Carlsbad Cavern from potential hazardous conditions due to contamination. To achieve these goals, the park is preparing an Environmental Assessment (EA) with alternatives to reduce the impacts from the park facilities on the cave.

To address parking lot runoff issues, the park is planning on treating some of the runoff and eliminating some paved areas altogether. The parking lots near the visitor center will be resurfaced to drain southward, away from the cave. The park will install oil-grit separators at the drainage points for these parking lots to treat the water before it enters the ground. The large parking lot in Bat Cave Draw has been closed to visitors in order to reduce the amount of vehicle fluid buildup. The EA proposes removing most of this parking lot and replacing it with a bus turnaround area and handicapped access path to the natural entrance. The turnaround would be more than 300 meters further west than it is now. The majority of the paved surface will be removed and replaced with natural vegetation to help restore natural drainage and infiltration conditions.

The sewage collection system in the housing and office area north of Bat Cave Draw is going to be replaced with new lines that will not be as prone to leaks. One option of the EA proposes that the main sewage line be rerouted southward to minimize the exposure of the caves to sewage leaks from this line.

The park has made important management decisions to protect the cave by placing funding priorities on sewage line repair and restructuring of the parking lots. Park residents are restricted from using pesticides, and other household chemicals which may be harmful to the cave. Residents are also restricted from maintaining vehicles and other activities which could spill hazardous materials on the ground. The park has developed a space allocation plan which will remove most of the residents from above the cave. This will reduce the amount of sewage in the system and will effectively reduce the number of vehicles parked above the cave on the north side of Bat Cave Draw. The park has also used this study as a springboard to make significant changes to park facilities through the Environmental Assessment.

## Lint Accumulation in Carlsbad Cavern

For many years, people have noticed a gradual discoloration of speleothems in Carlsbad Cavern. Close examination has revealed large accumulations of lint, clothing fibers, skin, and hair left behind by the thousands of cave visitors each year. This lint builds up on cave walls and speleothems and makes them appear dull and gray with obvious lumps of material. Organized efforts to clean up these deposits of lint since 1988 has removed over 70 kg of lint and more than 100 kg of other litter (Jablonski, et al., 1993).

During a lint cleanup in 1991, volunteers noticed that the calcite beneath the lint was pitted and had started to deteriorate (Jablonski, et al., 1993). Further investigations showed that the lint was a very good source of organic material for microbes, mites, and spiders. Microbiologists have suggested that the breakdown of the lint was probably generating organic acids that dissolved the calcite. The large amount of organic material may support a large population of microbes that thrive in a high-organic-energy environment. These microbes can out compete the native, low-organic-energy microbes and decimate their population.

Several different methods of reducing or controlling lint accumulations were investigated both at Carlsbad Caverns and at Wind Cave National Park in South Dakota. Researchers tried applying a vacuum to visitors to quantify lint deposition and to determine the effectiveness of lint removal using flowing air. Most methods for cleaning lint from visitors would require careful control of air circulation which is not possible in either Carlsbad or Wind Cave, so ways to reduce the migration of lint were examined.

Researchers looked at how lint migrated along the trail and onto the cave walls. They found that short rock walls along the sides of the trail contained much of the lint. In these areas, trail cleaning prevented the lint from breaking down into small enough particles to be carried into the air and deposited higher on the walls.

Based on the results and recommendations of these studies, Carlsbad Caverns is in the process of constructing rock walls along most of the five-kilometer-long trail. Twice a year, this trail is vaccumed with a HEPA vacuum to remove any lint that may have accumulated within the walls. In addition to these measures, a volunteer group works in the cave during a week-long “Lint Camp” to remove lint and other litter from throughout the visitor trail system. In places where there is a problem with visitors urinating in the cave and other places where the trail may become slick, water is used to clean the floor. In the past, this water was allowed to run off the trail, but the park is now studying methods to collect and remove this waste water from the cave. With these measures, the park hopes to significantly reduce the impacts from lint on the cave.

## MICROBIOLOGICAL RESEARCH IN LECHUGUILLA CAVE

Researchers from NASA have been looking at extreme environments across the planet that may be analogs for life on other planets. A team of scientists, primarily from the University of New Mexico and Biomes, Inc., have found previously unknown bacteria living on the rock and in the pools of Lechuguilla Cave. These bacteria survive hundreds of meters below the ground with no light and no organic input from the surface.

Some bacteria exist in thick mats of corrosion residue that line the walls, ceiling, and floors of many places in Lechuguilla Cave (Northup et al., 1994). These bacteria utilize small amounts of iron and manganese in the limestone bedrock to derive their energy (Northup et al., 1997). The researchers have found that almost one-third of the cells in the cave are actively respiring and that the process of microbial corrosion is still going on.

The pools in Lechuguilla Cave are also teeming with life (Northup et al., 1999). The bacteria in these pools have to compete fiercely with each other for the few nutrients that exist. To do this, they release enzymes that kill the competition. Some of these enzymes have been tested under laboratory conditions and have been found to attack leukemia cells. Much more testing needs to be done, but these bacteria may lead to a cure for some human diseases.

People carry foreign bacteria into caves on skin, hair, and clothing fibers. When they are shed into the caves, these bacteria out compete native microbes for food and destroy their populations. The magnitude of human impact was not known until some of the pools in Lechuguilla Cave were discovered. In some pools, the native microbe populations have been decimated by only a handful of contacts with cave explorers.

The park has changed it policy to require that everyone who enters Lechuguilla Cave has clean clothes and equipment to prevent microbes from other caves from being introduced into the Lechuguilla. Cave explorers and scientists often camp in the cave for several days. To minimize their impact, they are now required to eat and sleep on drop cloths to catch food, skin, and hair. Cave explorers and researchers are also encouraged to wear bandanas to contain hair and are required to eat their food over plastic bags to catch fallen crumbs. Some areas where there has already been contamination have been closed off to determine if, and how long it takes the native microbes to recover.

Cavers are restricted from getting near any pools found during exploration. When a pool is discovered, explorers report it to the park and to scientists studying the microbes. Scientists approach the pools wearing Tyvek clean suits and set up slides that sit in the cave for up to five years. The slides are collected later and the bacteria cultured in a lab where scientists hope to study them further.

## SUMMARY

Research in the fields of geology, hydrology, hydrochemistry, microbiology, and cave restoration have revealed negative impacts on the cave resources of Carlsbad Caverns National Park. The park has changed the way it manages both the surface and subsurface to mitigate these impacts. The park has used this research to engineer solutions to some of the problems of infiltration and contamination. Scientific research has changed and will continue to change the way the park manages both the caves and the surface around them.

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# National Cave and Karst Research Institute--History, Status, and Future

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## Abstract

The National Cave and Karst Research Institute Act of 1998 mandated the National Park Service to establish the Institute. The Act stipulated that the Institute will be located in the vicinity of Carlsbad Caverns National Park in New Mexico (but not inside Park boundaries), and that the Institute cannot spend Federal funds without a match of private funds. The main purposes of the Institute are to further the science of speleology, to encourage and provide public education in the field, and to promote environmentally sound cave and karst management. The Interim Director for the Institute reported in July 2000 for a two-year period to define the purview and scope of operation, design an organizational structure, form partnerships, find funding sources, find a physical facility, and define research needs.

## AUTHORIZATION

The National Cave and Karst Research Institute (hereafter referred to as the Institute) was mandated by act of Congress in 1998 (Public Law 105-325) under the organizational structure of the National Park Service (NPS). The 1998 Act was the culmination of many years of effort by the caving community, private and Federal, to have legislation enacted that facilitated gaining a scientific basis for cave and karst management. While the 1998 Act authorized the Institute, no funds have been appropriated yet.

During the decade before the 1998 Act was passed, other cave and karst related acts were passed. The Federal Cave Resources Protection Act of 1988 (Public Law 100-691) directs the Secretaries of the Interior and Agriculture to inventory and list significant caves on Federal lands, and provides a basis for protecting caves. Another enactment in 1990 (Public Law 101-578) further directed the Secretary of the Interior, through the Director of the National Park Service, to establish and administer a Cave Research Program, and to prepare a proposal for Congress that examined the feasibility of a centralized National Cave Research Institute. This proposal was presented to Congress in December 1994 and is the basis for the 1998 Act establishing the mandate for the Institute. The Lechuguilla Cave Protection Act (Public Law 103-169), passed in 1993, recognized the international significance of the scientific and environmental values of the cave.

Primary stipulations of the 1998 Act were that the Institute be located in the vicinity of Carlsbad Caverns National Park in New Mexico (but not inside Park boundaries), and that Federal funds must be matched by private funds.

The Interim Director for the Institute was recruited in July 2000 for a two-year period to move forward with NPS efforts to establish the Institute by defining the purview and scope of operation, designing an organizational structure, forming partnerships, finding funding sources and a physical facility, and defining research needs.

## MISSION AND GOALS

The mission provides a framework for the Institute to achieve its congressionally defined goals and to guide development of an appropriate scope of activities in the National interest:

> *The National Cave and Karst Research Institute furthers the science of speleology by facilitating research, enhances public education, and promotes environmentally sound cave and karst management.*

The goals (purposes) of the Institute are clearly and simply stated in the text of the 1998 Act. Following are expanded statements of goals that provide a broader view of the operational intent of the Institute:

- Further the science of speleology through coordination and facilitation of research.
- Provide a point-of-contact for dealing with cave and karst issues by providing analysis and synthesis of speleological information and serving as a repository of information.
- Foster partnerships and cooperation in cave and karst research, education, and management programs.
- Promote and conduct cave and karst educational programs.

In Eve L. Kuniansky, editor, 2001, U.S. Geological Survey Karst Interest Group Proceedings, Water-Resources Investigations Report 01-4011, p. 52-55

- Promote national and international cooperation in protecting the environment for the benefit of caves and karst landforms and systems.
- Develop and promote environmentally sound and sustainable cave and karst management practices, and provide information for applying these practices.

## ACCOMPLISHING THE GOALS

When the Institute becomes operational, a variety of activities can be undertaken to accomplish the stated goals. Some of the major activities could include:

- Develop and support short- and long-term speleological research studies by identifying research needs of National interest. Promote research focused on the needs through Institute research scientists, visiting scientists, and/or through a grant process.

An appropriate niche for the Institute is to support and encourage a body of research that would be the most useful nationally and for a variety of agencies and organizations. This focus would not preclude support of a wide variety of research topics, but research within the identified needs would be favored for grants and support. Issues for research would be determined by querying the needs and problems of a variety of agencies that manage caves and karst areas, and by consulting with universities and private organizations.

- Establish a comprehensive cave and karst library, research bibliography, and list of experts in speleological research.

No single institution is likely to be able to accumulate all references and knowledge on a particular topic; however, it is the goal of the Institute to amass as comprehensive a reference library as possible to provide a single source to research scientists and agencies. Linkages to other institutes and organizations housing resource libraries would also be a means to expand the virtual collection of the Institute. A fully functional research library is envisioned; however, a system of flexible access that requires only a minimum of staff to design, support, and maintain is critical. A database of scientists and research specialists could be an important service provided to those needing studies done or to scientists looking for collaborators.

- Develop a centralized data storage and retrieval system, and establish a comprehensive cave and karst information database. Develop methods to access and retrieve speleological data and information from worldwide sources. Establish a cave management information database.

Providing quick and easy access nationally and internationally to data and information would be critical to the information-management function of the Institute, and to facilitate synthesis of research findings. The Carlsbad location may not be accessible to all that need the resources, thus, remote and computer access to databases will be important. It is important to emphasize that the Institute would not hold proprietary information, such as cave entrance locations, that are not already publicly available.

- Sponsor national and international speleological and cave and karst symposia and cave and karst management seminars, and provide educational material.

Focused meetings and seminars would facilitate interaction among scientists to further the science, and provide a venue to convey management strategies to resource managers and decision-makers. A variety of cave related educational publications, visual materials and videotapes, and educational displays could be produced or sponsored and targeted to different audiences: scientists, managers, schools, and the public, who need to understand the implications of resource use or misuse.

- Establish and maintain contacts in other agencies, universities, corporations, and private sector groups engaged in activities related to caves and karst areas.

Formal agreements would be developed with other agencies having cave management responsibilities or research capabilities, with Universities and with cave and karst related organizations. Informal contacts also would be maintained with other organizations that have interest in the Institute programs.

- Develop a funding base to support the goals of the Institute.

A fund-raising plan will be developed that must be active and ongoing in order to provide adequate operational, research, and educational funds for the Institute. In addition to NPS funds, contributions and in-kind services would be solicited from other Federal agencies, State and local agencies, Universities, private organizations and individuals, and corporations. In order to support high level research in the National interest and viable educational programs, funding needs to be accumulated in a reserve that can be prudently budgeted consistent with the annual Federal funding cycles. Federal funding, as stated in the 1998 Act, can only be spent in the amount that it is matched by non-Federal funds.

## TIMELINE FOR FULL IMPLEMENTATION

The Institute will pass through several phases before it becomes a recognized force in the research community with the ability to sponsor a wide range of activities. The *Interim phase* is anticipated to span about three years (August 1999 to August 2002). This phase began when a Steering Committee convened to articulate expectations for the Institute and to draft specifications for recruitment of an Interim Director, and will end when the Interim Director completes the initializing tasks.

The *Gearing Up phase* is likely to take one additional year (2003), and would consist of staff recruitment, move into a building (possibly a temporary facility), initial operational setup, and the transition from the Interim Director to the Director. If funding and ability to operate permit, research grants could be distributed during this phase and the real work of the Institute can begin.

The *Basic Institute phase* would take another one to two years (2004-05) while the experience of the staff and the capacity of the Institute gradually increase, and financial resources for full operation are accumulated. If a building is constructed, it may be completed during this phase. A grant process would be operational, and results of research supported by the earliest grants may become available.

The *Fully Operational phase* should be attained by 2006, when the Institute becomes a significant and recognized resource in cave and karst research, education, and support of cave and karst management.

## PLANS AND ACTIVITIES--INTERIM PHASE

*Define the purview and scope of operation:* Discussions will be held with a variety of interested individuals and organizations to help determine the most appropriate activities for the Institute to undertake. The question of the Institute being only a granting organization or, additionally, having an in-house research staff, will be explored. The relation of the Institute to other institutes and organizations will be defined in conjunction with those groups.

*Design an organizational structure:* Staffing would be based on the scope of operation determined for the Institute. An important factor in the size and scope is the level of funding available to support the operation. Business models of other research institutions will be studied for ideas and to determine the most appropriate model for this Institute to adopt.

*Form partnerships:* Formal agreements with agencies, universities, and other organizations will be signed. A concerted effort is being made by the Interim Director to meet with a wide variety of groups. The Interim Director also is making international contacts in order to lay a foundation for international collaboration in cave and karst research and information exchange.

*Funding sources:* Private or state/local government funds must match Federal funds to support the Institute's operations. To accomplish this, a "bank account" of non-federal funding needs to be accumulated before matching funds would be available from Congress (2002 Federal funding requests will be developed early in 2001). A plan is in development to target likely non-federal donors and collaborators and establish partner agreements. Partnerships and collaborations with universities will be explored. State and other governmental levels are potential sources of non-federal funds for operations and for grants. Private or corporate contributions also are potential sources of non-federal funds for operations or grants.

*Find a physical facility:* The Institute has the option of renting space (or accepting existing space as an in-kind contribution) or constructing a building. The City of Carlsbad is a major player in the building decision, because the location of the Institute is a planning and economic factor in Carlsbad. The City and New Mexico Tech are collaborating to request building funds from the New Mexico legislature. If successful, the physical facility for the Institute would include office, laboratory, library, and computer space. If the request for full State funding is successful, a science center also would be built, which would include an auditorium, an IMAX theater, a museum, and an interactive science center focusing on cave and karst systems and resources.

*Assess research needs*: The Institute can provide a national scope and overarching goals to cave and karst research. These needs would be compiled through discussions with a wide variety of interest groups, scientists, and resource managers. Focus groups may be hosted at national professional meetings to provide a forum for input into the research needs.

## CURRENT STATUS

At the time of this writing (January 2001), the Interim Director has been in place for six months—one-quarter of the appointment period. Is one-quarter of the work accomplished? Probably not in all cases, but this is difficult to judge. Tasks in the initial phase are not proceeding in a linear fashion. Many aspects have been started or drafted, but none are complete. Each meeting with groups or individuals interjects new ideas into the process and revisions are made.

*Purview and scope:* A Federal Working Group has been formed to assist the Interim Director in developing the operating plan for the Institute. The Group is comprised of representatives from National Park Service, U.S. Geological Survey (USGS), Bureau of Land Management (BLM), Fish and Wildlife Service (FWS), and U.S. Forest Service (USFS). Additionally, each person in the Group has responsibility to represent and communicate with non-Federal constituency groups.

*Organizational structure:* It is envisioned that the initial Institute would be a staff of six or seven people that, in addition to the Director, might include a Scientist Coordinator, Education Coordinator, Computer/GIS, two administrative or support positions, and a Librarian if the library collection is significant. Some operational support, such as contracting and other administrative duties, may need to be supported out of other NPS units for a time. A voluntary Science Advisory Board is likely to be part of the organization, that would assist in defining research priorities and with a grant review and ranking process. Additionally, a volunteer Strategic Advisory Board may be formed to advise the Director of the Institute on the priority activities to focus on each year.

*Partnerships:* Partnerships may be informal, a mutual verbal agreement for parties to support each others missions and goals, or they may be formal agreements with specific tasks and activities. A partnership agreement has been signed between the Institute and New Mexico State University, which has a campus in Carlsbad, for a small amount of office space and administrative support during the interim and gearing up phases. A formal agreement among the Institute, New Mexico Tech, and the City of Carlsbad is under negotiation. The purpose of the agreement is to define the roles of all parties during the interim and gearing up phase of the Institute and to lay groundwork for partnerships when the Institute is fully functional.

Several NPS agreements, although not specific to the Institute, are available for use by the Institute if needs arise. An Interagency Agreement is being signed among Department of the Interior agencies (BLM, FWS, USGS, NPS) and Department of Agriculture (USFS) to achieve more effective and efficient management of caves through cooperative action among those agencies. The agreement identifies areas of mutual concern and establishes avenues for cooperation in the management, research, protection and conservation of caves and karst.

The NPS Mexico Affairs Office (MEAF) has formal agreements with Mexican government departments and universities. The Institute can work through the MEAF to form partnerships and collaborative activities with Mexico related to cave and karst activities. Plans are being made to meet with Mexico representatives. Similar activities would be initiated with Canada through existing agreements.

*Funding:* The focus on non-Federal funding at this time is for building construction. Discussions are in progress with the City of Carlsbad and New Mexico Tech who intend to solicit state funds for a building. If enough funding can be obtained, the Institute would be integrated into a Carlsbad Science Center that includes and educational museum, an IMAX, and an auditorium. The source(s) of Federal and private funding for basic Institute operations and for research and educational activities have not yet been clearly identified.

*Physical facility:* Carlsbad has been working with a contractor to conceptually design a riverfront commercial complex, which includes conceptual options for the Institute and an educational museum facility. Additionally, a renovated building near the proposed riverfront development may be an option rather than a new building. Parallel options are being pursued to assure at least a temporary location by about mid-2003. A construction project, if decided upon, would likely not be completed until 2005.

*Research needs:* Ideas for research needs are being accumulated through informal and formal discussions with scientists and resource managers. As a list grows, groupings of research areas should emerge, which would form the basis for articulating national research needs. The February 2001 Karst Interest Group Workshop constitutes one of the opportunities to discuss research needs and add to the growing list, and focus groups could be convened at professional meetings as a special session, such as at AGU or GSA.

## REFERENCES

National Park Service, 1994, National Cave and Karst Research Institute Study Report to Congress, 37 p.

# Steps Toward Better Models of Transport in Karstic Aquifers

By David Loper
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## Abstract

One of the most difficult challenges of geophysical sciences is the accurate modeling of contaminant transport in karstic aquifers. This task requires expertise, knowledge and resources which transcend any single agency, company or university; a cooperative approach to this problem is needed. The Hydrogeology Consortium is a new scientific organization seeking to catalyze the development and use of better models of flow and transport in karstic aquifers; see hydrogeologyconsortium.org. A new conceptual model of karstic aquifers has been developed and will provide the basis for the development of a new mathematical and computational model.

## INTRODUCTION

It is well known that models of contaminant transport which employ the advection-diffusion equation perform very poorly in karstic settings. There is a strong need for better models of flow and transport, which simulate realistically the conduit system of a karstic aquifer, and its effect on flow and transport. In the following section the nature of the problem is described, then an important and necessary administrative step in addressing the problem is described in the subsequent section. A new conceptual model of karstic aquifers is described in the following section, then the process of sequestration and the importance of springs are briefly discussed.

## NATURE OF THE PROBLEM

Karstic aquifers, as well as other highly heterogeneous aquifers, are characterized by the presence of preferred pathways (e.g., conduits) in which flow speeds are very much larger than average. Contaminants within these pathways travel much faster, and consequently have much shorter travel times, than predicted by standard models based on Darcy's law and the advection-dispersion equation. This is a dangerous situation; for example, conventional modeling may predict a water supply to be safe from contamination, when in fact it is not. The result of an inaccurate prediction of contaminant behavior can be a costly clean-up.

The problem of accurately predicting contaminant transport is made more difficult by the fact that the preferred pathways are difficult to observe. The most common and reliable (but tedious and costly) method of locating large karstic conduits is by scuba diving. In virtually all field situations, the available site-characterization data will be of unsatisfactory density and quality.

This problem has not been vigorously addressed by the scientific community for several reasons: it is large in scope, difficult to analyze and lacking in adequate data. However, contaminant transport is an important problem that is likely to get worse, as population pressures increase. It needs to be addressed.

## THE HYDROGEOLOGY CONSORTIUM

About three years ago a group of individuals concerned about aquifer quality and protection formed the *Hydrogeology Consortium*, in order to provide a forum in which to address these concerns. The HC is an independent, not-for-profit organization which is affiliated with the Geophysical Fluid Dynamics Institute at Florida State University.

The vision of the Consortium is to have an abundant supply of clean water for human use while maintaining a healthy natural environment. To help achieve that vision, the Consortium has chosen as its mission *to cooperatively provide scientific knowledge applicable to ground water resource management and protection*. The primary goal is improved effectiveness and efficiency of management and protection of water resources, particularly ground water, through better understanding and application of scientific knowledge. The activities of the Consortium include, but are not be limited to :

- endeavoring to develop the necessary scientific knowledge and collect field data, when and where they are found to be lacking, in order to continuously improve our conceptual understanding of hydrogeological environments and associated models.
- fostering the cooperative development of valid scientifically based models of ground water: including its flow and the behavior of waterborne contaminants, as critical factors in determining the health of complex three-dimensional ecosystems.

In Eve L. Kuniansky, editor, 2001, U.S. Geological Survey Karst Interest Group Proceedings, Water-Resources Investigations Report 01-4011, p. 56-57

- coordinating the development and implementation of specific pilot studies and laboratory simulations designed to calibrate and test these models.

Although the activities of the Consortium will be focused primarily on issues of direct concern to the State of Florida, the scientific problems of ground water it will address are endemic to many other areas.

## A NEW CONCEPTUAL MODEL

The distinguishing feature of karstic aquifers is the set of connected conduits which form preferred pathways for flow and transport. This feature forms the basis for a new model, consisting of a classic porous matrix, permeated by a tree-like network of conduits. A conduit is defined as any interconnected pathway of sufficient size to have turbulent flow. With a typical head gradient of $10^{-4}$, turbulent flow occurs in conduits having cross-sectional area greater than 10 $cm^2$.

In the simplest form of the model, water is assumed to enter the smallest conduits, then proceed through a series of increasingly larger conduits, finally emanating at a spring. Conservation of water yields a unique relation among conduit length, conduit area and map area. The distance between adjacent conduits depends on their size (i.e., cross-sectional area) and the recharge rate. In Florida, the spacing between smallest conduits is typically 50 – 100 m.

Resistance to flow in conduits is much less than that in the adjacent matrix. Consequently nearly all regional flow is carried by conduits. By the same token, matrix flow is local, directed toward the nearest conduit. The local gradient associated with this conduit recharge may be quite different than the gradient associated with regional flow.

Dispersion of contaminants occurs by a variety of mechanisms in this model. First is dispersion in the matrix, associated with the local recharge. It is likely that this can be adequately quantified by the advection-dispersion equation. Second is dispersion within a single conduit due to turbulent flow. Rather surprisingly, turbulent dispersion is much weaker than Taylor dispersion, so that this mechanism may be ignored as a first approximation. Third is travel-time dispersion caused by the existence of multiple conduit pathways having a spectrum of lengths, areas and flow speeds. This latter mechanism is likely to be dominant in field and regional settings, is certain to be site-specific and is likely to be difficult to quantify.

## SEQUESTRATION

The new aquifer model, consisting of a conduit network imbedded in a porous matrix, has the ability to simulate an important physical process: *sequestration*. In the parlance of continuum mechanics, the conduits and matrix form a set of two interpenetrating continua, each of which can have its own distinct pressure. If the conduit system has a larger pressure than the matrix (caused for example by a rainstorm flooding a sinkhole), water will be actively forced from the conduits into the matrix. Any contaminant carried by that water will be sequestered in the matrix, to be released at some later time when the pressure differential is reversed (as in a time of drought). The process of sequestration can have a profound effect on the break-through curve of the aquifer, greatly extending the residence time of contaminants. It is important to understand and quantify sequestration, but virtually nothing is known of this process at present.

## THE IMPORTANCE OF SPRINGS

Springs are singular points of an aquifer system. Typically an aquifer is recharged primarily by rainfall distributed over the catchment area (though sinking streams provide a few point sources), and is discharged at a few singular locations, called springs. The properties of the discharge at a spring (head, flow rate, water temperature, water chemistry) are governed by, or strongly modified by, the physical nature of the aquifer through which the water has moved. It follows that there is a wealth of information about the aquifer and the catchment area contained in these properties.

In order to extract information of aquifer properties from spring-discharge data, several things are needed. First is a realistic model of the aquifer, incorporating the salient physical processes in a minimal set of parameters, which can be run as a forward model to simulate the properties of the spring discharge. Second is an inverse model, based on then forward model, which is capable of determining the aquifer parameters from a perfect set of discharge data (which does not exist). Third is adequate data obtained at certain springs. “Adequate” data should be of the proper type (e.g., head, flow rate, temperature, etc.), of sufficient duration (e.g., several years), of sufficient temporal density (e.g., daily) and of sufficient accuracy.

Springs are the integrators of change occurring in their catchment basin, and the consequence of change can be deduced from a lengthy and detailed data record. Given the need for a relatively long data record, it would be prudent to begin now to collect the data necessary to characterize an aquifer basin associated with any springs of interest or concern.

# Comparisons Among Ground-Water Flow Models and Analysis of Discrepancies in Simulated Transmissivities of the Upper Floridan Aquifer in Ground-Water Flow Model Overlap Areas

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## Abstract

Discrepancies in simulated transmissivities of the Upper Floridan aquifer were identified in the overlap areas of seven ground-water flow models in southwest and west-central Florida. Discrepancies in transmissivity are generally the result of uncertainty and spatial variability in other aquifer properties. All ground-water flow models were used to simulate the potentiometric surface of the Upper Floridan aquifer for approximated steady-state conditions from August 1993 through July 1994 using the time-independent hydraulic properties assigned to the models. Specified-head and general-head boundary data used to generate boundary conditions appropriate to these models were obtained from the estimated annual average heads for the steady-state period. Water-use data and the approximated surficial aquifer system water table were updated to reflect conditions during the approximated steady-state period. Simulated heads at control points, vertical leakage rates to the Upper Floridan aquifer, and spring flows were used to analyze the discrepancies in transmissivities in model overlap areas. Factors causing transmissivity discrepancies in model overlap areas include differences among directly applied recharge rates, differences among model simulated vertical leakance values assigned to the overlaying confining unit resulting in varying leakage rates to the Upper Floridan aquifer, differences in heads and conductances used in general-head boundary cells, and differences in transmissivities assigned in the vicinity of springs. Additional factors include the grid resolution and algorithm used to approximate the heads of the surficial aquifer system when these are used as a source/sink layer.

## INTRODUCTION

Seven ground-water flow models in southwest and south-central Florida were analyzed to identify discrepancies in the simulated transmissivity in model overlap areas. The seven ground-water flow models of the Upper Floridan aquifer (UFA) encompass southwest and west-central Florida (fig. 1); study area hydrology and model details are presented in Ryder (1985), Yobbi (1989, 1996), Barcelo and Basso (1993), Blandford and Birdie (1993), Hancock and Basso (1993), and Metz (1995). The transmissivities of the UFA used in the simulations range from about 8,000 feet squared per day ($ft^2/d$) in southwest Levy County to greater than 12,000,000 $ft^2/d$ in Citrus County. This large range in transmissivity is typical in karst areas. In addition to the areal variations in transmissivities of the UFA, there are large differences among transmissivities used to simulate overlap areas in the ground-water flow models. For the purpose of this study, a discrepancy between simulated transmissivities within model overlap areas was identified whenever transmissivity values differed by more than twice or less than half the average transmissivity. Transmissivity discrepancies are a source of conflict for water-management regulators when evaluating water-use permits because they can result in different simulated potentiometric levels in the UFA under identical future water-use stresses.

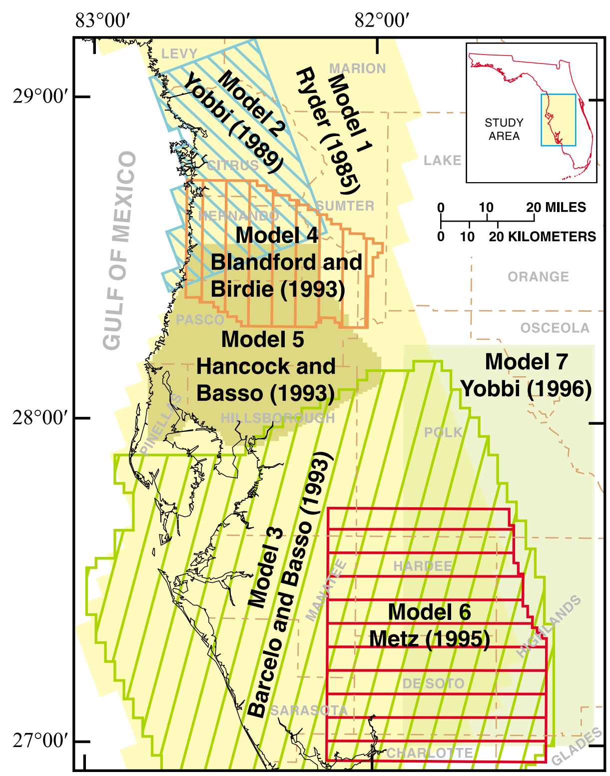


**Figure 1. Location and extent of the simulated areas in the Upper Floridan aquifer of ground-water flow models considered in the study.**

In Eve L. Kuniansky, editior, 2001, U.S. Geological Survey Karst Interest Group Proceedings, Water-Resources Investigations Report 01-4011 p. 58-67

## DESCRIPTION OF PREVIOUS MODELS

The Floridan aquifer system is a layered system and the conceptualization of the system varied among the seven models considered in this study (fig. 2). All models assume that the confining unit underlying the UFA is not leaky and, therefore, is simulated as a no-flow boundary (fig. 2). The surficial aquifer system (SAS) is simulated as a dynamic or active layer only by models 5 and 7. Models 1, 3, 6, and 7 simulate parts of the intermediate aquifer system in southwest Florida (fig. 2). Model 5 uses two layers to simulate the UFA. Models 2 and 4 are the only one-layer models, where recharge to the UFA is directly assigned to the layer; all other models compute the leakage to the UFA from the vertical leakance of the confining units and the hydraulic gradients between the UFA and either the intermediate aquifer system or the SAS. Model 1 used the computer code generated by Trescott and Larson (1976); all other models used MODFLOW (McDonald and Harbaugh, 1988). All models used block-centered grids where heads are computed at the center of the grid cells. A detailed discussion of these models is beyond the scope of this report.

The grids used by the ground-water flow models were variable in size and cells were not aligned along the same axis. Different grid scales generally caused some areas to be treated with higher resolution than others, increasing the spatial variability of hydraulic properties. The grid used to develop model 5 was of variable cell size, ranging from 0.25 square-mile cells to 1 square-mile cells (Hancock and Basso, 1993). All other models used grids of uniform, square cell size. The width of the cells of the uniform grids varied from 1 mile (model 7) to 4 miles (model 1). Grids of models 1 and 2 were rotated about 20 degrees west of due north. The variability in grid alignment required a scheme to identify the areas where discrepancies in transmissivity occur in overlap areas shown in figure 1.

## IDENTIFICATION OF DISCREPANCIES IN SIMULATED TRANSMISSIVITIES

The study area (fig. 1) was discretized into 5,000 foot-wide square cells to establish a framework grid for storing and analyzing the values of transmissivity assigned in the original models. The center points of the framework grid cells were intersected with the original model grid and the transmissivities assigned by the original model at the center points were stored at the corresponding framework grid cells. In addition, the center points of the original model grid cells were intersected with the framework grid and the transmissivities assigned by the original model at the center points were also stored at the corresponding framework grid cells. In cases where more than one transmissivity value was stored in a framework grid cell from any of the seven models, the geometric mean was computed from the multiple values obtained from that model. The transmissivities used in this study for model 5 were the sum of the transmissivities assigned to layers 2 and 3, because model 5 simulated the UFA with layers 2 and 3 (fig. 2). The transmissivities for all other models were taken as assigned.

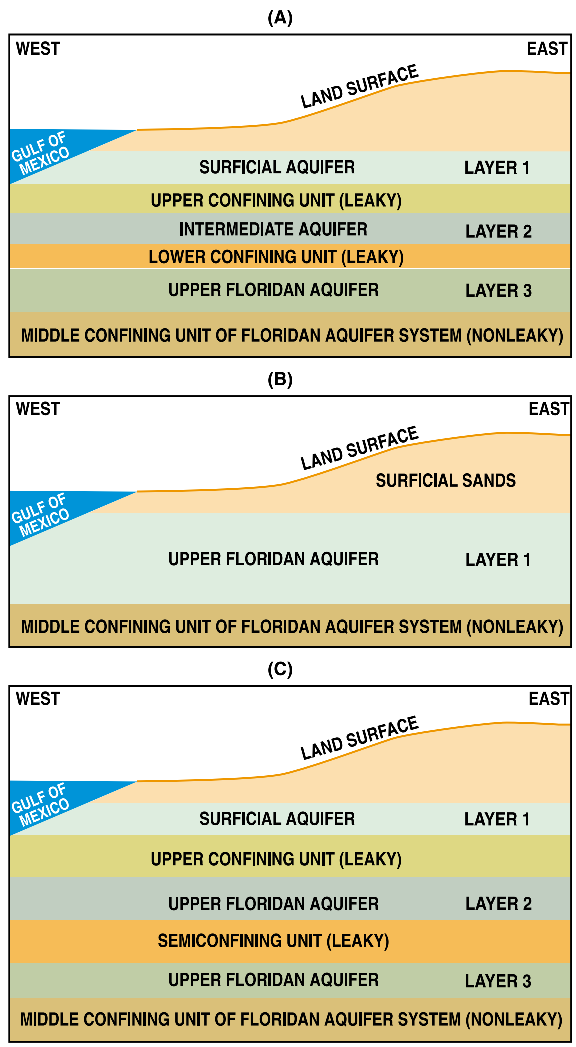


**Figure 2. Layering conceptualization of models (A) 1, 3, 6, and 7; (B) 2 and 4; and (C) 5. (See figure 1 for location of models.)**

A framework cell was identified as a transmissivity discrepancy if a transmissivity value from any model in an overlap area was either smaller than half or greater than twice the resulting geometric mean transmissivity. Cells with a discrepancy in transmissivity were grouped into 24 zones based on the geographical extent of the active areas of each model (fig. 3). The zones of transmissivity discrepancies were analyzed based on how well each model simulated the measured water levels and spring flows in the UFA and how realistic the assigned recharge rates or simulated leakage rates to the UFA seemed to be. All models were used to simulate average annual conditions in the UFA from August 1993 through July 1994.

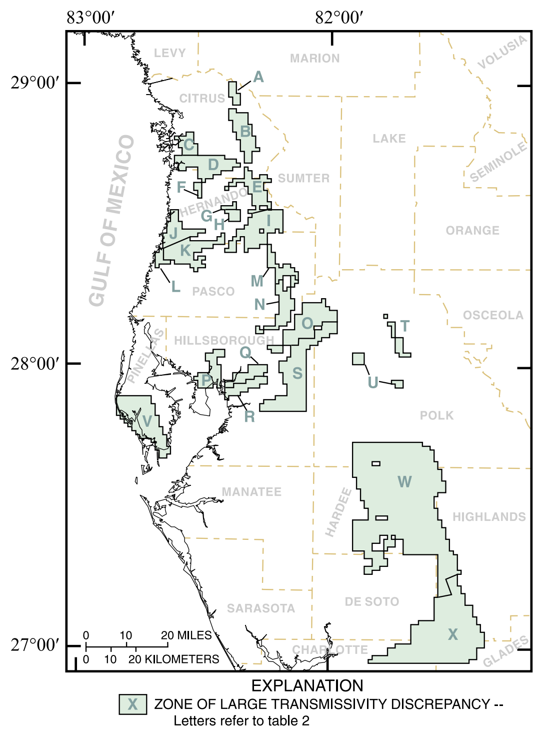


**Figure 3. Zones of discrepancies in simulated transmissivities of the Upper Floridan aquifer in ground-water flow model overlap areas.**

## MODEL INPUT AND CONCEPTUALIZATION

The seven models analyzed in this study were used to simulate the average potentiometric surface of the UFA from August 1993 through July 1994. The time-independent model input parameters including transmissivities, vertical leakances, spring and riverbed conductances, and conductances used for the general-head boundaries were compiled from the original published models. Time-dependent model input parameters, including specified heads in the SAS, water-use data, river stages, drain elevations, specified heads in the UFA, and heads used to specify the general-head boundaries, were updated to represent the prevailing hydrologic conditions of the simulation period. The combination of original, time-independent data and updated, time-dependent data was compiled for use with the computer code MODFLOW-96 (Harbaugh and McDonald, 1996) to perform the August 1993 through July 1994 steady-state simulation for each model. The simulated water levels and spring flows in the UFA were used to assess the sets of hydraulic parameters that better matched the measured ground-water levels and spring flows among the models. The simulated water levels used in this study for model 5 were the average of layers 2 and 3, because the hydraulic gradients between these two layers were negligible. The water levels for all other models were taken as simulated by the models.

### Water Table of the Surficial Aquifer System

The water table of the SAS was approximated using (1) the compiled data of lake elevations, river stages, and water-level measurements from surficial aquifer wells (fig. 4); (2) a digital elevation grid; (3) estimated lake elevations and river stages at ungaged lakes and rivers from the digital elevation grid; (4) an interpolated

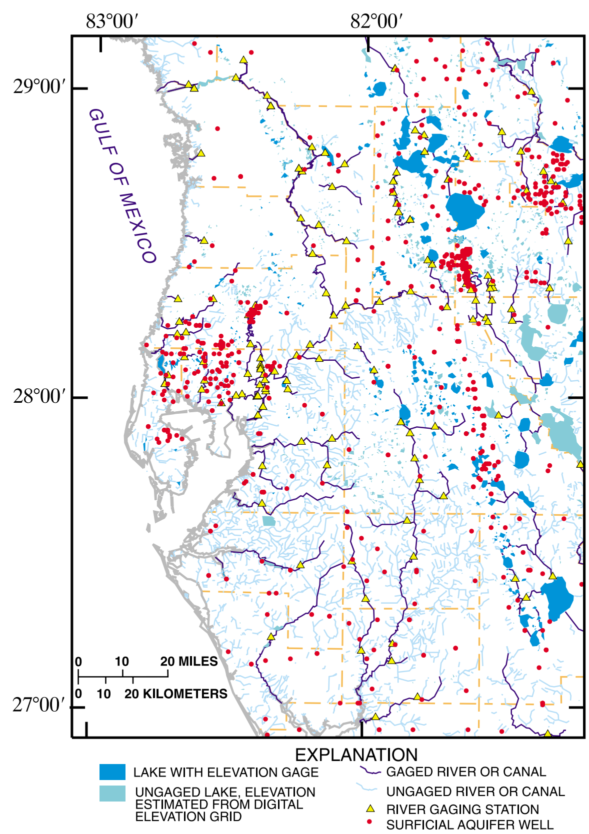


**Figure 4. Lakes, rivers, locations of stream gaging stations, and surficial-aquifer wells used to estimate the areal distribution of the surficial aquifer system water table.**

surface based on lake elevations and river stages, referred to as the “minimum water table;” and (5) a multilinear regression among the water-level measurements at surficial aquifer wells, the estimated minimum water table, and the land-surface elevation relative to the estimated minimum water table. Results of the multilinear regression were used to approximate the water table from the corresponding regression coefficients, the estimated minimum water-table elevation, and the land-surface elevation relative to the minimum water table. Data used to approximate the water table of the SAS was compiled from the U.S. Geological Survey (USGS), Southwest Florida Water Management District (SWFWMD), South Florida Water Management District (SFWMD), and St. Johns River Water Management District (SJRWMD) data bases.

From August 1993 through July 1994, average lake elevations were computed for gaged lakes in the study area, and average river stages were computed for gaged rivers (fig. 4). These rivers were divided into segments according to the location of the lakes and river gaging stations. River stage was computed at all discrete nodes located along the meanderings of the river segments using a linear approximation based on distance to upstream and downstream gages. Applicable distances were computed at all nodes forming each river segment. The computed lake elevations and river stages were assumed to be representative of the water-table elevation at the same sites, and elevations were referenced to the National Geodetic Vertical Datum of 1929.

The digital representation of the topography in the area was generated from 5-foot contour interval hypsography digitized by SWFWMD, SFWMD, and SJRWMD from 7.5-minute USGS topographic quadrangle maps. A digital elevation grid of square cells 100 feet wide was generated using the digitized hypsography for the study area, the lake elevations from gaged lakes, and the river stages computed along the meanderings of gaged rivers. Using this digital elevation grid, the land-surface elevation could be interpolated at any point in the study area.

Ungaged lake and river stages were computed from the digital elevation grid. Stages were interpolated along the discrete nodes forming the digital representation of the ungaged rivers. Although some of the ungaged lakes may not be representative of the regional water table (some of these lakes may be perched), excluding these lakes from the set of all ungaged lakes used to assess the areal distribution of the water table was beyond the scope of this study.

The minimum water table was generated by fitting quintic polynomials of continuous first and second derivatives between any two nodes of measured or estimated lake elevations, river stages, or ocean shoreline (which was assigned a water table of zero foot elevation). The minimum water table, water table, and land-surface elevation coincide at lakes and rivers (fig. 5). Elevations of the minimum water table at the surficial aquifer wells were interpolated from the minimum water-table surface previously generated. Land-surface elevations at the surficial aquifer wells were interpolated from the digital elevation grid, and the resulting elevations relative to the minimum water table were computed. A multilinear regression was computed between the measured water-table elevation as the dependent variable, and the minimum water table and the land-surface elevation relative to the minimum water table as the independent variables. A correlation coefficient of 0.99 shows that these variables are strongly correlated. The approximated water table computed from the multilinear regression was used to specify the heads of layer 1 in models 1, 3, and 6, in which the SAS was simulated as a constant-head layer. The root-mean-square error between the regressed and measured water table at surficial aquifer wells was 3.81 feet, whereas the absolute maximum of regressed minus measured heads was –8.12 feet.

Errors in the approximation of the source/sink heads of the SAS lead to errors in simulated hydraulic gradients, which in turn lead to errors in simulated leakage rates between the UFA and the SAS, or between the surficial and intermediate aquifer systems. The source/sink heads for the SAS in models 1, 3, and 6 were estimated from the land-surface elevation (Ryder, 1985; Barcelo and Basso, 1993; Metz, 1995). The algorithm described herein represents a uniform method for approximating the water table in the study area and generally agrees well with measured data.

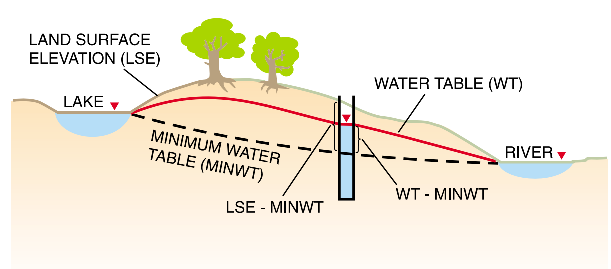


**Figure 5. Relation among water table, minimum water table, and land-surface elevation.**

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## Potentiometric Surface of the Upper Floridan Aquifer

Records of daily water levels from 1986 to 1996 from UFA wells equipped with continuous water-level recorders were evaluated to select a 1-year simulation period in which the error introduced by making a steady-state approximation was minimized. Only wells tapping the unconfined UFA were considered in this analysis, because the small storage coefficient typical of a confined aquifer generally make gain or loss of water from storage negligible. The smallest net changes in water levels in any 1-year period among the 22 unconfined UFA wells occurred from August 1993 through July 1994. Differences between water levels on July 31, 1994, and water levels on August 1, 1993, ranged from –1.07 to 1.23 feet, with a root-mean-square difference of 0.48 foot and a mean difference of –0.03 foot. If all 58 confined or unconfined UFA wells equipped with continuous water-level recorders were considered, then the differences ranged from –2.47 to 5.17 feet, with a root-mean-square difference of 1.89 feet and a mean difference of 0.89 foot. These differences indicate that the error introduced by making a steady-state approximation for this period is small.

Monthly averages for September 1993 and May 1994 and annual averages from August 1993 through July 1994 were computed from water levels in 58 UFA wells equipped with continuous water-level recorders. A multilinear regression was computed between the annual averages and the monthly averages for September 1993 and May 1994. A correlation coefficient of 0.99 for the multilinear regression indicated a strong correlation between the annual averages and the September 1993 and May 1994 averages. The multilinear regression equation used to compute the annual averages from August 1993 through July 1994 was: $h_{93\text{-}94} = 0.55\, h_{Sep93} + 0.45\, h_{May94} + 0.28$, where $h_{93\text{-}94}$ is the annual average for 1993-94 period, and $h_{Sep93}$ and $h_{May94}$ are the monthly averages for September 1993 and May 1994. This regression assumes that the water levels measured during September 1993 and May 1994 are representative of the monthly averages. The differences between the regressed and computed annual water-level averages at continuous water-level recorders tapping the UFA ranged from –0.67 to 3.50 feet, the root-mean-square difference was 0.74 foot, the mean difference was 0.15 foot.

A potentiometric-surface map of the UFA was generated to represent average hydrologic conditions from August 1993 through July 1994 in the SWFWMD and parts of SJRWMD and SFWMD (fig. 6). Approximately 90 percent of all water-level measurements were obtained from wells with surveyed land-surface elevations. The heads for the general-head boundaries for models 1-7 were interpolated from the potentiometric surface shown in figure 6. The conductances used to establish the general-head boundaries of these models were taken from the published models.

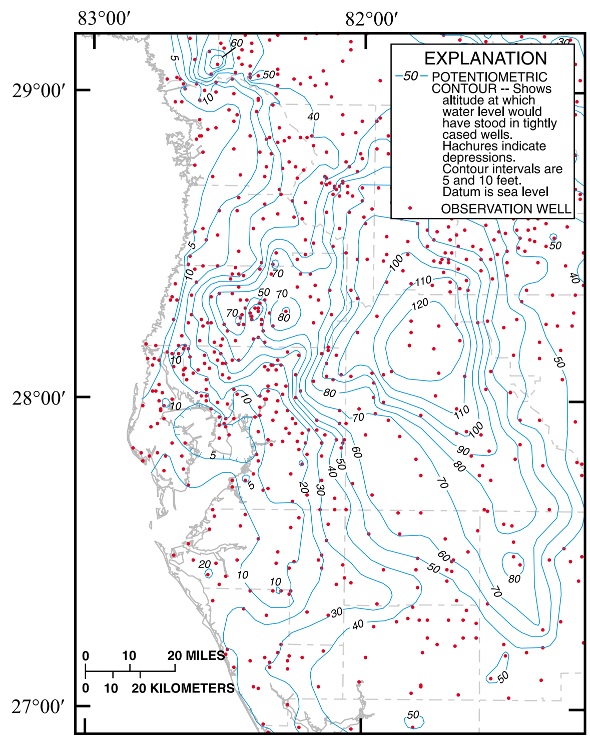


**Figure 6. Estimated potentiometric surface of the Upper Floridan aquifer, average conditions for August 1993 through July 1994.**

## Water Use

Ground-water withdrawals in the study area from August 1993 through July 1994 from the intermediate aquifer system and the UFA for public-water supply, commercial or industrial (including thermoelectric-power generation and recreational uses), and agricultural purposes were compiled or estimated (depending on the water-use type). Most of the ground-water withdrawals were compiled from consumptive user permit data bases and water-use data files from SWFWMD, SFWMD, and SJRWMD. Artificial recharge rates from injection wells were obtained from the Florida Department of Environmental Protection. Estimates of self-supplied domestic ground-water withdrawals were obtained from Richard L. Marella (USGS, written commun., 1998). Wells located inside the simulation areas of each model were used to generate the MODFLOW well-package input file needed to run each model. Approximate total ground-water withdrawals from the simulation areas of models 1 through 7, including self-supplied domestic water use and recharge from injection wells, were 895, 65, 594, 125, 280, 231, and 400 million gallons per day, respectively.

## Spring Flow

Spring flows within the study area originate mostly from the UFA. A major factor in spring flow is the net aquifer recharge from rainfall; however, spring response is delayed by aquifer-matrix storage. Higher spring flows usually occur in late fall after the rainy season, whereas lower discharges occur in late spring when rainfall has been low. Spring flows from the UFA tend to create depressions in the potentiometric surface. The areal extent of these depressions depends on the magnitude of the spring flow, and the aquifer and confining-unit properties in the vicinity of the spring.

Location and spring-flow data for springs originating from the UFA and located inside the zones of transmissivity discrepancies (fig. 3) were compiled in table 1 from several sources (Rosenau and others, 1977; Yobbi, 1989, 1992). Most of the flow measurements of springs in the zones of transmissivity discrepancies was estimated from previous flow measurements (Rosenau and others, 1977; Yobbi, 1989, 1992). A few spring-flow measurements in the zones of transmissivity discrepancies were made from August 1993 to July 1994 (table 1).

Average flows from May 1988 to April 1989 for a number of springs in parts of Citrus, Hernando, and Pasco Counties were calculated from Yobbi (1992). The average of flow measurements at Weeki Wachee Springs was 185 cubic feet per second ($ft^3/s$) from May 1988 to April 1989 (Yobbi, 1992); and 129 $ft^3/s$ from August 1993 to July 1994 (USGS, 1993, 1994), or about 70 percent the average flow from May 1988 to April 1989. Due to the lack of additional spring-flow measurements from May 1988 to April 1989 and from August 1993 to July 1994, the average flow from August 1993 to July 1994 for springs in parts of Citrus, Hernando, and Pasco Counties was estimated to be 70 percent of the average of flow measurements in Yobbi (1992) from May 1988 to April 1989 (table 1). Average spring flows from August 1993 to July 1994 for springs in Yobbi (1989), not listed in Yobbi (1992), were also estimated to be 70 percent of the average spring flows listed in Yobbi (1989).

Average flows for springs in table 1 but not in Yobbi (1992) or Yobbi (1989) were estimated from the product of the flow measurement from Rosenau and others (1977) and the ratio of the August 1993 to July 1994 rainfall to the year of flow-measurement rainfall. During the study period (1993-94), total rainfall recorded at National Oceanic and Atmospheric Administration stations in Citrus, Hernando, and Pasco Counties was about 85 percent of the 1961-90 average and about 75 percent of the rainfall from May 1988 to April 1989. Although the spring flow does not follow an exact rainfall ratio, average spring flows for springs not listed in Yobbi (1992) or Yobbi (1989) were estimated assuming an approximate rainfall to spring-flow ratio.

# COMPARISON OF MODEL RESULTS AND ANALYSIS OF DISCREPANCIES IN SIMULATED TRANSMISSIVITIES

The nonuniqueness of the solution to the ground-water flow equation and the uncertainty and spatial variability in hydraulic parameters generally yields discrepancies in parameter values among model overlap areas. Time-dependent parameters like specified heads, recharge rates to the unconfined UFA, and ground-water withdrawals from the UFA needed to be updated for the simulated time period. The transmissivity of the UFA, as well as the vertical leakance of the intermediate confining unit, generally are time-independent hydraulic parameters and, therefore, do not need to be updated. However, the uncertainty of hydraulic parameters can be reflected in the time-independent parameters, causing discrepancies among ground-water flow models.

The simulation of average conditions of the potentiometric surface of the UFA and average spring flows from August 1993 through July 1994 was accomplished for each model using the updated water-table distribution, UFA specified-head boundaries, ground-water withdrawals discussed earlier, and the hydraulic properties assigned to each original model. The residuals between simulated and measured water levels in the UFA were computed for zones of transmissivity discrepancies A through X (fig. 3). The residuals and a comparison of simulated and estimated spring flows for each model in zones A through X were used to analyze the reliability of the transmissivity and leakage rates of each model (table 2).

Measured heads and reliable UFA spring-flow measurements were used to determine which transmissivity values and leakage rates to the UFA are realistic among the considered models. Springs were located in zones C, D, J, L, N, P, and S (table 1). Zone D is an example of an area where both head measurements and spring flows are needed to assess the reliability of the hydraulic parameters used by models 1, 2, and 4. Zones F, G, and H suggest the need for

Table 1. Description and flow measurements of Upper Floridan aquifer springs in zones of transmissivity discrepancies

[Zone labels indicate zone where springs are located, see fig. 3 for zone locations; if more than one date of measurement is listed, flow is an average of measurements; $ft^3/s$, cubic feet per second; dates are shown in month-year format]

| Spring name | Zone | Latitude | Longitude | County | Flow, in $ft^3/s$ | Date(s) of flow measure-ment(s) |
|---|---|---|---|---|---|---|
| Halls River Springs | C | 284804 | 823610 | Citrus | [a]102.2 | |
| Hidden River Springs near Homosassa (including Hidden River Head Spring) | C | 284559 | 823520 | Citrus | [b]6.7 | |
| Homosassa Springs at Homosassa Springs | C | 284758 | 823520 | Citrus | [b]72.4 | |
| Southeast Fork Homosassa Springs at Homosassa Spring | C | 284751 | 823523 | Citrus | [a]43.1 | |
| Trotter Spring at Homosassa Springs | C | 284747 | 823510 | Citrus | [b]5.2 | |
| Baird Creek Head Spring near Chassahowitzka | D | 284230 | 823440 | Citrus | [b]3.2 | |
| Beteejay Lower Spring near Chassahowitzka (including Beteejay Head Spring) | D | 284131 | 823535 | Citrus | [b]7.3 | |
| Chassahowitzka Springs near Chassahowitzka | D | 284254 | 823435 | Citrus | [b]64.8 | |
| Crab Creek Spring | D | 284300 | 823434 | Citrus | [b]34.8 | |
| Lettuce Creek Spring | D | 284308 | 823437 | Citrus | [b]3.7 | |
| Potter Spring near Chassahowitzka (including Ruth Spring) | D | 284354 | 823548 | Citrus | [b]14.4 | |
| Rita Maria Spring near Chassahowitzka | D | 284126 | 823528 | Hernando | [b]3.3 | |
| Salt Creek Head Spring | D | 284323 | 823506 | Citrus | [b]0.4 | |
| Unnamed Tributary above Chassahowitzka Springs (including Bubba Spring) | D | 284254 | 823438 | Citrus | [b]20.5 | |
| Boat Spring at Aripeka | J | 282621 | 823929 | Hernando | [b].4 | |
| Bobhill Springs | J | 282607 | 823834 | Hernando | [b]1.8 | |
| Jenkins Creek Spring No. 5 | J | 283120 | 823804 | Hernando | [b]15.3 | |
| Magnolia Springs at Aripeka | J | 282558 | 823926 | Pasco | [b].5 | |
| Mud Spring near Bayport | J | 283240 | 823701 | Hernando | [b]17.0 | |
| Salt Spring near Bayport | J | 283246 | 823709 | Hernando | [b]22.3 | |
| Unnamed Spring No. 1 | J | 282600 | 823926 | Hernando | [a]6.3 | |
| Unnamed Spring No. 2 | J | 282720 | 823830 | Hernando | [a].7 | |
| Unnamed Spring No. 4 | J | 283118 | 823806 | Hernando | [a]6.3 | |
| Unnamed Spring No. 6 | J | 283254 | 823737 | Hernando | [a]2.8 | |
| Horseshoe Spring near Hudson | L | 282350 | 824121 | Pasco | [c]9.7 | 12-72 |
| Unnamed Spring No. 3 near Aripeka | L | 282352 | 824027 | Pasco | [c]17.8 | 08-60 |
| Crystal Springs near Zephyrhills | N | 281030 | 821120 | Pasco | 37.0 | 09-93, 05-94 |
| Sulphur Springs at Sulphur Springs | P | 280115 | 822705 | Hillsborough | 25.0 | 09-93, 05-94 |
| Lithia Springs Major near Lithia | S | 275158 | 821352 | Hillsborough | 31.1 | 09-93, 05-94 |
| Lithia Springs Minor near Lithia | S | 275201 | 821349 | Hillsborough | 8.0 | 09-93, 05-94 |

[a]Estimated to be 70 percent of average of flow measurements from Yobbi (1989).
[b]Estimated to be 70 percent of average of flow measurements from Yobbi (1992).
[c]Estimated from product of measured flow from Rosenau and others (1977) and ratio of August 1993 – July 1994 rainfall to year of flow-measurement rainfall recorded at nearest station.

reliable evapotranspiration data to better estimate leakage rates to the UFA. In zones where only head measurements were available (A, B, I, K, M, Q, R, T, U, and W), only generalizations can be made about the validity of transmissivity and leakage rates. In many of these zones, the mean residual can only indicate the direction in which the transmissivity and leakage rates should be changed once one parameter is kept constant, but not the direction in which both parameters should be changed.

The nonuniqueness of the solution to the ground-water flow equation implies, for example, that different combinations of transmissivities and vertical leakances can result in similar simulated heads. Matching

Table 2. Description of zones with transmissivity discrepancies in model overlap areas

[Tn, average transmissivity of the Upper Floridan aquifer (UFA) assigned by model n, in thousand feet squared per day; Ln, average leakage or recharge rate to the UFA, as the case may apply, assigned by model n, in inches per year; Rn, root-mean-square error of residuals from control points of UFA for model n, in feet, mean of residuals, in feet, and number of control points; Qn, sum of simulated spring flows by model n, in cubic feet per second ($ft^3/s$); Q, sum of measured or estimated spring flows, in $ft^3/s$. Several values of Tn are shown if T is areally variable. See figure 3 for zone labels; see figure 1 for model numbers]

| Zone | Models with active cells | Simulated transmissivity | Simulated leakage or recharge | Statistics of residuals and spring flow | Explanation or comment |
|---|---|---|---|---|---|
| A, B | 1, 2 | T1= 2,000, 500<br>T2= 450, 155 | L1= 11.0<br>L2= 16.4 | R1= 7.54, -4.87, 7<br>R2= 4.57, 3.06, 7 | T should be between T1 and T2 and L rate should be between L1 and L2. |
| C | 1, 2 | T1= 1,000<br>T2= 9,000 | L1= 10.0<br>L2= 9.0 | R1= 0.20, -0.20, 1<br>R2= 1.17, 1.17, 1<br>Q1=77, Q2=375<br>Q=230 | Similar L rates and Q1 lower than Q suggest T should be higher than T1. Higher Q2 than Q suggests T should be lower than T2. |
| D | 1, 2, 4 | T1= 1,000<br>T2= 6,500<br>T4= 900 | L1= 12.6<br>L2= 17.7<br>L4= 14.7 | R1= 3.80, 2.34, 2<br>R2= 3.78, 2.93, 2<br>R4= 0.94, 0.04, 2<br>Q1= 189, Q2=199<br>Q4= 28, Q=152 | Low R4 and low Q4 show the solution of the ground-water flow equation is not unique and calibration should make use of spring flows in addition to heads. T and L should vary areally, with T increasing towards the springs. |
| E | 1, 2, 4 | T1= 1,000<br>T2= 1,500<br>T4= 475 | L1= 11.6<br>L2= 13.5<br>L4= 11.1 | R1= 6.73, 6.26, 3<br>R2= 0.80, 0.61, 3<br>R4= 3.30, 2.24, 3 | Outflux through lateral boundaries of model 4 is four times higher than flux through same cells in models 1 and 2. T should be higher than T4. |
| F | 1, 2, 4 | T1= 250<br>T2= 250<br>T4= 1,300 | L1= 14.0<br>L2= 21.0<br>L4= 39.9 | R1= 1.63, 1.63, 1<br>R2= 3.03, 3.03, 1<br>R4= 0.25, 0.25, 1 | L rate should be between L1 and L2. Low R4 could also be achieved with T lower than T4 and L lower than L4. L4 seems too high. |
| G | 1, 2, 4 | T1= 1,000<br>T2= 1,500<br>T4= 160 | L1= 18.0<br>L2= 27.0<br>L4= 39.5 | R1= 9.14, -9.14, 1<br>R2= 1.40, 1.40, 1<br>R4= 6.79, 6.79, 1 | T should be between T1 and T2. L rate should be between L1 and L2. L4 seems too high. |
| H | 1, 2, 4, 5 | T1= 1,000<br>T2= 1,500<br>T4= 190<br>T5= 400 | L1= 16.5<br>L2= 16.5<br>L4= 37.6<br>L5= 5.8 | R1= 9.93, -9.93, 1<br>R2= 0.00, 0.00, 1<br>R4= 5.08, 5.08, 1<br>R5= 0.56, -0.56, 1 | Contrasting T and L values can produce similar water levels. Additional data, such as evapotranspiration estimates, are needed to determine L rates that are physically possible. |
| I | 1, 4, 5 | T1= 500, 130<br>T4= 575, 300, 18<br>T5= 150, 80, 35 | L1= 3.8<br>L4= 11.7<br>L5= 18.3 | R1=32.71,-28.63,7<br>R4= 8.04, -1.26, 7<br>R5= 7.75, 3.77, 7 | T could vary areally between T4 and T5 and L rate should be between L4 and L5. Mean residual for model 1 suggests L rate should be higher than L1. |
| J | 1, 2, 4, 5 | T1= 2,000, 1,000, 500, 250<br>T2= 2,000, 1,000, 500, 43<br>T4= 2,000, 1,000, 675, 420<br>T5=575,400,120 | L1= 10.3<br>L2= 1.8<br>L4= 5.7<br>L5= -2.9 | R1= 1.38, -0.72, 3<br>R2= 3.05, -1.63, 3<br>R4= 3.26, -2.40, 3<br>R5= 4.31, -3.75, 3<br>Q1= 82, Q2= 67<br>Q4= 91, Q5= 0<br>Q= 73 | Areal distribution of T is highly variable and contrasts from one model to another. No simulated spring flow by model 5 suggests T should be higher than T5 and L rate should be higher than L5. L rates vary areally because both recharge and discharge areas are included in zone. |
| K | 1, 4, 5 | T1= 500, 130<br>T4= 1,800, 790, 110, 26<br>T5= 115, 80, 55 | L1 = 17.0<br>L4 = 32.3<br>L5 = 9.4 | R1=24.08,-14.11,6<br>R4= 5.16, 3.48, 6<br>R5= 8.04, 0.79, 6 | T should be higher than T1 in areas where T1=130. Mean residuals for models 4 and 5 suggest L rate should be lower than L4 and T should be higher than T5. |
| L | 1, 5 | T1= 250<br>T5= 80 | L1= 0.0<br>L5= -4.8 | R1= 1.79, 1.79, 1<br>R5= 0.40, 0.40, 1<br>Q1= 13, Q5 = 8<br>Q= 27 | If T5 is increased near springs then Q5 would increase. Zone is mainly a discharge area, which suggests L could be lower than L1. |
| M | 1, 4, 5 | T1= 500, 40<br>T4= 150, 100, 20<br>T5= 400, 150, 55 | L1= -1.5<br>L4= 4.6<br>L5= 14.6 | R1=29.86,-29.49, 4<br>R4= 0.81, 0.44, 4<br>R5= 7.10, 7.01, 4 | L rate should be higher than L1. Mean residual for model 5 suggests L rate should be lower than L5. |

Table 2. Description of zones with transmissivity discrepancies in model overlap areas--Continued

[Tn, average transmissivity of the Upper Floridan aquifer (UFA) assigned by model n, in thousand feet squared per day; Ln, average leakage or recharge rate to the UFA, as the case may apply, assigned by model n, in inches per year; Rn, root-mean-square error of residuals from control points of UFA for model n, in feet, mean of residuals, in feet, and number of control points; Qn, sum of simulated spring flows by model n, in cubic feet per second ($ft^3/s$); Q, sum of measured or estimated spring flows, in $ft^3/s$. Several values of Tn are shown if T is areally variable. See figure 3 for zone labels; see figure 1 for model numbers]

| Zone | Models with active cells | Simulated transmissivity | Simulated leakage or recharge | Statistics of residuals and spring flow | Explanation or comment |
|---|---|---|---|---|---|
| N | 1, 5 | T1= 100<br>T5= 400 | L1= 1.8<br>L5= -2.0 | R1=21.03,-20.83, 2<br>R5= 5.49, 5.40, 2<br>Q1=0, Q5=36, Q=37 | No simulated spring flow simulated by model 1 suggests T should be higher than T1. L rates need to vary areally. L rates in discharge area could be less than L5. |
| O | 1, 5 | T1= 100<br>T5= 30 | L1= -0.9<br>L5= 10.0 | R1=40.50,-40.33, 7<br>R5=14.60,-17.39, 7 | L rate should be higher than L1. Conductances in model 5 on general-head boundary in northeast are too low. T should be higher than T5. |
| P | 1, 5 | T1= 200<br>T5= 50 | L1= 0.6<br>L5= 15.4 | R1=10.88, -8.48, 6<br>R5= 6.01, 5.64, 6<br>Q1= 0, Q5= 27<br>Q= 25 | T should be lower than T1 in a discharge area. L rate should be lower than L5 and higher than L1. |
| Q | 1, 3, 5 | T1= 200<br>T3= 33<br>T5= 400, 150 | L1= 0.3<br>L3= 0.7<br>L5= -29.4 | R1= 6.91, -5.18, 6<br>R3= 4.38, 2.48, 6<br>R5= 2.99, -2.07, 6 | L rate should be higher than L1. T should be near T3 in discharge area. |
| R | 1, 3 | T1= 200<br>T3= 33 | L1= 1.1<br>L3= 1.7 | R1= 3.47, -0.87, 3<br>R3= 8.03, 4.70, 3 | T should be higher than T3 and lower than T1 if L rates remain between L1 and L3. |
| S | 1, 3, 5 | T1= 130<br>T3= 130, 33<br>T5= 40 | L1= 5.8<br>L3= 10.0<br>L5= 5.9 | R1=22.33,-10.83,13<br>R3= 5.47, 3.53, 13<br>R5=10.01, -9.44, 3<br>Q1=13, Q3=18<br>Q5=0, Q=39 | Areally variable T as in model 3 should be used. Mean residual for model 3 suggests L rate should be lower than L3 or if L3 is used, then T should be higher than T3. Only a subset of zone is simulated in model 5. |
| T | 3, 7 | T3= 67<br>T7= 12 | L3= 5.6<br>L7= 6.2 | R3= 0.93, 0.93, 1<br>R7= 2.73, 2.73, 1 | L rates could be lower than L3 if T3 is used. T could be higher than T7 if L7 is used. |
| U | 1, 3, 7 | T1= 130<br>T3= 130<br>T7= 40 | L1= 6.3<br>L3= 4.1<br>L7= 6.6 | R1= 10.90, -7.71, 2<br>R3= 4.14, -2.89, 2<br>R7= 3.00, -2.95, 2 | These are small areas of recharge to the UFA. T should be closer to T7 and L rate should be closer to L7. |
| V | 1, 3 | T1= 30, 17<br>T3= 200 | L1= -1.3<br>L3= 1.3 | R1= 5.29, 4.20, 8<br>R3= 2.21, 2.40, 8 | Higher T requires higher L rate, lower T requires lower L rate. Aquifer test result suggests T could be even higher than T3. |
| W | 1, 3, 6, 7 | T1= 400, 130<br>T3= 134<br>T6= 400, 100<br>T7= 400, 130, 66 | L1= 2.4<br>L3= 3.2<br>L6= 1.4<br>L7= 2.1 | R1= 6.86, 4.87, 19<br>R3= 6.09, 3.31, 19<br>R6= 5.58, -1.87, 19<br>R7= 3.04, -0.19, 19 | If L rates remain between L6 and L3, then T should vary areally. If T is uniform, then L rates should vary areally. |
| X | 3, 6, 7 | T3= 400<br>T6= 100<br>T7= 400 | L3= 0.4<br>L6= -0.6<br>L7= 0.2 | R3= 4.00, 0.65, 6<br>R6= 4.26, -2.93, 5<br>R7= 1.86, -1.05, 5 | Fluxes through lateral boundaries are small for all models. Low T6 and low L6 yield a high R6 value. Model 3 extends further south than models 6 or 7. |

measured and simulated heads can indicate whether the simulated transmissivity should be increased or decreased if the vertical leakance values are not changed (or vice versa), but heads alone cannot indicate how both parameters should be changed. Zones I, K, Q, and W are examples of uncertainties where either the transmissivity or the vertical leakance (reflected by the leakage rates) could vary areally while the other parameter remains unchanged (table 2). The availability of either known flux rates or reliable aquifer performance tests could answer which hydraulic parameters are more representative of the aquifer properties.

Fluxes through general-head boundaries in zones E and O can be compared among models. In these zones, the reliability of the data used to establish lateral boundaries in some models can be analyzed by comparing fluxes simulated by models. Specified heads used for some general-head boundaries in models 4 and 5 suggest that unrealistic fluxes are simulated to be entering (model 5, zone O) or leaving (model 4, zone E) the model areas when compared to fluxes simulated by models 1 or 2. Errors in the interpolation scheme used to estimate the specified heads at the general-head boundaries may have translated into errors in the conductances specified at the general-head boundary cells in models 4 and 5.

## SUMMARY AND CONCLUSIONS

Seven ground-water flow models in southwest and south-central Florida were analyzed to identify discrepancies in the simulated transmissivity in model overlap areas. Average conditions from August 1993 through July 1994 in the UFA were simulated for each model in their respective areas using updated water-table elevations, UFA specified-head boundaries, water-use data for the period, and the hydraulic properties used by the original models. The simulated and measured heads and spring flows were compared to identify and analyze some of the reasons for the transmissivity discrepancies.

In general, the factors causing transmissivity discrepancies in model overlap areas include differences among directly applied recharge rates, differences among model simulated vertical leakance values assigned to the overlaying confining unit resulting in varying leakage rates to the UFA, differences in heads and conductances used in general-head boundary cells, and differences in transmissivities assigned in the vicinity of springs. Additional factors causing transmissivity discrepancies are the grid resolution and the algorithm used to approximate the heads of the surficial aquifer when these are used as a source/sink layer. This study underlines the need for reliable data to improve the quantification of some hydraulic parameters, particularly the recharge and leakage rates to the Upper Floridan aquifer.

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# Hydrogeology and Ground-Water Flow Simulation of a Karst Ground-Water Basin in the Valley and Ridge Physiographic Province near Hixson, Tennessee

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Cave Springs is one of the larger springs in Tennessee, one of the most heavily stressed, and an important source of water for Hixson, Tennessee. Cave Springs derives its flow from a carbonate rock (karst) aquifer in the Valley and Ridge Physiographic Province. The aquifer framework in the Cave Springs area consists of dense Paleozoic carbonate rocks with secondary permeability mantled by thick residual clay-rich regolith in most of the area and coarse alluvium in the valley of North Chickamauga Creek. Transmissivities estimated from well hydraulic tests conducted across the Cave Springs area span a wide range, varying from 240 to 900,000 foot squared per day ($ft^2$/d) with a medium value of 5,200 $ft^2$/d. Recharge to the aquifer occurs both from direct infiltration of precipitation and from losing streams. Most recharge occurs during the winter and spring months.

Current ground-water withdrawals from the Cave Springs area by the Hixson Utility District average about 8 million gallons per day (Mgal/d) from two well fields, Cave Springs (6 Mgal/d) and Walkers Corner (1.7 Mgal/d). Present and planned future withdrawals may be approaching the capacity of the ground-water system to supply the utility district needs. The U.S.Geological Survey, in cooperation with the Hixson Utility District, conducted a study integrating previously collected information about the Cave Springs area ground-water system to evaluate both the annual water budget and the effects of current and planned increases in ground-water withdrawals.

To address these questions, a numerical ground-water flow model of the ground-water system was constructed and calibrated using MODFLOW 2000. Preliminary results of the modeling effort indicate that losing streams along the base of Cumberland Plateau escarpment at the western edge of the study area are an important source of recharge to the ground-water system, supplying about 50 percent of the recharge to the study area. The other source of recharge, direct infiltration of precipitation, accounts for the remaining recharge to the study area. Current ground-water withdrawals (7.7 Mgal/d) equal about 14 percent of the total ground-water recharge with the remaining 86 percent discharging to rivers (46 percent), springs (18 percent), and Chickamauga Lake (22 percent). Current drawdown at the Walkers Corner well field is about 33 feet at the center of a cone of depression that is elongated along strike. If additional pumping at Walkers Corner increases total withdrawals to 9.7 Mgal/d, simulated drawdown at Walkers Corner well field increases to about 60 feet. The additional ground-water withdrawals result in a 5-percent decrease in discharge to Chickamauga Lake, a 4-percent decrease in discharge to rivers, and a 2-percent decrease in discharge to springs.

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# Travel Times Along Selected Flow Paths of the Edwards Aquifer, Central Texas

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## Abstract

Flow path travel times in the structurally controlled, karstic Edwards aquifer were estimated using simulated ground-water levels obtained from a finite-element model. For this analysis, simulated monthly ground-water levels were averaged over an 11-year calibration period to minimize the transient effect of short-term recharge and discharge events. The 1978-89 calibration period was characterized by average to wetter-than-average climatic conditions; simulated water-level and spring-flow compared favorably with measured data. Flow paths for which travel times were estimated range from 1,250 to 10,000 feet wide and from about 8 to 180 miles long. Effective aquifer thickness and effective porosity can be highly variable and is poorly defined throughout most of the aquifer. Accordingly, travel-time estimates were computed within known or inferred thicknesses and porosities within known or inferred ranges of 350 to 850 feet and 15 to 35 percent, respectively. The minimum rock matrix porosity for each element was divided by 10 to estimate a minimum time of travel (a worst case time of travel). Travel times range from 14 to 160 years for a flow path from the Blanco River Basin to San Marcos Springs and from 350 to 4,300 years for a flow path from the West Nueces River Basin to Comal Springs. Travel times near the minimum of the ranges are similar in magnitude to those determined from tritium isotopes in spring water, thus supporting the hypothesis that effective porosity and effective thickness of the aquifer is less than the respective ranges.

## INTRODUCTION

The structurally controlled, karstic Edwards aquifer is the sole source water supply for San Antonio, Texas. Water enters the Edwards aquifer from precipitation over its outcrop area and streamflow from the catchment area of the Hill Country. The gaining streams incised into the Trinity aquifer in the Hill Country cross the outcrop of the highly permeable and fractured rocks of the Edwards aquifer in the Balcones fault zone and disappear underground (fig. 1, Kuniansky, 1989). The major natural discharge from the Edwards aquifer is at springs. Two of the major springs, Comal and San Marcos are habitat for endangered species. This karst system is unique due to its existence in a semi-arid area and the geologic structure that controls the direction of ground-water movement in the aquifer. A finite-element model of the Edwards and Trinity aquifers within the Hill Country and Balcones fault zone in central Texas (fig. 1) (Kuniansky, 1994, 1995; Kuniansky and Ardis,[in press]) was designed to incorporate the geologic and hydrologic conditions affecting ground-water flow and to better understand the flow system. Faulting throughout the study area, and particularly in the Balcones fault zone, results in horizontal anisotropy that strongly influences regional ground-water flow patterns. The finite-element method is one of the few numerical methods that can represent hydraulic characteristics that vary in the horizontal direction and was well suited for developing a heterogeneous continuum model of this karst system. A detailed deterministic numerical model synthesizes known information including geologic structure, recharge and discharge, and ground-water level by solving the ground-water flow equations for water levels given boundary conditions, parameters (hydraulic properties), and stresses (pumping and recharge).

The purpose of this extended abstract is: to describe the geologic structure that affects ground-water flow direction within the Edwards aquifer; to describe how flow paths were determined from the average simulated potentiometric surface (1978-89); to show flow paths from points where water enters the aquifer at streams to major natural discharge features (Comal or San Marcos Springs); and to provide worse-case (fastest) estimated times of travel along these flow paths. The model design, layering, and boundary conditions are published in Kuniansky, 1994 and 1995, and are not described herein.

## Description of the Study Area

The Hill Country is characterized by rough rolling terrain dissected by the headwaters of the streams within the Nueces and Guadalupe River Basins. These streams have been eroding headward into the Edwards Plateau forming narrow valleys with steep carbonate walls. Wider stream valleys along the major streams may have formed by lateral cutting and karstic processes during periods of greater rainfall (Wermund and others, 1974).

In Eve L. Kuniansky, editor, 2001, U.S. Geological Survey Karst Interest Group Proceedings, Water-Resources Investigations Report 01-4011, p. 69-77

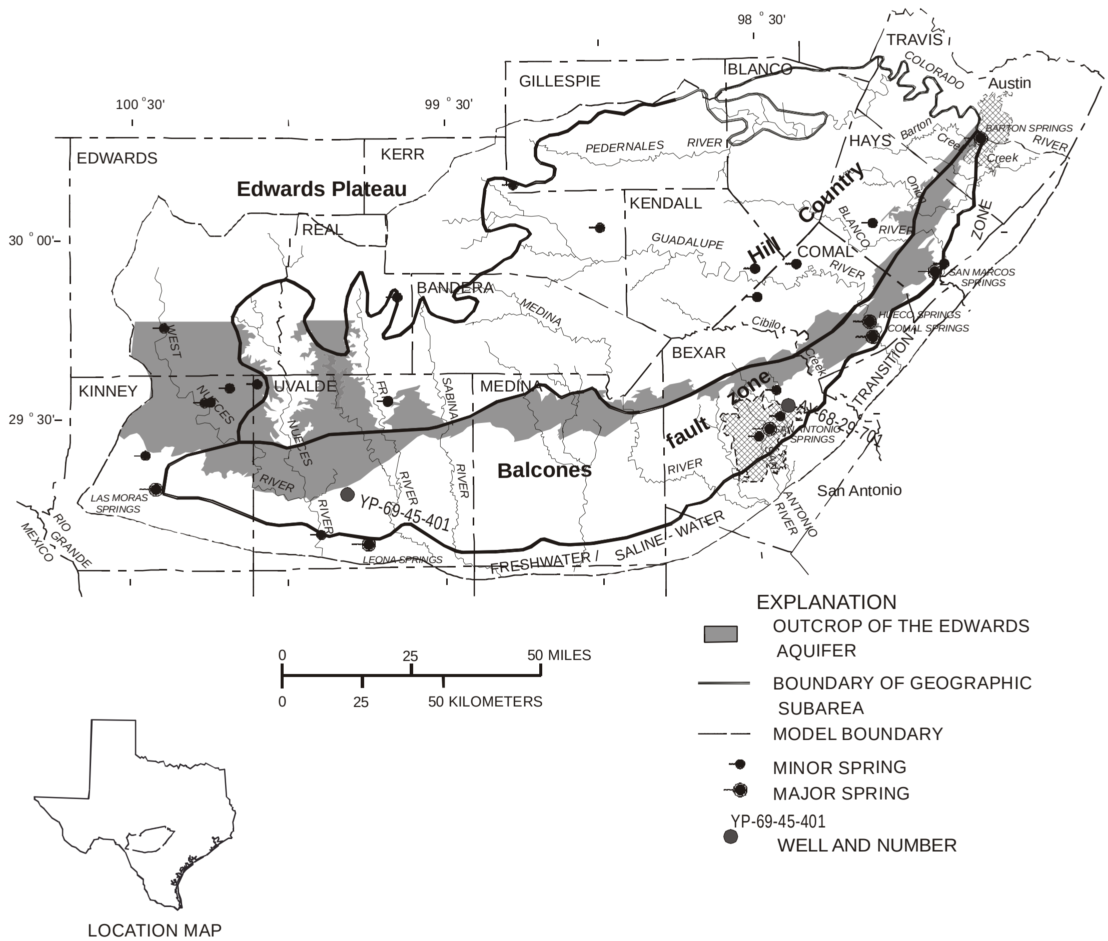


**Figure 1.** Location of study area.

The Balcones fault zone is characterized by a series of en echelon faults that trend southwest to northeast along the length of the region (fig. 2). The terrain within the Balcones fault zone is much less rugged than the Hill Country. Gently rolling hills and wide alluvial valleys are typical near the southeastern border of the fault zone. Surface karst features of karren [surface grooves ranging in width from a few inches to 5 feet (ft)] and tinajitas (dissolved pools in streambeds or formed by springs) are commonly found in and along streams. Shallow sinkholes and swallow holes also are fairly common.

Major rivers within the study area include the Nueces, Frio, Sabinal, Medina, Guadalupe, Blanco, Pedernales and Colorado Rivers all of which incise the Edwards and Trinity aquifers. Within the Hill Country, the majority of the streams are gaining streams. Within the Balcones fault zone, many streams become intermittent because of losses to the Edwards aquifer. Streamflow losses and percolation of rainwater account for the majority of recharge to the Edwards aquifer along its outcrop.

The climate is classified as subhumid, subtropical in the eastern part of the study area and semiarid in the western part. Mean annual rainfall ranges from 32 inches per year (in/yr) in the east to 20 in/yr in the west (1951-80). There are two rainy seasons, spring and fall. Rainfall varies greatly from year to year, but long-term seasonal averages indicate that winter is the driest season. Mean annual temperature is 69 degrees Fahrenheit (Riggio and others, 1987).

Over most semiarid regions of the Edwards Plateau and Hill Country, soil development is poor and generally less than 1 ft thick. In the Edwards Plateau, soils tend to be calcareous stony clays vegetated by desert shrubs in the west and juniper, oak, and mesquite in the east. The Hill Country soils and vegetation are similar to those of the Edwards Plateau. In the northeastern part of the Balcones fault zone, soils are calcareous clay, clayey loam, and sandy loam with some prairie vegetation. West of San Antonio in the southwestern part of the Balcones fault zone, vegetation is predominantly juniper, oak and mesquite (Kier and others, 1977).

In Eve L. Kuniansky, editor, 2001, U.S. Geological Survey Karst Interest Group Proceedings, Water-Resources Investigations Report 01-4011, p. 69-77

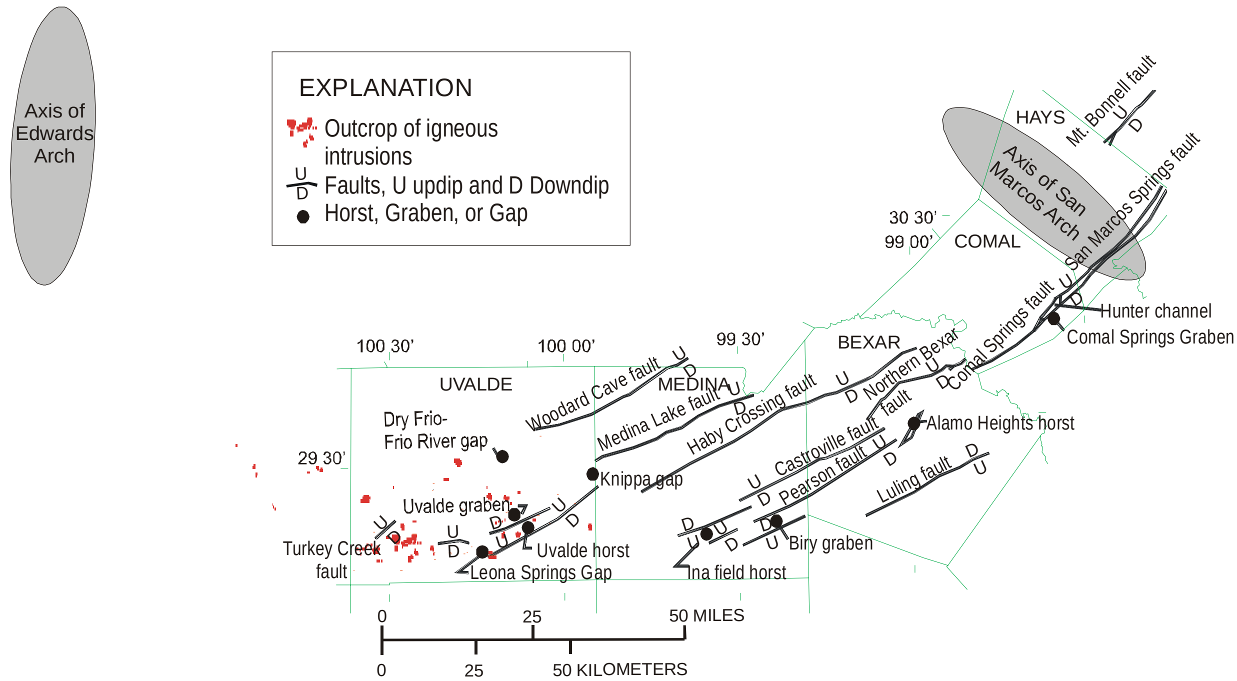


Figure 2.--Structural features that affect ground-water flow direction.

The major aquifers are the Trinity in the Hill Country and the Edwards in the Balcones fault zone. All rock units are of Cretaceous age (Barker and other, 1995; Barker and Ardis, 1996). The Trinity aquifer is composed of dolomitic limestone with interbeds of sand, shale, and clay. The Lower Glen Rose Limestone and the Hensel Sand are the most productive units of the Trinity aquifer. The Upper Glen Rose has been eroded, exposing rocks of the Lower Glen Rose along the Blanco, Guadalupe, and Medina Rivers and Cibolo Creek. The Hensel Sand is exposed along the Pedernales River (Ashworth, 1983). Rocks of the Edwards aquifer (the Edwards Group) have been mostly eroded and cap a few hills in the eastern part of the Hill Country.

The Lower Glen Rose is cavernous in the area of Cibolo Creek (Wermund and others, 1978). Near the confluence of the Pedernales and Colorado Rivers at the northeastern limit of the Hill Country, the lower part of the Trinity aquifer is exposed along the streams. In this area, the most productive units of the Trinity aquifer are the Hosston and Sligo Formations.

The Edwards aquifer is unconfined in a narrow strip where rocks of the aquifer crop out along the southern edge of the Hill Country and the Edwards Plateau. Most of the Edwards aquifer is confined downdip of the outcrop. Rocks that compose the Edwards aquifer tend to be honeycombed, horizontally bedded, and more permeable than rocks of the adjacent Trinity aquifer. Dissolution of rocks that parallel faults and joints has resulted in large secondary permeability. Numerous caves have been mapped within the study area (Wermund and others, 1978).

## STRUCTURAL CONTROLS ON GROUND-WATER FLOW

Faults and structural lineaments have been mapped extensively in the Hill Country and Balcones fault zone. Locations of major faults within the Hill Country and Balcones fault zone are shown in figure 2 along with the location of positive anticlinal features in the pre-Cretaceous surface and the outcrop of igneous intrusions.

Faults, joints, and dissolution of the rocks has greatly affected the ground-water flow system. In part, this is a result of the depositional and diagenetic character of the carbonate bedrock ( Barker and Ardis, 1996). The limestone and dolomite of the Edwards-Trinity is not pure containing clay, shale, and sand. Diagenetic alternation of burrowed limestone beds has resulted in the development of vuggy porosity. The burrowed limestone bedrock members of the Edwards-Trinity aquifer are not the most permeable part of the aquifer system. Solution caverns formed along joints and faults represent the zones of greatest permeability. Fault and fracture zones within the Balcones fault zone created an avenue for meteoric water to percolate through the carbonate rocks. Along with the faulting, joints parallel and perpendicular to the fault system provide an opportunity for the movement of ground water. As streams incised bedrock in the Hill Country

In Eve L. Kuniansky, editor, 2001, U.S. Geological Survey Karst Interest Group Proceedings, Water-Resources Investigations Report 01-4011, p. 69-77

and Balcones fault zone, the development of spring flow further increased the dissolution of rock. Over geologic time, dissolution of carbonate rock developed into a system of caverns and dissolution channels. More caverns formed in the Edwards aquifer, in the Balcones fault zone, than in the Hill Country. These caverns tend to be linear and parallel to the faults or joints (Fieseler, 1978, fig. 4; Wermund and others, 1978, fig. 12; Woodruff and others, 1989, figs 6 and 14; Veni, 1988 p. 12-13). Many caves parallel faults, with some aligned with joints perpendicular to the faults. Veni (1988, p. 13) hypothesized that tensional joints corresponding with many of the en echelon faults, provided preferential ground-water flow paths for the development of caverns and preceded the fault movement.

En echelon normal fault movement has resulted in a series of horst and graben structures. Many of the fault structures form barriers restricting or diverting the lateral movement of ground water. Grabens form flow conduits in the Edwards aquifer (fig. 2, Maclay and Land, 1988).

Two important barrier faults are present along the central part of the Haby Crossing and Pearson faults; here the Edwards aquifer is completely displaced. Other barrier faults include Woodard Cave, Turkey Creek, Medina Lake, Castroville, Northern Bexar, Luling, Comal Springs, San Marcos Springs, and Mount Bonell (Maclay and Small, 1984; Maclay and Land, 1988). In areas where rocks of the Edwards aquifer crop out, erosion and upthrown horst structures have combined to help reduce the saturated thickness of the Edwards aquifer. In the confined part of the system, horst structures have juxtaposed less permeable Trinity rocks with the more permeable rocks of the Edwards aquifer. Important horst stuctures include Uvalde, Ina Field, and Alamo Heights (Maclay and Land, 1988). The Woodard Cave and Mount Bonell faults mark the southeastern boundary of major blocks of the Edwards aquifer, juxtaposing the Trinity aquifer to the northwest with the Edwards aquifer to the southeast (Small, 1986).

The horst and graben structures may combine to divert ground-water flow. The Uvalde graben lies north of the Uvalde horst. Ground water that would normally flow downgradient is obstructed horizontally by the horst structure and thus moves parallel to the horst within the dropped block of the Uvalde graben. The Comal Springs graben, bounded by the Comal Springs fault on the northwest and a series of upthrown blocks to the south, is a narrow area of highly transmissive rocks. The Hunter channel (fig. 2), between Comal and San Marcos Springs, contains highly transmissive rocks.

A series of gaps have formed in areas where minor fault displacement has occurred; the diversion of ground-water flow in these areas is less common. Major gaps include the Dry Frio-Frio River, Leona Springs, and Knippa gaps (fig. 2).

The San Marcos arch is a pre-Cretaceous positive anticlinal feature (fig. 2). The Edwards-Trinity aquifer is thinner over the San Marcos arch (Ashworth, 1983, fig. 7). Localized highs in the pre-Cretaceous base of the aquifer system can reduce the saturated thickness of the more permeable Cretaceous rocks (Barker and Ardis, 1992; Ardis and Barker, 1993) restricting regional ground-water movement. The San Marcos arch has been associated with a ground-water divide in the Edwards aquifer often used as a no-flow boundary for local model studies of the Edwards aquifer (Klemt and others, 1979; Maclay and Land, 1988; Slade and others, 1985). The Edwards Arch is another positive anticlinal feature formed in the pre-Cretaceous surface that resulted in less deposition of lower Trinity rocks near the apex of the arch. The apex of this arch occurs within Edwards County along a south-southwest to north-northeast axis.

Basaltic igneous rocks occur in Uvalde and Medina Counties (fig. 2) and intrude overlying Cretaceous rocks, locally affecting ground-water flow. Although, the subsurface extent of these intrusions are not known, they may impede lateral movement of ground water. Calibration to observed ground-water levels in Uvalde County was improved when the intrusions were simulated as localized areas of reduced transmissivity.

## TRAVEL TIMES ALONG SELECTED FLOW PATHS

Travel times were estimated along flow paths in the Edwards aquifer using simulated ground-water levels. For this analysis, simulated monthly water levels were averaged over an 11-year calibration period (1979-89) to reduce the transient effects of short-term recharge and discharge events. The 1978-89 period was characterized by average to wetter-than-average climatic conditions.

The finite-element, transient flow model of the Edwards aquifer was calibrated to 10 water-level hydrographs, the major and minor springs, and base flows of continuously gaged streams (not shown). The hydrographs shown in fig. 3 show the best and worst fits of the simulation. Simulated water levels at well YP-69-45-401, in Uvalde County, range from approximately 5 to 90 ft too high (worst fit). The well in Bexar County, AY-68-29-701, near the index

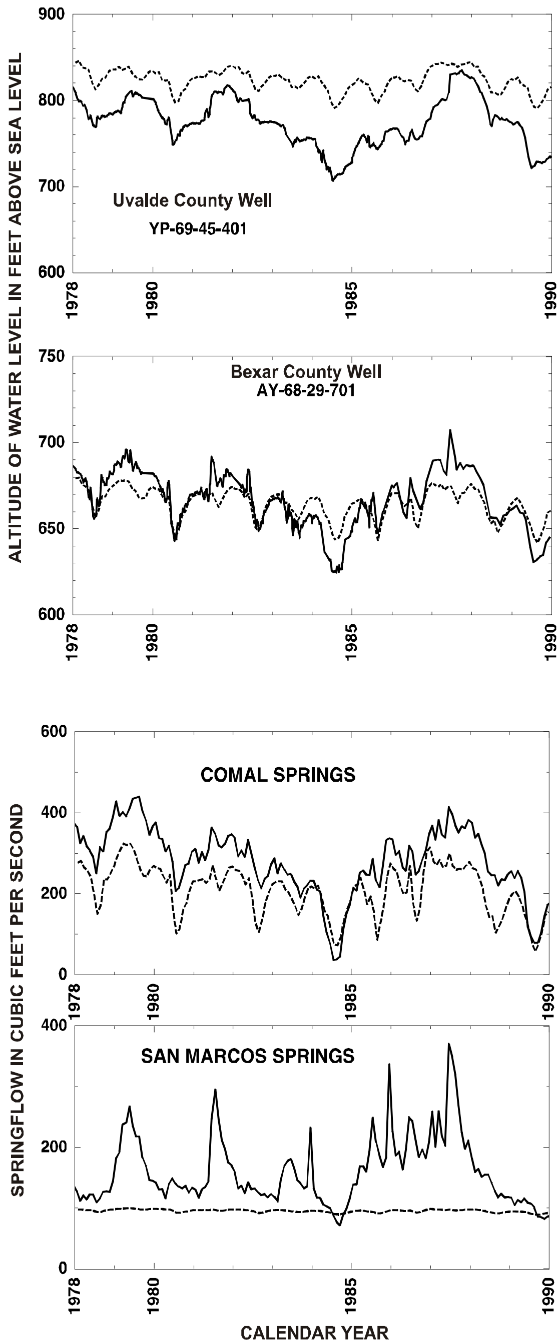


Figure 3.-- Observed (solid line) and simulated (dashed line) ground-water levels and springflow.

well (also known as, J-17) for Comal springs and simulated water levels match observed levels much of the time to a maximum error of approximately 35 ft. The simulated water level is too high in the west (YP-69-45-401 in fig. 3), which means that the simulated gradient is steeper than the actual gradient. A steeper gradient results in faster velocity and travel-time estimates.

Comal Springs matches some of the time, but is simulated with the worst error approximately 110 cubic feet per second less flow than observed springflow. The local recharge at San Marcos Springs was not included in recharge estimates for the model, thus only the base springflow was matched (fig. 3). The effect on travel times of underestimating the springflow is slower estimated velocity and travel times.

The method for estimating times of travel is straightforward. Simulated Darcy flux vectors are calculated for each element of the finite-element model using the average head value for 1978-89 at each node to compute the local gradient for each element (Kuniansky, 1990). The local coordinate system is oriented in the direction of anisotropy, such that all cross products of the transmissivity tensor are zero, thus only the maximum and minimum transmissivity, $T_{xx}$ and $T_{yy}$ , respectively, are non-zero. The gradient in the local coordinate system ($\delta h/\delta x$ and $\delta h/\delta y$) is multiplied times $T_{xx}$ and $T_{yy}$ to compute the Darcy flux ($ft^2$/day) in the local x and y directions. The local flux vectors are then converted to the global coordinate system using the angle of the anisotropy (Kuniansky, 1990). The transmissivity ranges shown on figure 4 are $T_{xx}$, the maximum transmissivity. In the areas with faults (fig. 4), the angle of anisotropy is along the strike of the faults shown. In areas with no major faults or gaps (fig. 2), the aquifer is simulated as isotropic. Dividing the flux vector by aquifer thickness (ft) and porosity (dimensionless) provides an estimate of the advective velocity of a particle of water for that element. Porosity and thickness data (not shown) were obtained from published maps by Hovorka and others (1993).

Flow paths were selected manually by plotting the flow vectors computed from the average simulated potentiometric surface (fig. 4), selecting a starting point, and following the flow vector to an adjacent element until the endpoint (Comal or San Marcos Springs) was reached. The average velocity and distance between elements is computed from the two adjacent elements (fig. 5). The time of travel from one element to the next is computed by dividing the distance by the velocity and summed up along the flow path. In general, the flow paths support much of the work on the conceptual framework of the Edwards aquifer described by Maclay and Small, 1984; Maclay and Land, 1988; and Groschen, 1996.

Flow paths for which travel times were estimated range from 1,250 to 10,000 feet wide and from about 8 to 180 miles long. Effective aquifer thickness and effective porosity can be highly variable and is not well defined throughout most of the aquifer.

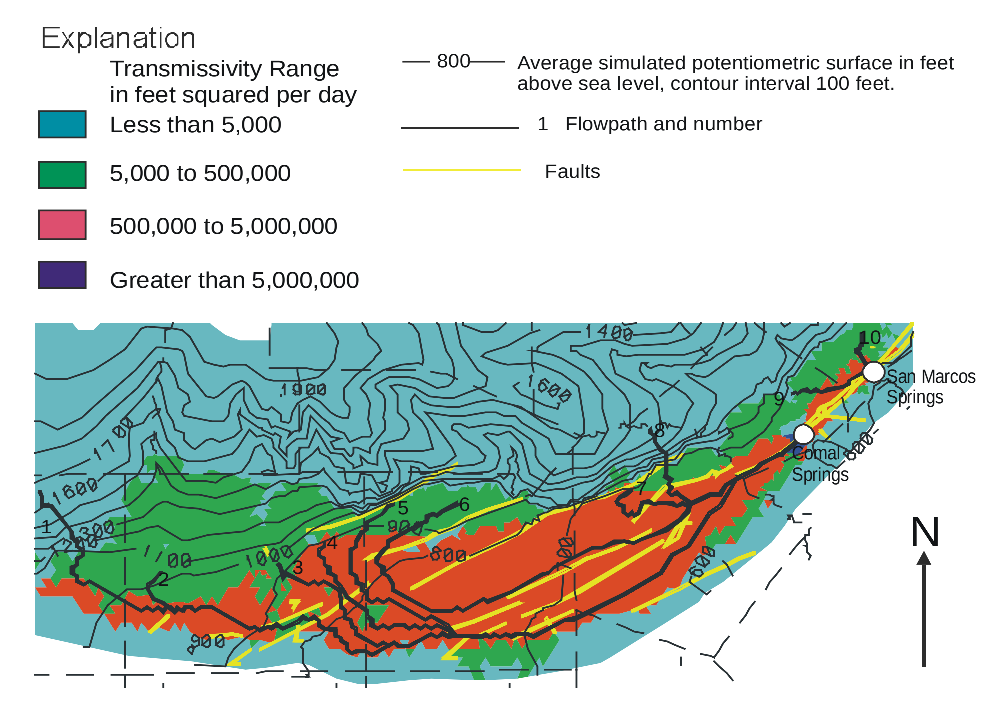


Figure 4.--Selected flow paths from recharge areas to Comal or San Marcos Springs.

Estimates of travel times were computed from aquifer thickness and rock matrix porosities within known or inferred ranges of 350 to 850 ft and 15 to 35 percent, respectively (table 1). Computations involving total aquifer thickness and maximum rock matrix porosity yield maximum travel times. In a karst system, such as the Edwards, the entire thickness of the aquifer may not be the permeable or transmissive zone. Additionally, the rock matrix porosity may not be representative of the effective porosity (connected void spaces). For example, Small and Maclay (1982) report porosity of less than 3 percent for parts of the Edwards aquifer; Sieh (1975) report porosity of less than 1 percent for parts of the Edwards aquifer; Hovorka and others (1993) report effective porosities as low as 5 percent. The minimum rock matrix porosity for each element (range along flow path, tab. 1) were divided by 10 to estimate a minimum time of travel. Travel times range from 14 to 160 years for a flow path from the Blanco River Basin to San Marcos Springs and from 350 to 4,300 years for a flow path from the West Nueces River Basin to Comal Springs. Minimum travel-time estimates are similar in magnitude to the estimates of the age of the water at these springs determined from tritium isotopes in water (Pearson and Rettman, 1976; Pearson and others, 1975). This supports the hypothesis that effective porosity and effective thickness of the aquifer is probably less than its respective range (tab. 1).

Various authors used the tritium data of Pearson and Rettman (1976) to interpret ages for the waters of the Edwards aquifer. Campana and Mahin (1985) used a discrete state compartment model to describe the observed tritium concentrations. This model assumes that water moves from one cell to another as a discrete unit, then mixes completely with water within that cell. Calculated ages were determined as 47 to 132 years from Uvalde County, 57 to 123 years from Medina County, and 38 to 123 years from Bexar County. The estimated age of water was 91 years from Comal Springs, and 16 years from San Marcos springs. More recently, Shevenell (1990) used two hydrologic models, well-mixed and piston flow, to describe the observed tritium concentrations. These two end-member hydrologic models allow determination of interpreted minimum and maximum age dates for observed tritium concentrations. The well-mixed model indicated water from Uvalde County as 96 to 187 years old, Comal Springs water 318 to 521 years old and San Marcos Springs water 61 to 75 years old. The piston-flow model indicated Uvalde County water was 12.5 to 17.9 years old, Comal Springs water was 14.5 to 17.5 years old, and San Marcos Springs water was 10.5 to 15 years old.

The estimated dates obtained from the well-mixed model (Shevenell, 1990) agree more closely with the numerical model than the other hydrogeochemical models. In general, both the numerical model estimates and the geochemical models indicate that the waters obtained from Comal Springs are a mixture of older waters than those obtained from San Marcos Springs.

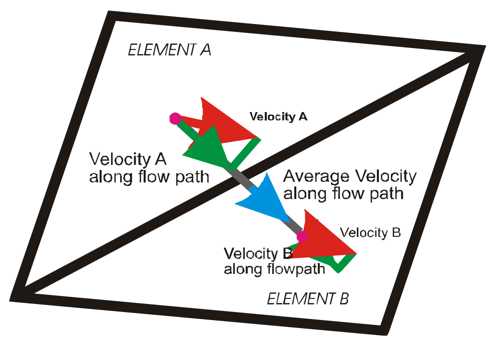


Figure 5.-- Diagram of average velocity calculated between two elements.

## Limitations of the Model and Flow Path Analysis

In developing a numerical model of an aquifer system, many simplifications of the system are required in order to approximate it mathematically. In this quasi three-dimensional finite-element model, ground-water flow is simulated as horizontal and two-dimensional within two model layers, with vertical leakage occurring between layers. The ground-water flow equation solved by the numerical model is the continuity equation for flow with the incorporation of Darcy's law, derived from the principals of conservation of mass and energy along with the assumptions that water is incompressible and of constant viscosity. Mathematically, this is a boundary-value partial-differential equation that is solved numerically. The partial-differential equation is solved for aquifer head, given boundary conditions of specified head, specified flux, or head-dependent flux and aquifer parameters and stresses. The equation is valid for ground-water flow problems when the velocity of ground water is slow and laminar. In karst terranes, it is possible for flow through caverns and dissolution channels to be turbulent. Thus, the equation is not valid for the entire flow domain of the Edwards and Trinity aquifers. A simplification is to assume laminar flow everywhere and an effective transmissivity, so that results are consistent with known hydraulic gradients. The only method to mathematically approximate the effect of horst and graben structural control on lateral ground-water movement in a bedded carbonate unit is to vary transmissivity and to vary the direction and magnitude of anisotropy in a model layer. The range in transmissivity and storage coefficients for the Edwards aquifer used in the model were taken from maps and data published in Maclay and Small (1984); Hovorka and other (1993); and Hovorka and others (1995). In the Hill Country, hydraulic properties for the Trinity aquifer were obtained from well test data and from calibration of a regional one-layer model (Kuniansky and Holligan, 1994). Vertical leakage coefficients between layers were estimated from confining unit thickness (Barker and Ardis, 1996) and rock properties, but were adjusted to be leakier in areas where data indicate cross-formational flow along faults and joints.

A modified version of MODFE (Torak, 1992), a two-dimensional finite-element ground-water flow model was used to simulate ground-water flow in the karstic Edwards aquifer system. This code has not been tested elsewhere, thus, programming errors may exist in the code. Verification of the model code was conducted by comparing the results of an equivalent finite-element mesh (Kuniansky, 1990) using the MODFLOW (McDonald and Harbaugh, 1988) model; both model codes appear to simulate similar ground-water levels and head-dependent flux values. A 20-hour simulation time for 1978-89 using monthly stress periods made parameter estimation and calibration difficult. Thus, it is likely that the model calibration could be improved. Additionally, the lower layer of the model was simulated as a constant head layer using the steady-state simulated initial conditions with both layers actively simulated. This was incorporated to eliminate transient instability in the solution for head in the lower model layer. Transient instability occurred during efforts to simulate the 12 highest monthly recharge events conducted over 144 monthly stress periods within small areas in the lower model layer (relatively low permeability Trinity aquifer beneath outcrop of high permeability Edwards aquifer). The solution for head in the Edwards aquifer did not change as a result of simulating the lower layer as constant head rather than active during the transient simulation.

With all of the limitations described above, simulated heads, spring flows, and base flows reasonably match observed data (Kuniansky and Ardis, [in press]) and transmissivities used for the Edwards aquifer fall within the ranges published by Maclay and Small (1984) and Hovorka and others (1995). Thus, the estimated direction of flow and Darcy flux along selected flow paths is considered to be reasonable. The least conclusive aspect of the analysis is associated with estimates of pore velocity and times of travel due to the poor understanding of effective aquifer thickness or the distribution of effective porosity within the Edwards aquifer.

Table 1. Summary of flow path analysis for average simulated potentiometric surface, 1978-89.

| Flowpath<br>Number and description | Thickness[1](feet)<br>minimum to maximum, average | Porosity[1] (percent)<br>minimum to maximum, average | Distance (miles) | Average Velocity (feet per day)<br>minimum to maximum Average | Time (years)<br>minimum to maximum |
|---|---|---|---|---|---|
| 1. West Nueces River to Comal Springs | 450 to 850, 620 | 15 to 35, 23 | 180 | 0.024 to 50, 6.9 | 350 to 4,300 |
| 2. Nueces River to Comal Springs | 450 to 850, 610 | 15 to 35, 22 | 149 | 0.024 to 50, 8.1 | 210 to 2,600 |
| 3. Frio River to Comal Springs | 450 to 850, 600 | 15 to 35, 22 | 122 | 0.40 to 50, 9.8 | 69 to 790 |
| 4. Sabinal River to Comal Springs | 450 to 850, 580 | 15 to 35, 23 | 114 | 0.017 to 50, 9.8 | 65 to 780 |
| 5. Hondo Creek to Comal Springs | 450 to 750, 560 | 15 to 35, 22 | 120 | 0.99 to 60, 12 | 49 to 600 |
| 6. Verde Creek to Comal Springs | 450 to 750, 530 | 15 to 28, 22 | 111 | 0.65 to 50, 13 | 29 to 360 |
| 7. Northwest of San Antonio to Comal Springs | 450 to 450, 450 | 15 to 28, 24 | 46 | 0.24 to 50, 14 | 27 to 320 |
| 8. Cibolo Creek to Comal Springs | 350 to 450, 430 | 15 to 28, 24 | 43 | 0.05 to 50, 15 | 200 to 2,400 |
| 9. Guadalupe River to San Marcos Springs | 400 to 500, 460 | 24 to 28, 26 | 16 | 0.14 to 23, 8.6 | 23 to 270 |
| 10. Blanco River to San Marcos Springs | 400 to 500, 450 | 24 to 28, 26 | 8 | 0.30 to 7.3, 2.7 | 14 to 160 |

[1] From Hovorka and others, 1993

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# Freshwater Macrofauna of Florida Karst Habitats

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## INTRODUCTION

The caves, springs, and seeps that comprise karst habitats within the United States harbor unique and diverse faunal assemblages that are biologically important due to their high degree of endemicity. Culver and others (2000) summarized the described obligate cave fauna of the U.S. and determined that 594 (61%) of a total of 973 taxa are confined in distribution to one county; nearly a third of all taxa occur at single sites. The aquatic fauna is a major component of the karst biota and includes about one third of all known cave species. In fact, the distribution of the aquatic biota across karst habitats accounts for an estimated 4% of the total surface area of the contiguous U.S. (Culver and others, 2000). Karst systems of Florida contain among the most diverse aquatic faunas nationwide. Within Florida, the greatest karst biodiversity is found in the northern peninsula and east-central panhandle. Franz and others (1994) reviewed the cave faunas of Florida and southern Georgia and identified 267 biologically important caves serving as critical habitat for populations of 27 invertebrate and one vertebrate taxa, of which nearly all species (93%) are aquatic.

Compared to cave faunas, fewer synoptic studies are available for the myriad of spring habitats and species of the U.S. Williams and Smith (1990) provided an extensive international bibliography of spring habitats and their faunas. Few comprehensive surveys exist of the biota of Florida's extensive spring habitats. Woodruff (1993) summarized previous literature, conducted a limited survey of 13 selected Florida springs, and developed a classification system based on a cluster analysis of springs using water chemistry data provided by Rosenau and others (1977), the U.S. Geological Survey, Water Management Districts, and other sources. Mattson and others (1995) examined the biota of springs and spring-influenced streams of the Suwannee River drainage in northwest Florida and included a synopsis of the periphyton and benthic invertebrate communities.

The purpose of this paper is to summarize the relevant literature and information on the aquatic macrofauna of Florida karst habitats. The biota of submerged caves and springs are considered together because of the integral connection between subterranean and surface ground-water habitats in the state and the similar ecological conditions that exist for karst-adapted species. Major sources of information on organisms in Florida's karst environments are the synoptic works by Thompson (1968), Franz and Lee (1982), Woodruff (1993), Franz and others (1994), Deyrup and Franz (1994), Mattson and others (1995), and numerous original descriptions of both hypogean and epigean species.

## STUDY AREA

Florida has expansive karst areas that include a combination of diverse and globally unrivaled large-magnitude springs, caverns, caves, sinks, disappearing streams and lakes, and complex subterranean aquifers (Rosenau and others 1977; Lane 1986; Miller 1997). Four principal aquifer systems are exposed at the surface or covered by a thin layer of confining soils and receive recharge primarily via precipitation (Figs.1, 2).

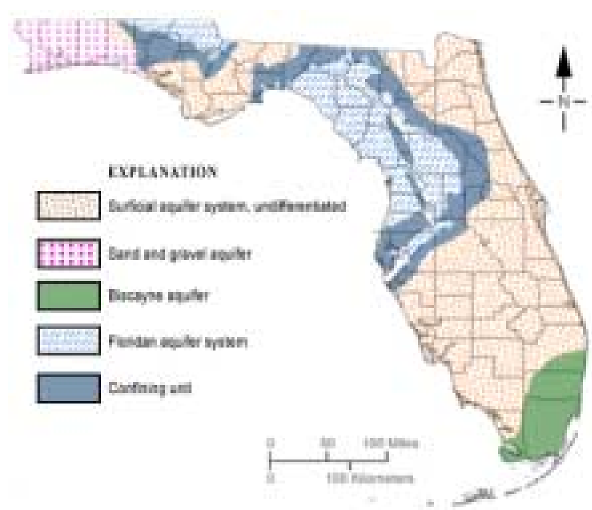


**Figure 1.** Principal aquifers of Florida exposed at or near the land surface (after Miller 1997).

In Eve L. Kuniansky, editor, 2001, U.S. Geological Survey Karst Interest Group Proceedings, Water-Resources Investigations Report 01-4011, p. 78-88


**Figure 2.** Extent of surficial contact with the Floridan Aquifer (from Miller 1997; unconfined = absent or thin; thinly confined < 100 ft; confined > 100 ft).

The largest system and major source of ground water, the Floridan aquifer, consists of a thick sequence of Tertiary carbonate rocks with a complex lithological profile defined by soil characteristics, permeability, geological depositions, and erosional history (Miller 1997). The smaller Biscayne aquifer, which serves as the primary source of ground water for largely urbanized areas of southeast Florida, is also comprised mainly of carbonate rocks. These two carbonate-rock aquifers are more mineralized than other Florida aquifers that are composed of siliclastic rocks. Because of their complex history, geomorphology, and ecological characteristics, the carbonate-rock aquifers provide important habitats for unique assemblages of spring- and cave-adapted organisms.

## METHODS

Information on aquatic species in karst habitats of Florida was obtained from a variety of published sources. Emphasis is on macroscopic mollusk and crustacean invertebrates and one vertebrate that are obligately associated with ground-water environments. Synoptic taxonomic, distributional, and microhabitat data were obtained primarily from Hobbs (1942, 1989), Hobbs and others (1977), Thompson (1968, 1984), Franz and Franz (1990), Deyrup and Franz (1994), and Franz and others (1994). Common names of decapod crustaceans and mollusks follow Williams and others (1989) and Turgeon and others (1998) except where species are unnamed, where other authors (e.g., Franz and others, 1994) propose alternative names, or names are suggested herein by inclusion within quotation marks. Conservation status was determined from Deyrup and Franz (1994), Taylor and others (1996), and the official list of state-protected taxa by the Florida Fish and Wildlife Conservation Commission. The following terminology, after Franz and others (1994), Culver and others (2000), and additional sources, applies to taxa associated with cave and ground-water habitats: troglobites are forms that are confined to caves and exhibit unique morphological specializations for subterranean life (e.g., depigmentation, reduction or loss of eyes, and development of accessory sensory structures); stygobites are aquatic troglobites; phreatobites are taxa that inhabit interstitial ground waters and may also occur regularly in some cave and spring habitats; troglophiles are taxa that may complete parts or all of their life history in caves as well as epigean habitats, but lack the extreme morphological specializations of troglobites; trogloxenes occur regularly in caves but do not complete their life cycles in them; accidentals are species that do not normally inhabit caves but are occasionally encountered in them; endemic taxa (= precinctive of Franz and others, 1994) are those that occur in highly localized habitats, often single caves, springs and spring-runs, and single aquifer systems. Occurrence of species by drainage or faunal region is based exclusively on published distributional information (Fig. 3, Table 1).

## RESULTS AND DISCUSSION

### Cave Habitats

Cave habitats in Florida with readily accessible entry points have historically generated considerable interest to speleologists and scientists. Franz and others (1994) provided a detailed history of biospeleological exploration of the state's cave habitats. This history includes notable contributions by various early explorers during the late 1800s, prolific crustacean studies of the late Horton H. Hobbs, Jr., and associates that culminated in a landmark monograph on Florida's crayfish fauna (Hobbs, 1942), and the important modern period (1970-1992), during which time Richard Franz (Florida Museum of Natural History) and colleagues continued a rich period of exploration, discovery, and description of new species, coupled with a synthesis of existing information and elucidation of distributional, phylogenetic, and ecological data.

In Eve L. Kuniansky, editor, 2001, U.S. Geological Survey Karst Interest Group Proceedings, Water-Resources Investigations Report 01-4011, p. 78-88


**Figure 3.** Surficial hydrologic subregions identified by Rosenau and others (1977). Each subregion is an assemblage of smaller hydrologic units. 1=Choctawhatchee-Yellow-Escambia; 2=Apalachicola-Chattahoochee-Flint; 3=Ochlockonee; 4=Aucilla-Suwannee; 5= Altamaha-St. Marys; 6=St. Johns; 7=west coastal area (Withlacoochee, Hillsborough, Peace, and others); 8=southern Florida.

Approximately 630 caves have been identified in Florida by the Florida Speleological Society, of which Franz and others (1994) reported at least 267 (42%) as having important macroscopic troglobitic faunas. Most of the species in these caves are aquatic, although these authors also included terrestrial taxa as well as those that are considered to be facultative in cave environments (i.e., troglophiles, trogloxenes, and accidentals). The most important cave habitats are concentrated in the north-central peninsula and eastern panhandle, with over half of the biologically significant caves identified by Franz and colleagues located in four counties : Alachua (47 caves), Suwannee (43), Jackson (34), and Marion (27). Twenty additional counties in Florida and two in southern Georgia have caves, ranging in number from one to 15, that are important in the distribution of troglobites.

The known stygobites of Florida and southern Georgia include 18 crayfishes (including two subspecies and intergrade populations), three isopods, two amphipods, one shrimp, one snail, and one salamander (Franz and others, 1994; Table 1). Several of these taxa are undescribed due to a paucity of critical material with diagnostic morphological characters (e.g., reproductively mature male crayfishes). Greatest study has been devoted to the decapod crustaceans due in part to their remarkable radiation in Florida's cave habitats. In addition to the free-living species, there are at least four obligate commensals on stygobites in circum-Florida karst assemblages: *Cambarincola leoni* and other unidentified branciobdellid annelids reported from seven crayfish populations; and, three species of entocytherid ostracods (*Uncinocythere ambophora*, *U. equicurva*, and *U. lucifuga*), also recorded from several crayfish species. Several additional ostracods and copepods were retrieved either free swimming or from the gut of the stygobitic salamander, *Haideotriton wallacei*; however, the taxonomy, distribution, and ecology of these microcrustaceans are largely unstudied. Additionally, troglobitic and trogloxenic springtails (Insecta: Collembola) are often closely associated with subterranean aquatic habitats and may be important prey for stygobites. Franz and others (1994) provided a detailed, annotated list of other cave-associated species that includes 23 troglophiles, 47 trogloxenes, and 37 accidentals. The non-troglobitic organisms reported by these authors include a variety of terrestrial and aquatic non-arthropod invertebrates, arthropods (especially insects, epigean crustaceans, and arachnids), and vertebrates. The principally epigean vertebrate fauna consists of nine fishes, 16 amphibians, nine reptiles, two birds, and 13 mammals, including six bats.

Study of the zoogeography of Florida's troglobitic fauna has been refined through the years as new distributional data have become available and as geologic and hydrologic information has been incorporated. Early investigations such as those by Hobbs (1958), Caine (1974a), Relyea and others (1976), Hobbs and others (1977), and Franz and Lee (1982) were summarized and updated by Franz and others (1994) to include the most recent interpretations. The obligate cave faunas fall into six distinct regions, each characterized by a unique combination of hydrological, geological, and ecological characteristics: (1) Econfina Creek Fauna, with two endemic taxa (*Dasyscias franzi* and *Caecidotea* sp. 1), is known from a single cave associated with a limited karst area in Washington and Bay counties; (2) Apalachicola Fauna, with three endemic taxa (*Cambarus cryptodytes*, *Islandiana* sp., *Haideotriton wallacei*), in two segments (Marianna Lowlands, Jackson County, FL, and Dougherty Plain, lower Flint River, GA) with caves that are developed near the boundary of the Ocala Group, Suwannee, and Marianna limestones (all formations of the Floridan Aquifer); (3) Woodville Fauna, with two endemic taxa (*Procambarus horsti*, *P. orcinus*) associated with the Ocala Group limestones in eroded segments of the Tallahassee Hills and Woodville Karst Plain near and below the Cody Scarp (boundary between Northern Highlands and Gulf

Coastal Lowlands), paralleling riverine karst areas of the Wakulla-St. Marks and Wacissa rivers; (4) Ocala Fauna, containing a rich stygobitic group that includes nine endemic taxa (*Palaemonetes cummingi*, *Procambarus erythrops*, *P. franzi*, *P. leitheuseri*, *P. lucifugus lucifugus*, *P. l. alachua*, *P. l. lucifugus* X *P. l. alachua*, *P. pallidus*, *Troglocambarus maclanei*) in six geographically- and faunistically-distinct assemblages occurring in mature and riverine karst areas associated with the Ocala Group limestones from the Suwannee River drainage southward to Pasco County; (5) St. Johns River Fauna, containing seven endemics (*Caecidotea* sp. 2, *Procambarus acherontis*, *P. attiguus*, *P. delicatus*, *P. morrisi*, *P.* (*Lonnbergius*) sp., and *Troglocambarus* sp.) in two assemblages developed near contact zones of the Hawthorne Formation and underlying Ocala Group limestones, the Wekiva Assemblage in a small karst area along the Wekiva River, and the Lake George Assemblage, occurring along the western fringe of Lake George, Alexander Springs, and the lower Oklawaha River; and, (6) the Miami Fauna, a small assemblage containing one endemic species (*Procambarus milleri*) in solution holes and shallow wells associated with the Miami oolite formation of the Biscayne Aquifer. Several faunal regions include troglobitic taxa that also occur in at least one other region. Hobbs and others (1977), Franz and Lee (1982), and Franz and others (1994) provided detailed discussion of hypothesized zoogeographic relationships among the stygobitic crayfishes, including possible speciation and dispersal events.

Ecological data on stygobites of Florida are largely anecdotal. The most detailed invertebrate study to date was the comparison of hypogean and epigean crayfishes by Caine (1974b, 1978). Other studies documenting life-history and habitat information of cave crayfishes include Relyea and Sutton (1973a, b). Published taxonomic descriptions of some species include limited data on ecology, habitat, and associated species. Perhaps the most significant aspect of the ecology of Florida's stygobites, like troglobites elsewhere, are the limitations imposed by restricted energy input into the systems in which they occur. In many caves, allochthonous nutrients and detritus transported via ground water are the primary source of energy that supports troglobitic communities. However, some Florida caves have important bat colonies that produce large amounts of guano that sustains the subterranean fauna (Franz and others, 1994). Streever (1996) reported relatively low organic carbon influxes to Sim's Sink cave (Suwannee County) and speculated on possible energetic limitations to the endemic crayfish *Procambarus erythrops*. In a few systems, there may be flocculent mats of chemautotrophic bacteria that serve as a trophic base for stygobitic grazers (Hobbs and Franz, 1992; Franz and others, 1994).

Unlike the rich invertebrate radiation that has evolved in ground waters of the state, submerged Florida caves are unusual in their lack of vertebrate diversity. Other karst regions of North America (e.g., Edwards Aquifer, Interior Highlands) have prominent stygobitic fishes, but Florida has no true troglobitic fishes. Nevertheless, various cave systems in Florida are utilized by a small number of fish species. The Redeye Chub (*Notropis harperi*) exhibits the closest apparent association with Florida springs and caves (Marshall 1947) and has been observed deep in subterranean waters of the St. Johns River drainage. Relyea and Sutton (1973c) observed bullhead catfishes (*Ameiurus natalis*) in close association with Redeye Chub in isolated caves of the Suwannee River drainage, and speculated that both species may have been spawning in ground-water habitats. Diet analyses of the catfish revealed significant predation on troglobites, and fish exhibited fin abnormalities suggesting possible inbreeding from low population size. However, it remains unknown as to whether either of the above species is capable of completing its life history within caves. Other notable trogloxene and accidental fishes observed in Florida caves include the American Eel (*Anguilla rostrata*), Pirate Perch (*Aphredoderus sayanus*), Eastern Mosquitofish (*Gambusia holbrooki*), Striped Bass (*Morone saxatilis*), Brown Bullhead (*Ameiurus nebulosus*), Bluegill (*Lepomis macrochirus*), and Black Crappie (*Pomoxis nigromaculatus*) (Franz and others, 1994). There have been few ecological studies of subterranean fish populations in Florida. Helfman (1986) observed daily movements of American Eel from food-limited cave interiors to shallow-habitat foraging areas.

## Spring Habitats

Florida is renown for its extensive springs and spring-fed rivers. In their landmark monograph that continues to serve as a primary source of information, Rosenau and others (1977) surveyed the state's springs and provided detailed data on water chemistry and physical features. Springs are generally classified on the basis of geomorphology, flow, temperature, water chemistry, and other physical characteristics, excluding, for the most part, biological communities. There are 27 first-magnitude springs (those having a water discharge > 100 $ft^3$/sec) in the state, providing Florida with the largest number and discharge of major springs in any area of similar geographic size; these large springs account for about 80% of the total ground-water discharge for the state. Most of Florida's springs are artesian, formed as hydrostatic pressure of ground water in karst sediments rises to the surface and resurges through natural breaches in impermeable or thin confining layers (Fig. 1). There are also numerous water-table springs in the state, formed by percolation of surface water through permeable sediments and sheet flow along the gradient of an impermeable layer to an outcrop point where water issues forth as a seep or non-artesian spring. Many of the major springs in Florida are situated in the northern half of the peninsula and are associated with the high Tertiary stratum known as the Ocala Uplift. The elastic-sediment veneer of the uplift was deposited more thinly and eroded more quickly than other confining layers throughout the state, thus exposing the Floridan Aquifer at the surface. In addition, other spring concentrations are found in areas of surficial down-cutting, such as sinks and river valleys (e.g., the Suwannee River drainage). The Floridan Aquifer is not the sole source of water for the state's artesian springs; for example, many springs in the Central Highlands and eastern panhandle are supplied by intermediate aquifers (Woodruff, 1993). General descriptions of the geology and hydrology of Florida's ground-water resources is provided by Rosenau and others (1977), Lane (1986), and Miller (1997).

Woodruff (1993) proposed a revised classification of Florida's springs based on a detailed analysis of water chemistry data. Using cluster analysis of the six predominant ions, he recognized four basic groupings among 170 Florida springs: (1) low ionic springs; (2) calcium bicarbonate springs; (3) mixed springs; and, (4) salt springs. There is a gradient of increased ionic concentration across the groupings from low-ionic to salt springs, especially due to sodium and chloride. The most common spring type, calcium bicarbonate (76% of springs examined), are formed from dissolution of limestone and dolomite principally in the Floridan Aquifer and are typically associated with karst terrain subjected to river down-cutting extending from the panhandle southward through the peninsula. Woodruff (1993) recognized three subgroupings of calcium bicarbonate springs: a low-calcium bicarbonate subgroup distributed in the western panhandle and Central Highlands; an intermediate subgroup with a similar range but also including springs in the eastern panhandle, upper Suwannee, and St. Johns river drainages; and, a high-calcium bicarbonate subgroup largely confined to the Ocala Uplift region within the Suwannee River drainage. Salt springs (12% of those examined) and mixed springs (8%) occur primarily near coastal zones where salt deposits or saltwater intrusion influence ionic composition. Along the west coast, these spring types are found from the St. Marks River drainage (Wakulla County) southward to Sarasota County, and on the east coast these spring types emerge inland within the St. Johns River drainage. Springs of the smallest group, the low-ion type (4%), are confined to the panhandle and northern peninsula where they resurge from surficial or intermediate aquifers. Woodruff's (1993) delineation differs slightly from that of Miller (1997; Fig. 6.4.1), but the distribution and physicochemical composition of each spring type corresponds well with regional geological and hydrological features of the aquifer system.

The extensive springs and spring-fed streams of Florida provide important habitats for rich biological communities (Nordlie 1990) that include a number of obligate spring taxa. Neverthelesss, relatively few detailed studies exist on the diversity, distribution, and ecology of the facultative and spring-dependent fauna and associated habitats in Florida springs. Woodruff (1993) found that the diverse chemical composition of Florida's springs has a strong influence on representative aquatic assemblages. Although the relatively stable physicochemical conditions of many springs contribute to extremely high productivity rates, other aspects of water chemistry (e.g., low dissolved oxygen levels) may be limiting for many species, especially near the vent (McKinsey and Chapman, 1998). Florida springs and spring runs support diverse, productive algal and macrophyte communities that serve as a trophic base for primary consumers while providing essential habitat for other species (Whitford 1956; Woodruff 1993; Mattson and others, 1995). The obligate spring-dwelling macrofauna of Florida is dominated by the diminutive hydrobiid gastropods, including many species that have highly restricted ranges often confined to single springs or reaches of spring runs (Table 1; in addition to those listed, there are at least four recently discovered new species of *Cincinnatia* confined to small springs in Seminole State Forest, Lake County; F.G. Thompson, pers. com.). Many other invertebrates and vertebrates utilize springs facultatively. Populations of some benthic invertebrates (e.g., gastropods, crustaceans) may reach extraordinary levels in spring habitats. These species are important primary consumers and a significant prey base for other organisms. Woodruff (1993) reported densities of amphipods exceeding 26,000 per $m^2$ in some springs. The survey by Mattson and others (1995) provides an example of the rich diversity of macroinvertebrates that populate Florida springs and spring-fed rivers (Table 2). Among the more visible groups of facultative species in springs are fishes; notable studies of Florida fishes in spring habitats and spring-fed streams include those of Hubbs and Allen (1943) and Hellier (1967). The ichthyofauna of Florida springs includes a large number of resident freshwater species, as well as many marine invaders that are able to persist due to the ionic composition of bicarbonate spring effluents. Many marine-derived species penetrate spring-fed systems far inland, have established breeding populations, and/or are diadromous, including the Striped Mullet (*Mugil cephalus*), Hogchoker (*Trinectes maculatus*), Atlantic Stingray (*Dasyatis sabina*), Gulf Pipefish (*Syngnathus scovelli*), Mountain Mullet (*Agonostomus monticola*), Gulf Sturgeon (*Acipenser oxyrinchus desotoi*), and many other species. One of the most familiar vertebrates in Florida springs is the West Indian Manatee (*Trichechus manatus latirostris*).

Spring seeps are a small component of karst habitats in Florida, yet they often have unusual and highly localized biological communities. The majority of biologically important seep habitats occur in the northern portion of the peninsula and in the panhandle. Table 1 includes six species of dragonflies that have limited distributions near seep areas within Florida; some of these species have relatively widespread distributions in the southeastern U.S., but their habitat requirements appear to be confined to seepages. Other species may also be limited to or exhibit strong ecological affinities for seep habitats. Some burrowing crayfishes, especially in Coastal Plain drainages of the panhandle, are closely tied to seep areas (e.g., *Cambarus pyronotus*, *C. striatus*, *Procambarus rogersi*). *Procambarus geodytes* is a burrowing species that appears to be confined to sulfur and mineral seeps in the St. Johns River drainage (Hobbs, 1942).

## CONSERVATION

The endemic fauna of Florida karst habitats includes many highly vulnerable species. Of the 27 stygobites listed in Table 1, nearly one third are known from only one cave system, and 67% of the taxa are reported from 10 sites or less. The most recorded endemic taxa are from the St. Johns River (8) and the Suwannee River (6) drainages, with the fewest known from the western panhandle and southern Florida regions. The Florida Committee on Rare and Endangered Plants and Animals (FCREPA; Deyrup and Franz, 1994) recognized three species of special concern, 15 as rare, two as threatened, and one as endangered. In contrast, Taylor and others (1996) considered the crayfish fauna alone to include two species of concern, three as threatened, and nine as endangered. The Florida Fish and Wildlife Conservation Commission currently recognizes the following taxa on its official list of endangered (E), threatened (T), and species of special concern (SSC): Squirrel Chimney Cave Shrimp, *Palaemonetes cummingi* (T); Santa Fe Cave Crayfish, *Procambarus erythrops* (SSC); and, Georgia Blind Salamander, *Haideotriton wallacei* (SSC). Currently, *Palaemonetes cummingi* is the only stygobite of Florida karst that is afforded protection under the U.S. Endangered Species Act. The spring-obligate gastropod fauna includes species with highly restricted ranges, yet of those listed in Table 1, only two are considered endangered, three threatened, and three of special concern (Deyrup and Franz, 1994). Fourteen (78%) of the hydrobiid and pleurocerid snails are known from only single sites, and the majority of Florida's karst-limited gastropods occur in isolated springs of the St. Johns River drainage. At present, the endemic stygobitic and spring species of Florida's karst habitats may be inadequately protected insofar as listing by state or federal natural-resource agencies. All seep-associated dragonflies in Table 1 were considered imperiled by FCREPA, but none of them are currently pending or listed by the U.S. Fish and Wildlife Service.

Like many other geographic regions with prominent karst systems, the most significant threats to Florida's karst biota relate to human activities (Drew and Hötzl, 1999; Walsh, 2000). Spring and cave species are especially susceptible to habitat loss, ground-water contamination, aquifer withdrawals, saltwater intrusion, and competition or predation by nonindigenous species. Springs are frequently modified for consumptive or recreational purposes, with concomitant impacts on aquatic organisms. Many of Florida's karst species are threatened by habitat modifications due to their very localized distributions. For instance, the Enterprise Siltsnail (*Cincinnatia monroensis*) may have been eliminated by hydroelectric development at the single spring that it is confined to. Perhaps the most serious potential threat to Florida's hypogean and spring faunas is ground-water pollution and/or saltwater intrusion as land surface is developed and aquifer resources are increasingly tapped. Streever (1992, 1995) reported on a kill and post-kill recovery of the troglobitic Santa Fe Cave Crayfish (*Procambarus erythrops*) and three troglophiles that may have been due to physicochemical changes associated with flushing of contaminants and/or Suwannee River water during a flood event. In recent years, there have been notable increases in contaminants and nutrients within some Florida ground-water sources (e.g., Katz and others, 1999). Eutrophication in spring habitats may result in greater algal growth, increased turbidity, and physicochemical and biological changes that can be detrimental to native species. Although ecological effects of nonindigenous animals in Florida karst are largely undocumented, introduced Asiatic thiarid snails (Fawn Melania, *Melanoides turricula*; Quilted Melania, *Tarebia granifera*) have been found to displace native hydrobiid snails in several Florida springs (Thompson, 1984).

Florida has a rich and globally significant karst biota that has long interested biospeleologists. Consequently, considerable information is known about the taxonomy and zoogeography of this fauna. However, species remain to be discovered, systematic studies are incomplete, ecological data are largely lacking, and distributional and status information is dated. Thus, there is an urgent need to better document Florida's karst organisms in order to understand, protect, and effectively manage these assemblages.

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**Table 1.** Obligate and predominate macrofaunal species of Florida karst habitats. Distributional occurrences (X) correspond with surficial hydrologic subregions of Rosenau and others (1977; see Figure 3; no obligate species are recorded in drainage 5). Conservation status categories are: E=Endangered, T=threatened, R=Rare, SSC=Species of Special Concern (FCREPA; Deyrup and Franz, 1994); E=Endangered, T=Threatened, SC=Special Concern (AFS; Taylor and others, 1996). Sites=number of cave or spring localities reported.

| | | | **Hydrologic Subregion** | | | | | | | **Conservation Status** | |
|---|---|---|---|---|---|---|---|---|---|---|---|
| | | **sites** | **1** | **2** | **3** | **4** | **6** | **7** | **8** | **FCREPA** | **AFS** |
| **CAVES** | | | | | | | | | | | |
| Gastropoda | | | | | | | | | | | |
| Hydrobiidae | | | | | | | | | | | |
| *Dasyscias franzi* | Shaggy Ghostsnail | 1 | X | | | | | | | | |
| Amphipoda | | | | | | | | | | | |
| Crangonyctidae | | | | | | | | | | | |
| *Crangonyx grandimanus* | Florida Cave Amphipod | >20 | | | X | X | X | X | X | SSC | |
| *Crangonyx hobbsi* | Hobbs' Cave Amphipod | >20 | | | X | X | X | X | X | SSC | |
| Isopoda | | | | | | | | | | | |
| Asellidae | | | | | | | | | | | |
| *Caecidotea hobbsi** | Florida Cave Isopod | 10-11 | | X | | X | X | | | SSC | |
| *Caecidotea* sp. 1* | “Econfina Cave Isopod” | 1 | X | | | | | | | | |
| *Caecidotea* sp. 2* | “Rock Springs Cave Isopod” | 1 | | | | | X | | | | |
| *Remasellus parvus** | Swimming Little Florida Cave Isopod | 4 | | | X | X | | | | R | |
| Decapoda | | | | | | | | | | | |
| Cambaridae | | | | | | | | | | | |
| *Cambarus cryptodytes* | Apalachicola Cave Crayfish | 20 | | X | | | | | | R | T |
| *Procambarus acherontis* | Orlando Cave Crayfish | 6 | | | | | X | | | T | E |
| *Procambarus attiguus* | Silver Glen Springs Cave Crayfish | 1 | | | | | X | | | R | E |
| *Procambarus delicatus* | Big-Cheeked Cave Crayfish | 1 | | | | | X | | | R | E |
| *Procambarus erythrops* | Santa Fe Cave Crayfish | 5 | | | | X | | | | R | E |
| *Procambarus franzi* | Orange Lake Cave Crayfish | 4 | | | | | X | | | R | E |
| *Procambarus horsti** | Big Blue Springs Cave Crayfish | 3-4 | | | X | X | | | | R | E |
| *Procambarus leitheuseri* | Coastal Lowland Cave Crayfish | 8 | | | | | | X | | R | E |
| *Procambarus lucifugus alachua* | Alachua Light-Fleeing Cave Crayfish | 13 | | | | X | | | | R | T |
| *Procambarus lucifugus lucifugus* | Withlacoochee Light-Fleeing Cave Crayfish | 2 | | | | | | X | | R | E |
| *P. l. lucifugus* X *P. l .alachua* | [intergrade populations] | 16 | | | | X | X | | | | |
| *Procambarus milleri* | Miami Cave Crayfish | 2 | | | | | | | X | R | E |
| *Procambarus morrisi* | Putnam County Cave Crayfish | 1 | | | | | X | | | T | E |
| *Procambarus orcinus* | Woodville Karst Cave Crayfish | 15 | | | X | | | | | R | T |
| *Procambarus pallidus* | Pallid Cave Crayfish | >20 | | | | X | | | | R | SC |
| *Procambarus* sp. | “Hawthorne Cave Crayfish” | 1 | | | | | X | | | | |
| *Troglocambarus maclanei* | Northern Spider Cave Crayfish | 16 | | | | X | X | X | | R | SC |
| *Troglocambarus* sp. 1 | Orlando Spider Cave Crayfish | 1 | | | | | X | | | | |
| Palaemonidae | | | | | | | | | | | |
| *Palaemonetes cummingi* | Squirrel Chimney Cave Shrimp | 1 | | | | X | | | | E | |
| Caudata (Vertebrata) | | | | | | | | | | | |
| Plethodontidae | | | | | | | | | | | |
| *Haideotriton wallacei* | Georgia Blind Salamander | 12 | | X | | | | | | R | |

In Eve L. Kuniansky, editor, 2001, U.S. Geological Survey Karst Interest Group Proceedings, Water-Resources Investigations Report 01-4011, p. 78-88

**Table 1.** (continued)

| | | sites | Hydrologic Subregion 1 | 2 | 3 | 4 | 6 | 7 | 8 | Conservation Status FCREPA | AFS |
|---|---|---|---|---|---|---|---|---|---|---|---|
| **SPRINGS** | | | | | | | | | | | |
| Gastropoda | | | | | | | | | | | |
| Hydrobiidae | | | | | | | | | | | |
| *Aphaostracon asthenes* | Blue Spring Hydrobe | 1 | | | | | X | | | T | |
| *Aphaostracon chalarogyrus* | Freemouth Hydrobe | 1 | | | | | X | | | E | |
| *Aphaostracon hypohyalinum* | Suwannee Hydrobe | 7 | | | | X | | | | | |
| *Aphaostracon monas* | Wekiwa Hydrobe | 1 | | | | | X | | | T | |
| *Aphaostracon pycnum* | Dense Hydrobe | 1 | | | | | X | | | SSC | |
| *Aphaostracon theiocrenetum* | Clifton Spring Hydrobe | 1 | | | | | X | | | T | |
| *Aphaostracon xynoelictum* | Fenney Spring Hydrobe | 1 | | | | | | X | | SSC | |
| *Cincinnatia helicogyra* | Crystal Siltsnail | 1 | | | | | | X | | | |
| *Cincinnatia mica* | Ichetucknee Siltsnail | 1 | | | | X | | | | SSC | |
| *Cincinnatia monroensis* | Enterprise Siltsnail | 1 | | | | | X | | | E | |
| *Cincinnatia parva* | Pygmy Siltsnail | 1 | | | | | X | | | | |
| *Cincinnatia petrifons* | Rock Springs Siltsnail | 1 | | | | | X | | | | |
| *Cincinnatia ponderosa* | Ponderous Siltsnail | 1 | | | | | X | | | | |
| *Cincinnatia vanhyningi* | Seminole Siltsnail | 1 | | | | | X | | | | |
| *Cincinnatia wekiwae* | Wekiwa Siltsnail | 1 | | | | | X | | | | |
| *Spilochlamys conica*** | Conical Siltsnail | >20 | | | X | X | | | | | |
| *Spilochlamys gravis*** | Armored Siltsnail | >20 | | | | | X | | | | |
| Pleuroceridae | | | | | | | | | | | |
| *Elimia vanhyningiana* | Goblin Elimia | 3 | | | | | X | | | | |
| **SEEPS** | | | | | | | | | | | |
| Odonata | | | | | | | | | | | |
| Cordulegastridae | | | | | | | | | | | |
| *Cordulegaster obliqua fasciata* | Arrowhead Spiketail | — | X | X | X | X | | | | R | |
| *Cordulegaster sayi* | Say's Spiketail | — | X | X | | X | X | | | T | |
| Corduliidae | | — | | | | | | | | | |
| *Somatochlora provocans* | Treetop Emerald | — | | X | X | | | | | T | |
| Gomphidae | | — | | | | | | | | | |
| *Dromogomphus armatus* | Southeastern Spinyleg | — | X | X | X | X | X | X | | R | |
| *Progomphus bellei* | Belle's Sanddragon | — | X | X | X | | | | | R | |
| Petaluridae | | — | | | | | | | | | |
| *Tachopteryx thoreyi* | Gray Petaltail | 8 | X | ? | X | X | | | | R | |

*occurs in caves and springs
**spring preference, not obligate

In Eve L. Kuniansky, editor, 2001, U.S. Geological Survey Karst Interest Group Proceedings, Water-Resources Investigations Report 01-4011, p. 78-88

**Table 2.** Benthic macroinvertebrate diversity from spring-influenced reaches of the Suwannee River, Florida, reported by Mattson and others (1995). Some species represent taxonomic complexes, are undescribed, or were not identified below genus level. Unidentified species represent those unassigned below family level. Parenthetical numbers following select insect orders indicate numbers of families represented.

| Phylum | Subphylum/Class/Subclass/Order | Genera | Species | Unident. |
|---|---|---|---|---|
| Cnidaria | Hydrozoa | 1 | 1 | |
| Platyhelminthes | — | 1 | 1 | |
| Nemertea | — | 1 | 1 | |
| Aschelminthes | Nematoda | | | 1 |
| Mollusca | Gastropoda | 16 | 25 | 1 |
| | Bivalvia | 6 | 9 | |
| Annelida | Oligochaeta | 16 | 24 | 3 |
| | Hirudinea | 4 | 9 | 1 |
| Arthropoda | Chelicerata—Acarina | 14 | 14 | |
| | Chelicerata—Araneae | 1 | 2 | |
| | Crustacea—Amphipoda | 2 | 2 | |
| | Crustacea—Isopoda | 2 | 2 | |
| | Crustacea—Decapoda | 2 | 4 | |
| | Insecta—Collembola | 4 | 4 | 1 |
| | Insecta—Coleoptera (11) | 45 | 73 | |
| | Insecta—Diptera (10) | 89 | 133 | |
| | Insecta—Ephemeroptera (10) | 19 | 36 | |
| | Insecta—Hemiptera | 19 | 29 | |
| | Insecta—Lepidoptera | 1 | 1 | |
| | Insecta—Megaloptera | 3 | 3 | |
| | Insecta—Odonata (suborder Zygoptera) | 7 | 9 | |
| | Insecta—Odonata (suborder Anisoptera) | 19 | 29 | |
| | Insecta—Orthoptera | 2 | 2 | |
| | Insecta—Plecoptera | 8 | 12 | |
| | Insecta—Trichoptera (10) | 23 | 33 | |
| Bryozoa | — | 1 | 1 | |

# Preliminary data on microcrustacean communities from ground waters in the southern Everglades

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## ABSTRACT

We studied ground-water crustacean communities by collecting samples from three sets of wells located in the Rocky Glades area of Everglades National Park (ENP), the Atlantic Coastal Ridge northeast of ENP, and along canals southeast of ENP. In each well, we sampled at different depths, monthly from June to December 2000. Copepods collected included both calanoid species, 22 of the 27 species of cyclopoids, and 2 of the 8 species of harpacticoids reported from Everglades National Park. We also collected several isopods, amphipods, and ostracods that were identified at the generic level. The numbers of copepod species and individuals varied with time, and with depth class, suggesting that surface water copepods use ground water as a refuge from predators, or as a feeding site during the wet season. Surface-dwelling taxa become more numerous when the ground-water level decreased, at the end of the wet season. We documented faunal exchange from surface to ground waters and vice versa, and for the wells along the Coastal Ridge canals, movements to and from the canals into ground water.

## INTRODUCTION

The ground-water realm has been included in freshwater ecological studies only recently, and the knowledge of its function and structure are still very incomplete (Danielopol, 1989). Ground-water ecology is a field of study that has developed slowly in the past, mainly due to methodological difficulties related to the sampling and observation of the subsurface environment. Although ground water is known to be a complex living ecosystem, faunistic data have often been neglected. Many authors have demonstrated that surface and subterranean fauna colonize the different habitats of the highly structured, subsurface environment according to their ecological tolerance and preferences (Danielopol, 1991a; Rouch, 1991; Dole-Olivier and Marmonier, 1992). The faunal assemblages also reflect certain hydrological, geomorphological, physical and chemical processes (Amors and Mathiaeu, 1984; Creuzé de Chatellierers and Reygobellet, 1990; Lafont and Durbec 1990). Therefore, faunal assemblages should be used to monitor environmental changes (Danielopol, 1991b; Gibert, 1992).

The aquatic stygobitic fauna differs in composition from the surface-water fauna, owing to the absence of insects and to the dominance of crustacean species. In Europe, stygobitic species account for approximately 40% of the total crustacean fauna (Danielopol, 2000). At a global scale, 41% of crustacean orders have stygobiont representatives in ground waters, and 6 of the orders are essentially stygobionts (Mystacocarida, Gelyelloida, Syncarida, Mictacea, Thermosbenacea, Nectiopoda) (Stoch, 1995). For other crustacean groups such as Cyclopoida, Harpacticoida, Amphipoda, Isopoda and partly Ostracoda, the species richness in ground-water is near, or even higher than those recorded at comparable scales in surface freshwater habitats (Danielopol, 2000). There is also a large number of stygobitic relict species, which have became extinct in surface-water environments (Rouch and Danielopol, 1997).

The subclass Copepoda includes 10 orders (Huys and Boxshall, 1991), with over 11,500 known species (Humes, 1994). Only four orders (Cyclopoida, Calanoida, Harpacticoida, Gelyelloida) have free-living freshwater species. Harpacticoida are usually benthonic and rarely found in plankton. The order includes 53 families, but only Ameiridae, Canthocamptidae and Parastenocarididae are widely represented in freshwater habitats with about 1,000 species and subspecies (Dole-Olivier et al, 2000). The order Cyclopoida includes 12 families and is primarily marine epibenthic. It has secondarily invaded freshwater, mostly with the family Cyclopidae (Huys and Boxshall 1991), with 900 freshwater species and subspecies, mostly from the subfamilies Cyclopinae and Eucyclopinae. Calanoida are planktonic, occasionally found in benthic habitats of subterranean lakes or in karstic springs and within the hyporheos

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after floods (Dole-Olivier et al, 2000). Gelyelloida are represented by only two ground-water species in European karstic systems and from an undescribed interstitial stream habitat in South Carolina (Reid, pers. comm.). With more than 800 species/subspecies, a number surely underestimated, copepods inhabit all kinds of aquifers (karstic, fissured, porous), as well as surface/ground-water ecotones (land/water, water/water).

In general, crustaceans are extremely successful in colonizing ground water and in certain areas, hypogean species may be equal or exceed the number of epigean ones (Stoch, 1995). For example, detailed studies on the ground-water habitats in northeastern Italy and neighboring Slovenia, including the extremely developed karstic system showed that for copepods, 47.14% of the collected species were stygobiont. Nonetheless, the faunistic and taxonomic knowledge of ground-water organisms is rather scarce and species richness of stygobiont copepods is highly underestimated in several geographical areas. Moreover, there are usually few data on non-stygobites (i.e., stygophile: epigean organisms that occur in both surface water and ground-water without adaptation to subterranean life, stygoxene: typical epigean organisms that appear rarely, and mostly at random, in ground-water), organisms that are also needed to formulate a correct theory of hypogean species diversity. The biogeographical statement that stygobiont species are rare in tropical caves (Mitchell, 1969: Sbordoni, 1994) might be true for the Florida karstic system, but data are not in agreement with this statement (Rouch and Danielopol, 1987) and the knowledge of both stygobites and non-stygobites component of the fauna must be extended in order to support or reject these views. It must also be mentioned here that the species richness reported for a territory is related to the sampling effort. A comparison of the benthic copepod fauna shows that fewer than half as many species are known from North America as from Europe (Reid, 1992a), and the discrepancy is not due to a poorer fauna in North America, but to the lack in taxonomic studies (Reid, 1992a).

Several studies (see Rouch and Danielopol, 1997) show that extrapolating methods fail in determining species richness in ground water. A strong sampling effort is needed, together with a good knowledge of ecological/physical background data (Rouch and Danielopol, 1997). Moreover, within a given subterranean area, species richness of a taxocoenosys depends on the structure of the habitat and the functioning of the ecosystem within which these habitats are located: higher diversity can be expected in habitats that receive high quantities of energy and matter, e.g., the superficial aquifers close to surface water. Deep ground-water habitats (the “phreatic” ones) with their slower dynamics might have lower species richness (Rouch and Danielopol, 1997).

In Everglades National Park (ENP), the marl prairies of the Rocky Glades host plant and animal species adapted to variable hydroperiods, which has allowed some of them to survive the drainage of this region during the intensive development by agricultural and urban interests. Marl prairies are typically formed on a limestone karst substrate that provides a vertical dimension of habitat for aquatic organisms. In the Rocky Glades, the presence of solution holes and below-ground, near-surface habitat allows for survival of a greater variety of organisms. This is related to small differences among the different solution holes and because surface-dwelling aquatic organisms find refuge in the solution holes during the dry season (Loftus *et al*., 1992). The marl soils also retain some water and allow plants and animals to survive the drought. As a consequence, marl prairies still contain a very diverse community of plants and animals, some of which are endemic. The assessment of species composition in the marl prairies, and the following monitoring of changes in the communities appears to be of primary importance; the stress imposed over these habitats brings the risk of losing species, and of pushing to the point of extinction even organisms that can withstand large variations in physical habitat.
The transition area between the marshes and the Atlantic Coastal Ridge is also important to subterranean organisms, and this is one of the few areas where a gradient of surface/ground-water habitats is relatively intact. The surface-ground-water connections and interchange are much greater in this area than most geographic areas (based on faunal similarities), and the study of ground-water communities might give support to the necessity of protecting these areas in order to preserve ground water supplies.

The study of ground-water organisms in ENP is providing interesting preliminary results. Because it is the first time that ground waters in South Florida have been investigated for microcrustaceans, any collection is important in order to assess the species composition. Faunistic lists for ground-water organisms must be as complete as possible, and the ecology, biology and life cycle of each taxon has to be inspected and evaluated, for the purposes addressed above.

Ground-water organisms are comparatively protected from major climatic changes, and hypogean animals may persist in ground-water refugia; for the Everglades surface organisms, the possibility of entering the ground-water system can help them to survive the droughts. The identification of such surface organisms in ground water could therefore indicate a hydraulic connection between surface waters and the aquifer.

Monitoring the changes that might occur in the composition of ground-water communities can therefore be a useful tool in assessing changes in hydrology and water quality, particularly in ENP where major changes will be expected when the restoration projects are implemented.

## MATERIALS AND METHODS

We selected a set of wells in each of the following areas (Fig. 1):

- Rocky Glades (13 wells)
- L31W-C111 canals, south-east of ENP (6 wells)
- Atlantic Coastal ridge northeast of ENP (15 wells)

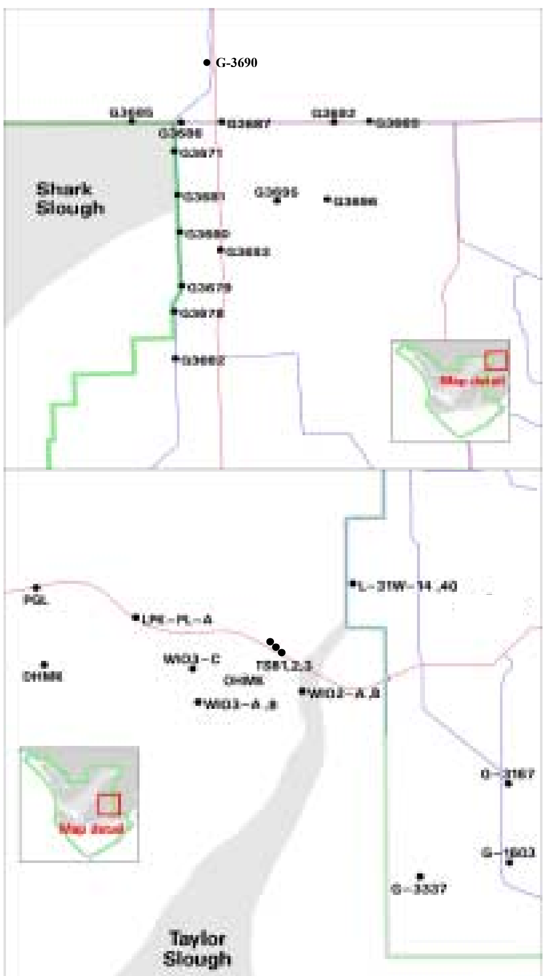


Figure 1 – Maps of sampling wells in the Atlantic Coastal Ridge Area (top), and in Rocky Glades/southeast of ENP (bottom)

Samples were collected monthly and, for the Atlantic Coastal ridge wells, cores and optical logs previously taken by USGS allowed sampling of different depths in each well, corresponding to highly permeable strata. For the wells in the Park area, we sampled on multiple depths, in order to reach high permeability layers.

A total of 91 samples have been collected monthly, starting in June 2000. In the Rocky Glades area, well PGL was flooded from June to October, OHMK and DHMK were flooded from July to September, LPK-PL-A and LPK-PL-B were flooded from August to September, and therefore all these wells were not sampled during these intervals. December samples form Atlantic Coastal Ridge wells were collected after a rainstorm that raised the water levels of about 10 cm.

Qualitative samples were collected using a Wayne® 1/2HP portable pump connected to a Coleman® 1750 Watt portable generator, and a set of 1.5 m long PVC pipe that were connected to the pump through a flexible plastic hose. 1,000 L of water were filtered using a 63-µm mesh, 20-cm-diameter plankton net. Water depth and environmental variables such as temperature, pH, conductivity, salinity, and dissolved oxygen were recorded using a YSI 85® Multi-Parameter Water Quality Meter.

All samples were fixed in the field with 5% buffered formalin; specimens were sorted and counted in the lab using a Leica® Stereoscope. Specimens of each taxa were mounted in permanent slides with Faure's medium and studied with a Leica® DMLS phase contrast microscope at 10X, 20X, 40X, 100X. All crustaceans were identified to at least family level, and for copepods the species, sex, and developmental stage have been determined. Results will be therefore presented only for this last group.

## RESULTS

For the taxa collected, both species of calanoids, 22 of the 27 species of cyclopoids (81.5% of the species) and 2 of the 8 species of harpacticoids (25% of the species) (Reid, 1992b; Loftus and Reid, 2000; Bruno et al., 2000; Bruno et al., in press, Bruno unpubl. data) were present in the ground-water samples. Five cyclopoid species are new records for Florida, and one is the first record for North America.

At the beginning of the wet season, in June, when the groundwater level was still low (Figs. 2, 3), we collected copepods in quite good density, and the species number had the highest value (Fig. 4). Individuals were collected only in the first 9 m depth

In Eve L. Kuniansky, editor, 2001, U.S. Geological Survey Karst Interest Group Proceedings, Water-Resources Investigations Report 01-4011, pp. 89-97

(Fig. 5), with 56.8% at 0-3 m, 32.1% at 3-6 m, and 11.1% at 6-9 m (Fig. 6).

| | 0-3 m | 3-6 m | 6-9 m | 9-12 m | > 12 m | Total | N |
|---|---|---|---|---|---|---|---|
| Jun. | 0.515 | 0.291 | 0.101 | 0.000 | 0.000 | 0.907 | 17 |
| Jul. | 0.295 | 0.243 | 0.037 | 0.015 | 0.000 | 0.590 | 11 |
| Aug. | 0.175 | 0.188 | 0.122 | 0.002 | 0.003 | 0.490 | 12 |
| Sep. | 0.081 | 0.063 | 0.020 | 0.001 | 0.000 | 0.165 | 7 |
| Oct. | 0.705 | 0.078 | 0.039 | 0.000 | 0.000 | 0.822 | 13 |
| Nov. | 2.798 | 0.449 | 0.012 | 0.001 | 0.000 | 3.260 | 13 |
| Dec. | 0.124 | 0.211 | 0.039 | 0.003 | 0.002 | 0.379 | 10 |

Table 1- Density of individuals (N ind./L) collected at each depth class by month, and number of species per month.

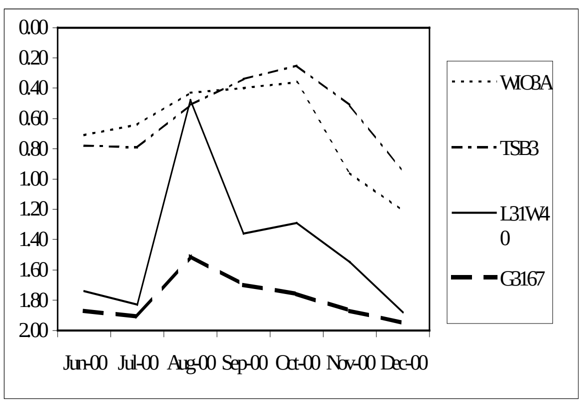


Figure 2 - Ground-water levels at selected Atlantic Coastal Ridge wells

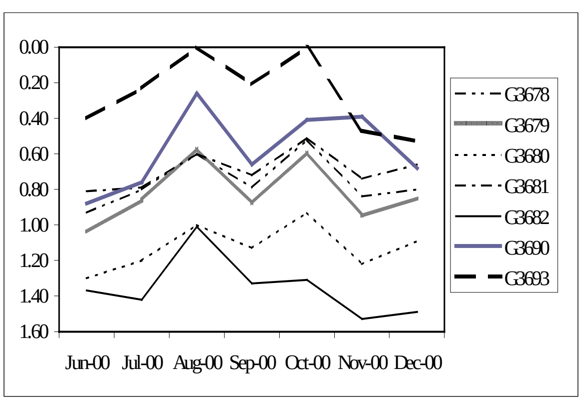


Figure 3 - Ground-water levels at selected Rocky Glades/south-east ENP wells

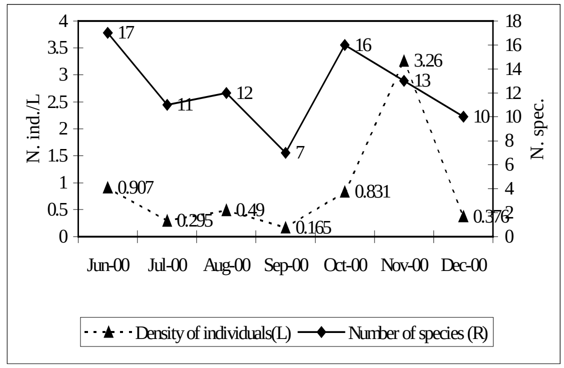


Figure 4 – Density of individuals (left) and of species (right) collected each month.

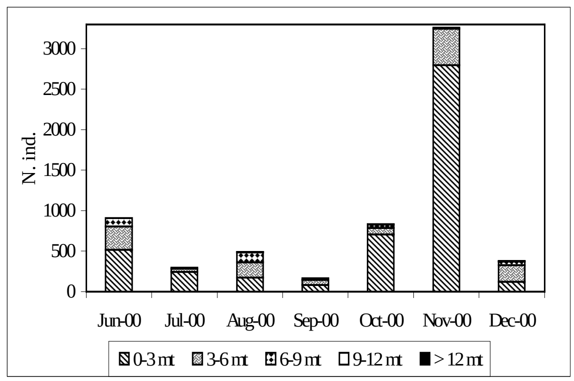


Figure 5 – Number of individuals collected at each depth class by month.

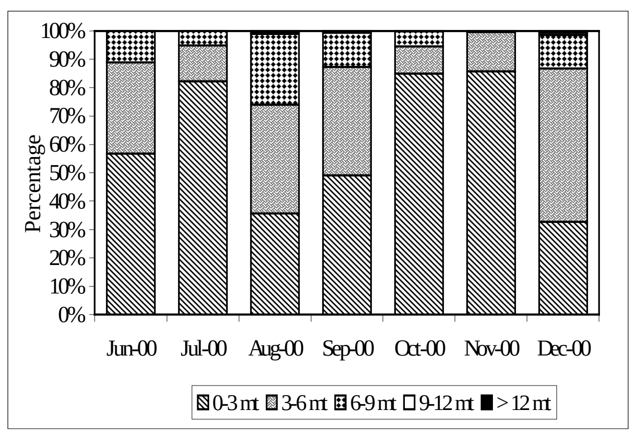


Figure 6- Percentage of individuals collected from each depth class, each month.

In August, in correspondence with high rainfall, the ground-water level was high at most sites (Figs 2,3), reaching almost the surface. In this month, the density of individuals was still low, and one more species was recorded in respect to the previous month (Fig. 4, Tab. 1). For the first time, a few individuals were collected at depths below 12 m: 35.7% of the individuals were collected at 0-3 m, 38.4% at 3-6 m, 24.9% at 6-9 m, 0.4% at 9-12 m, and 0.6% at below 12 m (Figs. 5,6).

In September, ground-water level generally dropped (Figs. 2,3); both number of species and density of individuals had the lowest value (Figs. 4,5; Tab. 1). All individuals were collected at depths above 12 m, in particular: 49.1% at 0-3 m, 38.2% at 3-6 m, 12.1% at 6-9 m, and 0.6% at below 9 m (Fig. 6).

In October ground-water level rose again, due to extremely high rainfall (Fig. 2,3) and we recorded the highest species number (Fig. 4; Tab. 1) and densities of individuals since June. All specimens were collected at depths above 9 m, with 85.8% at 0-3 m, 9.5% at 3-6 m, and 4.7% at 6-9 m.

In November, at the end of the wet season, ground-water level dropped at most stations (Figs. 2,3), and the density of individuals was the highest recorded, more than three times higher than in October, but with less species (Fig. 4); most individuals were collected at shallow depths: 85.8% at 0-3 m, 13.8% at 3-6 m, and only 0.4% at 6-9 m (Figs, 5,6).

In December, when water level dropped (Fig. 4,5, data for Atlantic Coastal ridge wells were collected after a rainstorm (see materials and methods), and densities of individuals and number of species were low (Fig. 4; Tab. 1); 32.7% of the individuals occurred at 0-3 m, 55.7% at 3-6 m, 10.3% at 6-9 m, 0.8% at 9-12 m, and 0.5% at below 12 m (Figs. 5,6).

Regarding the ecology of the collected organisms, calanoids are planktonic species, therefore strictly epigean, and can be considered as stigoxene. Of the two species, *Arctodiaptumus floridanus* (Marsh 1926) was collected only in July with 83 individuals. The second species, *Osphranticum labronectum* (S.A. Forbes 1882), was collected through the entire sampling period with very few individuals until November, when we took 1,719 individuals. This species is usually very abundant in surface waters of ENP in the same time of the year (Bruno et al., submitted.).

Two species of cyclopoids, *Thermocyclops parvus* (Reid 1989) and *Diacyclops nearcticus* (Kiefer 1934), are associated with hypogean habitats and are stygophiles. Both had already been collected in ENP but mostly from ground water. A third species, a *Diacyclops* sp. very similar to *D. jeanneli* (Chappuis 1929), has been collected with only one specimen, therefore the taxonomic classification is still unsure, however *D. jeanneli* has only been collected in caves in Indiana. If *Diacyclops* sp. will is classified as *D. jeanneli*, it will be a new record for Florida of this stygobitic species. All other species of cyclopoids collected are epigean, but very euryoecious and tolerant of different physical-chemical conditions.

Of the two harpacticoids, *Phyllognathopus viguieri* (Maupas 1892) is one of the most widely distributed. It has been collected from virtually every possible freshwater habitat, ground water included (Dussart and Defaye, 1990). *Elaphoidella marjoryae* (Bruno and Reid 2000) had been previously collected both from ground waters and in solution holes during the dry season, when the deep solution holes that do not dry are connected and recharged by ground water (Bruno et al., 2000). This is therefore another species strictly related to ground-water habitats, and can probably be considered as stygoxene.

The trend in species richness and species composition is interesting, since the stygobite and stygophile species become very rare in September October. For example, we collected 38 specimens of *Diacyclops nearcticus* in August, 11 in September, and 10 in October, however, for this last month, this species was not present at all in the Rocky Glades wells, where it previously always occurred over the last 2 years of sampling. When the water level started to decrease, the number of individuals of *D. nearcticus* increased; we collected 67 individuals in November and 47 in December, most of them from the Rocky Glades wells. Other typically planktonic species are present in October, such as all species belonging to the genus *Paracyclops* and *Microcyclops*.

Regarding crustaceans other than copepods, we collected a few specimens of amphipods, isopods, and ostracods. Just two specimens of the isopod, *Caecidiotea* sp., were collected in good condition, and are being identified; the amphipods belonged to the genus *Crangonix*. Only a few ostracods were collected and were identified as both epigean, swimming forms, as well as hypogean, stygobitic species.

## DISCUSSION

High diversity and densities of individuals would be expected in habitats that receive high inputs of energy and matter, as is the case of the superficial aquifers located closely to surface water. In the Rocky Glades area of the Everglades, a high degree of dissolution of the oolitic limestone bedrock has occurred with time, producing a typical karstic landscape with thousands of solution holes (Hoffmeister, 1974). Some of those are connected with ground water (Loftus et al. 1992). Surface organisms move to ground water to find a refuge from the drought and from predators (Loftus et al., 1992; Bruno et al., submitted.). Therefore, the presence of surface water organisms in our ground-water samples was expected. The dominance of strictly epigean taxa, such as calanoids, in ground-water samples, is indicative of surface-water intrusion in the aquifer, followed by passive dispersal of the organisms. Nonetheless, planktonic organisms are typically unable to actively move for long distances; their dimensions are often less than 1 mm, and therefore even the strongest swimmers (the largest species, such as calanoids) are limited in their dispersal abilities. We, therefore, assumed that species collected at a certain depth come from adjacent permeable layers.

Most of the specimens were collected in the USGS wells northeast of ENP, probably because we were sampling the high permeable layers. All those wells are along canals, and it is possible that planktonic organisms could enter ground water through lateral connections. In most of the ENP wells, surface organisms could be transported into the wells from the surrounding sloughs when the area is flooded during the wet season. Even if most of the copepods are not strictly epigean, they are able to survive in ground-water habitat and complete their life cycle there. In most months and wells, we collected individuals at different larval stages, suggesting that reproduction takes place in ground water for both epigean and hypogean taxa, or that both larval stages and adults can disperse in ground water and survive there.

The particular high diversity recorded in the shallow samples can be explained considering that the surface-water/ground-water interface sites are where intense hydraulic exchanges occur and the biogeochemical activity is higher than in the adjacent systems (Gibert et al., 1997). It has been assessed that the main characteristics of these interfaces are their great variety of elasticity, permeability, biodiversity and connectivity (Gibert et al., 1990). The marl prairies of the Rocky Glades have a high permeability (Fish and Stewart, 1992), allowing rapid surface/ground-water exchange. This happens particularly during the wet season, when ground waters are recharged by rainfall; surface organisms probably then disperse passively in ground water.

The wells may be divided in three groups: Atlantic Coastal Ridge, L-31W/C-111, and Rocky Glades. The first group always had the highest number of individuals, underlining the importance of using GPR and video cameras to locate permeable layers when ground-water colonization is studied in karstic habitats. In L-31W and C-111 wells, only few individuals were sporadically collected, even though these wells were located along canals, as in the previous group. We assume that the sampling depth did not correspond to highly permeable layers.

The Rocky Glades wells are not located along canals, but are in or very near areas that are flooded during the wet season. The exception are the TSB wells, that are on the main ENP road shoulder in the section crossing Taylor Slough, from which copepods might enter into ground water. The entire area has high permeability, with a transmissivity of about 300 $ft^2$/day (Fish and Stewart, 1992). Those wells had high numbers of individuals, with wide seasonal variations, and characteristic high numbers of calanoids, even when surface water had disappeared (e.g. PGL in November). Therefore, in this area surface copepods disperse into ground water and survive for long time, following the recession of the water table. The faunal assemblage of these wells was constituted by stygoxenes and by stygophiles, particularly *D. nearcticus, T. parvus,* and *Orthocyclops modestus* (Herrick 1883).

Our data may confirm the presence of seepage from the canals to the ground waters, at least for the wells along the Atlantic Coastal Ridge. These are mostly located on the sides of L-31N, L29, L30, and C-4 canals, and they are where most of the specimens were collected. For the Rocky Glades area, organisms may migrate vertically from surface waters, and when surface water disappears, they remain in ground-waters, since here they can not move back into the canals, as may be possible for organisms collected in the Atlantic Coastal Ridge wells. This seemed to happen particularly at the end of the wet season; in November, when the water table fell to deeper levels, calanoids survived by entering ground water, demonstrated by the high numbers (1,687) of individuals collected at PGL, a station that had been flooded until the previous month.

To our knowledge, faunistic exchanges in a karstic system such as the Everglades have never been studied. Nonetheless, data for other subterranean habitats raise the possibility of surface/ground-water exchanges, in relation to variation of ground-water levels. For example, Malard et al., 1994, found that ground-water communities were dominated by epigean organisms during low water periods when infiltration rates of surface waters were high. During periods of intensive ground-water recharge, these epigean organisms were displaced upstream and downstream of the study area and were also disseminated throughout the adjacent fissure network. They were associated with hypogean organisms, which were collected only when ground waters circulated though the opening of the site. Therefore, a close relationship existed between the ground-water flow, the movement of surface waters towards and within the aquifer, and the spatial and temporal distribution of hypogean and epigean species. Data from other ground-water habitats confirm this information. For example, Marmonier and Creuzé des Châtelliers (1991) studied the dynamics of interstitial assemblages after a spate and during low discharge, in a regulated channel of the Upper Rhône River. They found that most stygobites decreased in abundance or disappeared just after a spate while, during low discharge, stygobites were more abundant and diversified.

Ground-water ecosystems in the Everglades are diverse, and this diversity should be protected or even increased by ecological remediation procedures. As already stated by Rouch and Danielopol, 1997, species richness in subterranean habitats depends not only on the diversity of available habitats, but also on the global functioning of the ecosystems to which these habitats belong. Particularly important are the relationship between surface and subsurface systems. Therefore, in order to protect the subterranean fauna, the surrounding surface environment should be protected, since they are a component of the surface-subsurface hydrological exchanges.

In the Everglades, protection of the ground-water ecosystem will be achieved when environmental protection activities are finally extended to the entire system, to which this unique karstic habitat structurally and functionally belongs.

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# Geochemistry of the Upper Knox Aquifer in Tennessee

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## Abstract

The Knox Group, a thick sequence of Cambrian- and Ordovician-age limestones and carbonate rocks, occurs at land surface in East Tennessee and primarily in the subsurface in Middle and West Tennessee. West of the Valley and Ridge Physiographic Province, the Knox Group occurs in the subsurface and crops out in two locations, the Sequatchie Valley and the Wells Creek cryptoexplosive structure. In Middle Tennessee, the Knox Group occurs at depths ranging from about 500 feet below land surface to several thousand feet below land surface in areas outside the Central Basin. The carbonate rocks of the Knox Group were subjected to subareal dissolution, karstification, and structural deformation. In the upper 300 to 500 feet of the Mascot Dolomite of the Knox Group, a paleokarst zone developed that forms the upper Knox aquifer.

Ground water in the upper Knox aquifer occurs in a series of solution openings, fractures, and brecciated zones with varying geochemical composition. The upper Knox aquifer receives recharge from the overlying Ordovician-age carbonates in the Central Basin of Tennessee. In the Recharge area, ground water in the upper Knox aquifer is primarily a calcium sulfate bicarbonate water type with generally less than 1,000 milligrams per liter dissolved solids. Ground-water flow in the upper Knox aquifer is primarily to the southwest; discharge from the aquifer occurs along the western valley of the Tennessee River near the structural high of the Clifton Saddle. Downgradient, toward the apparent discharge areas, ground water in the upper Knox aquifer is a sodium bicarbonate chloride water type with dissolved-solids concentrations of about 1,000 to 1,500 milligrams per liter. East of the recharge area, the ground water becomes a sodium chloride water type with 8,000 to greater than 10,000 milligrams per liter dissolved-solids concentrations. Northwest of the Tennessee River, ground-water in the upper Knox aquifer is primarily a sodium chloride water type with about 5,000 milligrams per liter dissolved solids.

In Eve L. Kuniansky, editor, 2001, U.S. Geological Survey Karst Interest Group Proceedings, Water-Resources Investigations Report 01-4011, p. 98

# Inorganic Carbon Flux and Aquifer Evolution in the South Central Kentucky Karst

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## Abstract

The geometries of carbonate karst aquifer "plumbing systems", and the rates at which they evolve, depend on the fluxes of water and $CO_2$ through them. A one-year, high temporal resolution study of flow and carbonate chemistry within the humid-subtropical Mammoth Cave System's Logsdon River quantifies significant variations in these fluxes over storm and seasonal time scales. Undersaturated storm waters dissolve rock within a 25-30 m thick flood zone. Waters were only undersaturated, and thus capable of dissolving the aquifer framework, 31% of the year. Rates of aquifer evolution are thus strongly influenced by time-varying processes. Although flood conditions occur during a small percentage of the time, they dominate chemical passage enlargement. Of the $7.8 \times 10^3 \pm 1.9 \times 10^3$ kg ha$^{-1}$ of total inorganic carbon leaving the river's 25 km$^2$ catchment during the year, 1% entered the aquifer as recharge, 57% was derived from carbonate mineral dissolution, and 42% was produced by biological activities. Comparison of sources of inorganic carbon suggests that during large floods the aquifer is "rinsed" of the diffuse limestone dissolution products that accumulate during more normal conditions by waters moving downwards through the vadose zone. A dual approach, coupling quantitative modeling with calibration and refinement of the models by careful measurement of processes within real karst aquifers, provides a framework for developing a comprehensive understanding of karst system behavior.

## Introduction

Because of the low flow resistance typical of well-developed karst aquifers, significant changes in both flow and groundwater chemistry, and associated water-rock interactions, occur over a variety of timescales. Under some conditions these changes can be rapid. Since aquifer dissolution rates, in turn, depend on the spatial and temporal variations in the chemistry of the through-flowing waters, investigating the nature of these changes is necessary to understand karst aquifer and landscape development. Using a new method that provides high frequency quantification of important carbonate water chemistry parameters, this paper reports on the results of a one year sampling program (May 5, 1995 to May 4, 1996) designed to understand rates and magnitudes of water and carbon dioxide fluxes within one of the major underground rivers of the humid-subtropical south-central Kentucky karst aquifer. Using this information we evaluate how these changes, at time scales ranging from minutes to months, impact conduit evolution and inorganic carbon transport through the aquifer. This work is being undertaken in cooperation with the International Geological Correlation Program, Project 379: " Karst Processes and the Global Carbon Cycle."

## Field Site

Logsdon River is one of the major underground streams of south-central Kentucky's Mammoth Cave karst aquifer (Figure 1). Within the Proctor Cave section of the system, near the south-western known edge of the Mammoth Cave System, Logsdon River converges with Hawkins River, which drains about 75 km$^2$ near Park City, Kentucky. Logsdon River drains the 25 km$^2$ Cave City groundwater basin. Within Logsdon River, about 100 m upstream from the confluence of the two rivers, are two 15-cm diameter, 145-m deep observation wells intersecting the river (Figure 1). One well is equipped with a compressor-driven pump for collecting water samples from the surface, and through the other electronic probes deliver high-resolution (two-minute) data on stage, velocity, temperature, and specific conductance. The bottoms of the wells can be reached underground (for pump and probe installation and maintenance) through the Doyel Valley Entrance to Mammoth Cave, requiring a 3-km round-trip through the cave. The conduit at the Logsdon River well site is about 3 m tall and 7 m wide, and lies near the downstream end of one of the world's longest known sections of accessible underground river passage, continuously traversable for over 8 km.

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Figure 1. Field site at Logsdon River, the main trunk flow path of the Cave City Basin of the Mammoth Cave karst aquifer. Another well just out of the picture to the left of the passage delivers probes that record flow and chemical conditions of the river with two-minute resolution. Photos by Chris Groves.

## Methods

The primary objective of the research has been to collect sufficient high-resolution data on both flow and chemical characteristics of the river to quantitatively evaluate the magnitudes and rates of change of carbonate chemistry and water/rock interactions at a variety of time scales. Four probes are installed through the wells into each stream, including temperature, specific conductance, river stage, and velocity. A Campbell CR10 multi-channel data logger queries the probes every thirty seconds, and averages these readings every two minutes. Using an accurately surveyed passage cross-section at the well site, stage data were used to calculate flow cross-sectional area for discharge calculations.

With data resolution varying from two minutes during storms to hourly during static conditions (in the interest of both data storage and computational efficiency), these values were determined for 21,473 observations between May 5, 1995 to May 4, 1996. Equipment problems led to a loss of 19.9 days (5.4% of the period) of conductance data during the year. With the large number of conductance chemographs during storms available over the rest of our study period, however, we were able to closely estimate these missing periods using a procedure analogous to unit hydrograph analysis (Chow, 1964). In addition, electrical problems created offsets of the data (29.2 days, or 8.0%) that were corrected by calibration and translation. We used direct velocity measurements (with two minute resolution) to estimate discharge under open channel conditions, but these were limited by measurement at a single location in the center of the flow cross-section in each stream. To account for the vertical velocity distribution the velocity at 0.6 flow depth, which has been shown to be close to the mean velocity in the vertical profile, was related to the apparent velocity at the fixed-depth point of measurement assuming a logarithmic velocity distribution (Chow, 1959). In order to correct for longitudinal velocity variations, discharge measurements were made on eight occasions using standard wading-rod gaging techniques and a 0.3 meter longitudinal spacing (Chow, 1959), and the average ratio between mean and maximum velocity (assumed at the center of each channel, where the velocity sonde is located) was determined. The vertically corrected apparent velocity in each stream was then multiplied by this ratio (0.57).

Hourly water chemistry measurements have been made during storms of various magnitudes by sampling from the surface through the wells. Calcium has been measured with either titration or with a Varian Spectra 10/20 atomic absorption spectrophotometer (AA) at the Ogden Environmental Water Lab at Western Kentucky University, and magnesium has been measured exclusively with atomic absorption. Samples for dissolved calcium and magnesium AA analysis were filtered (0.45 μm) and acidified. Conductivity and pH were measured immediately after sample collection. A gel filled combination electrode using two-point calibration was used to measure pH. Bicarbonate alkalinity was measured by acid titration either immediately after collection, or in some cases samples were stored on ice, and the titrations were done at the Ogden Laboratory within 24 hours.

## Analysis

With this information we have used least-squares regression analysis (Draper and Smith, 1981) to relate high-resolution specific conductivity data from within the river to calcium, magnesium, and bicarbonate concentrations, as well as pH (Figure 2). These relationships, along with direct temperature measurement, then allow calculation of several important components of carbonate system behavior with high temporal resolution, including $CO_2$ pressures, dissolution rates, and total inorganic carbon (*TIC*) fluxes through the aquifer contributed from atmospheric, mineral, and biological sources.

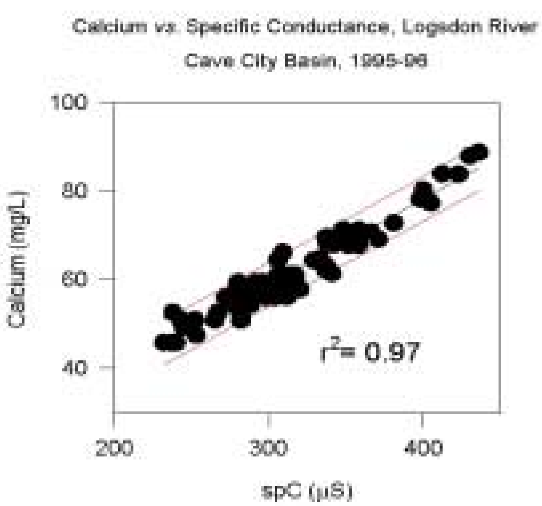


Figure 2. Statistical relationship between calcium and continuously measured specific conductance (spC) used in carbonate chemistry calculations.

Activities of $Ca^{2+}$, $Mg^{2+}$, and $HCO_3^-$ are first obtained using the extended Debye-Hückel limiting law (Stumm and Morgan, 1981). Then activities of $CO_3^{2-}$ and $H_2CO_3^*$ are calculated using appropriate equilibria, where

$$[H_2CO_3^*] = \frac{[H^+][HCO_3^-]}{K_1} \qquad (1)$$

and

$$[CO_3^{2-}] = \frac{K_2[HCO_3^-]}{[H^+]}., \qquad (2)$$

with activities indicated by square brackets. These are then summed with bicarbonate activities to obtain *TIC* for each observation, where

$$TIC = [CO_3^{2-}] + [HCO_3^-] + [H_2CO_3^*], \qquad (3)$$

where $H_2CO_3^*$ is the sum of $H_2CO_3^o$ and aqueous $CO_2$ (Stumm and Morgan, 1981).

Calcite dissolution rates assumed reaction-limited kinetics and were calculated for each observation using the rate law of Plummer *et al.* (1978), where

$$Rate = k_1[H^+] + k_2[H_2CO_3*] + k_3[H_2O] - k_4[Ca_2^+][HCO_3^-], \qquad (4)$$

with *Rate* expressed in mass of mineral lost per time per surface area of fluid/mineral contact, and where the *k*'s are temperature dependent kinetic rate constants (Plummer *et al.*, 1978). For ease of interpretation, we report results in this paper in the rate of conduit wall retreat (mm/yr) following the example of Palmer (1991).

We also investigate aquifer hydrochemical/ evolution characteristics using a recently developed method to discriminate between the various sources of inorganic carbon cycling through karst aquifers (Groves and Meiman, 2000). In this analysis we assume that there are three classes of sources for total inorganic carbon, *TIC*, leaving a limestone or dolomite karst groundwater basin:

a. dissolved inorganic carbon in precipitation, in equilibrium with the atmospheric background $CO_2$ pressure, entering the aquifer as recharge, $C_a$,

b. biological processes that produce gaseous $CO_2$ in the soil, vadose, and saturated zones, including microbial respiration, oxidation of organic material, and plant root respiration, $C_b$, and

c. carbonate mineral dissolution, $C_m$ .

Thus over a specified time interval t ,

$$\int_0^t TICdt = \int_0^t C_a dt + \int_0^t C_b dt + \int_0^t C_m dt. \quad (5)$$

Since $C_a$ and $C_b$ originate in the atmosphere as $CO_2$ gas, and leave the karst system as dissolved inorganic carbon species, we define the inorganic component of the $CO_2$ sink, $\xi$ due to flow and biogeochemical processes within the karst landscape/aquifer system:

$$\xi = (\int_0^t C_a dt + \int_0^t C_b dt) / tA \quad (6)$$

where $t$ is the length of the sampling period, and $A$ is drainage basin area.

$C_a$ is estimated by calculating a carbon mass flux leaving the basin, using measured river discharge and assuming *TIC* in equilibrium with a constant representative atmospheric $CO_2$ pressure, nominally set at 360 parts per million over the study period, using (3). In the current study, input data with two-minute resolution of chemical and flow conditions within the river are able to encompass all significant features of storm-scale and seasonal variations, so that close estimates of the appropriate integrals can be obtained.

To obtain both $C_m$ and $C_b$, we make use of the fact that during dissolution of limestone or dolomite one mole of C is released from the mineral for each mole of $Ca^{2+} + Mg^{2+}$, regardless of the elementary reaction involved (Plummer *et al.*, 1978; Busenberg and Plummer, 1982), so that

$$\int_0^t C_m dt = \int_0^t (Ca^{2+} + Mg^{2+})\, dt, \quad (7)$$

and by substituting (7) into (5) and rearranging,

$$\int_0^t C_b\, dt = \int_0^t TICdt - \left( \int_0^t (Ca^{2+} + Mg^{2+})\, dt + \int_0^t C_a\, dt \right). \quad (8)$$

## Results

### Flow and Chemistry Variations

Fifty-six storm events caused measurable changes in the flow or chemistry within the river during the study (Figure 3), with thirteen of the storms flooding the conduit to the ceiling. During the largest storm, on May 18, 1995, the water level reached a maximum height of 28.2 m, with a maximum rate of rise of 6.03 m/hr. The ceiling of the conduit under these conditions is about 25 m below the water table, and the conduit remained totally flooded for 114 hours. Based on observed flow characteristics, the year can be divided into a dry and wet season (Figure 3). After the May 1995 flood subsided, the conduit flooded only twice over the next 224 days, and one of those events was the remnants of a hurricane (Opal, study day 141, October 5), a relatively infrequent weather condition for Kentucky. In contrast, during the last 135 days of the study, beginning on Jan 2, 1996, floodwaters filled the conduit ten times, or an average of about once every two weeks.

The chemistry of the system also showed a seasonal signature (Figure 3), with the most highly concentrated fluids during the dry season, presumably reflecting a larger proportion of diffuse flow, in contrast to more diluted storm flow, with higher concentrations of dissolved limestone. A third-order regression curve of the 21,473 dissolution rate observations shows a sine-like form, apparently reflecting a higher frequency of storms during the winter and spring which do not allow the river's waters to return to a "chemical baseflow" condition, in which river flow is completely maintained relatively concentrated diffuse flow waters. Dissolution rates, in contrast, continue to fall for several months between the infrequent rains of the dry summer period. This suggests that the full relaxation time for the waters to reach a "chemical baseflow" condition in the river is on the order of months, rather than days in the case of river stage.

The maximum dissolution rates (associated with minimum conductivities) reached as a result of storms nearly all fell within a close range, suggesting a maximum bounding value of about 1 mm/yr, associated with conductivity values of 190 µS that, for sufficiently large rainfalls, is independent of season, rainfall amount, and antecedent moisture conditions. The one exception is the large May 1995 flood, where the conductivity reached a low of 167 µS.

A striking feature is that the waters were oversaturated (negative values in Figure 3) about 69% of the year, so that dissolution took place only about one-third of the time (Groves *et al.*, 1999). Sampling upstream from this location to explore the geochemical evolution of Logsdon River (Anthony, 1998) suggests that $CO_2$ outgassing along the stream is responsible for this oversaturated condition. Our data suggest that $CO_2$ pressures at this point in the river system, in the lower end of the basin, are influenced primarily by the relative contributions of high $PCO_2$ recharge from discrete inputs and during storms and lower $PCO_2$ diffuse flow inputs. The latter have been in close, prolonged contact with limestone, and much of the gas has been consumed by the resulting carbonate mineral dissolution. This seasonality in $CO_2$ pressures, and thus dissolution rates, is more closely in phase with seasonal hydrologic changes (including the influence of both precipitation

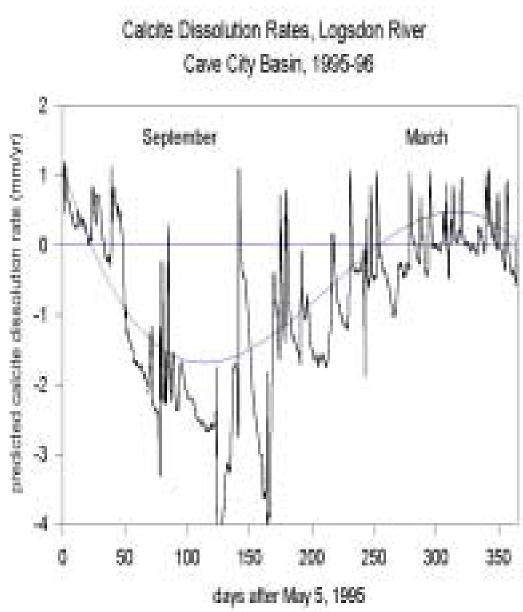


Figure 3. Predicted calcite dissolution rates showing both storm and seasonal scale variations, based on the rate law of Plummer *et al.* (1978). Negative values indicate mineral precipitation.

and evapotranspiration rates) than changes in seasonal soil $CO_2$ levels, which are in general highest in the warm seasons (Miotke, 1974).

This oversaturated condition might suggest that the dissolution occurring during storms and the wet spring season is offset by mineral precipitation in the conduit during dryer periods. In fact, the mean rate over all 21, 473 observations is -0.13 mm, indicating net *precipitation* over the year. Several observations, however, suggest that precipitation is limited. It has been reported in other settings that calcite saturation index (*SI*) values on the order of about +1 are required to exceed kinetic thresholds required for the initiation of calcite precipitation (*e.g.*, Herman and Lorah, 1988), and the maximum calculated *SI* values during the study period were less than 0.7. No macroscopic physical evidence of travertine deposition, such as rimstone dams, is present at the site. We have examined bed sediment grains collected at the end of the fall dry period (October, 1996) (Smith, *et al.*, 1997) under a scanning electron microscope, and washed sediment in acid to identify increased calcium concentrations, and have found evidence for minor calcite coatings. It is not clear, however, whether this precipitation occurred at the site, or when the grains were further upstream, where rimstone is common in some areas. Making the assumption that solutions do not become sufficiently oversaturated during the year to initiate calcite precipitation at the site (Figure 4), by setting all negative rates to zero, the mean calculated rate becomes 0.36 mm/year of dissolution. In either case, the importance of storm flows in dominating conduit enlargement is clear.

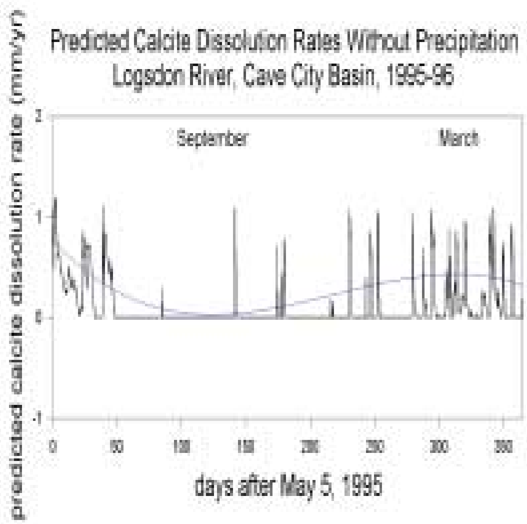


Figure 4. Predicted calcite dissolution rates based on the rate law of Plummer *et al.* (1978), assuming that kinetic barriers for precipitation are not met, and thus that the dissolution rate goes to zero during periods of oversaturation.

## Inorganic Carbon Budget and Transport Processes

Groves and Meiman (2000) measured the relative sources of the inorganic carbon flux in transport through the system over the study period discussed in this paper (eqs. (5)-(8)), and found that of the $7.8 \times 10^3 \pm 1.9 \times 10^3$ kg ha$^{-1}$ of total inorganic carbon leaving the 25 km$^2$ Cave City Basin during the year, 1% entered the aquifer as recharge, 57% was derived from carbonate mineral dissolution, and 42% was produced by biological activities. Further interpretation of those data can reveal details of aquifer transport processes. It is the flux of water and carbon dioxide through the system that ultimately controls the evolution of the aquifer framework.

The magnitudes of the total inorganic carbon flux as well as those from mineral and biological sources were summed over each of the twelve months of the study period and plotted against mean monthly discharge for the same period (Figure 5).

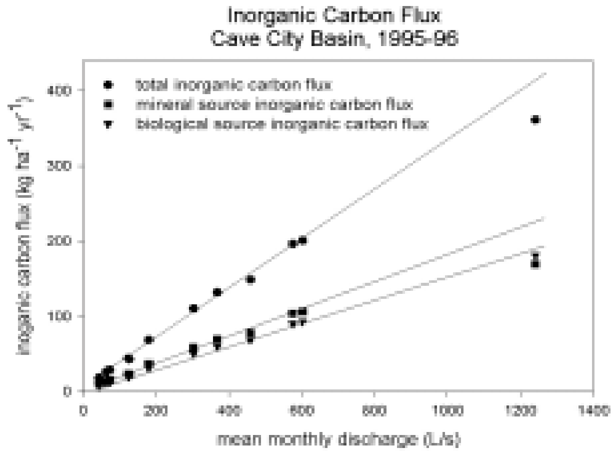


Figure 5. Magnitudes of inorganic carbon fluxes and discharge summed over monthly intervals.

In general, they show a nearly linear relationship, except for the point from May, 1995, which was dominated by a major flood, produced by roughly 25 cm of rainfall over less than one day (rightmost point on Figure 5), causing a stage increase of more than 28 m in twelve hours at the study site. In particular, the May point is in line with the biological source inorganic carbon (triangles on Figure 5), but that produced by dissolution of the carbonate bedrock falls far below the regression line for the eleven smaller flows. This suggests, first, that the inorganic carbon washed through the system produced by biological processes in the soil and vadose zone (Atkinson, 1977; Wood and Petraitis, 1984; White, 1988) occurs in an abundant enough supply that the amount transported through the system is roughly proportional to the amount of water available. In contrast, in May there is a clear deficit of mineral source carbon from that expected if the same relationship holds (Figure 6). A plausible explanation is that for most flows mineral carbon from diffuse parts of the vadose zone washes through the aquifer in proportion to the flow of water, but in sufficiently large flows this carbon is "rinsed" from the aquifer, and the water that follows becomes correspondingly diluted with respect to the products of dissolved limestone. The fact that this more dilute water should be significantly more undersaturated with respect to limestone than typical flows further highlights the dominant role that large storms appear to have on karst aquifer development.

The conclusions drawn from this carbon budgeting process do not consider the impacts of organic carbon, and thus may be influenced accordingly. We are currently beginning to investigate both organic carbon transport and microbial ecology of the Mammoth Cave karst aquifer (Elliott *et al*. 2000, and Vaughn, 1998) to better understand and quantify carbon dynamics and how these may influence aquifer evolution.

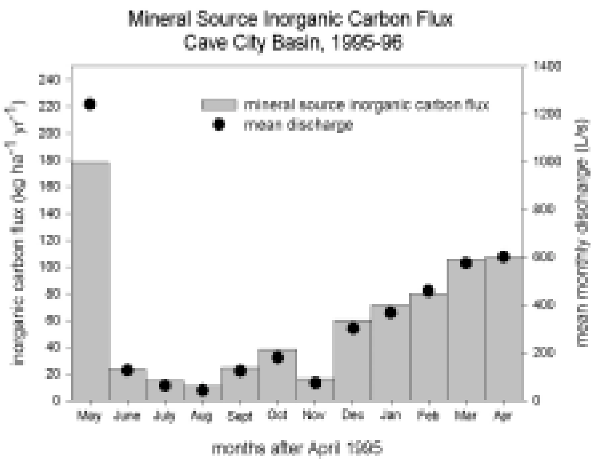


Figure 6. Mineral source organic carbon and discharge summed over monthly intervals.

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# DIRECT LINKAGES BETWEEN ONSHORE KARST AQUIFERS AND OFFSHORE MARINE ENVIRONMENTS: CRESCENT BEACH SPRING, FLORIDA

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Crescent Beach Spring is a continuously flowing offshore spring located 4 km off the northeast coast of Florida in the Atlantic Ocean. The source aquifer for Crescent Beach Spring is the Ocala Limestone (upper Eocene), which is the most productive formation of the Floridan Aquifer. In addition, Crescent Beach Spring discharges relatively fresh (9.4 milliSemens/centimeter) particulate-laden water from the Ocala at a water depth of 38 meter, creating zero visibility in the throat of the spring. The high flow rates (10 - 300 cubic feet per second) display a noticeable boil on the sea surface and a hydrogen sulfide odor is distinguishable for some distance downwind.

The likelihood of contaminating source waters and subsequently transporting contaminants from onshore aquifer systems to offshore marine environments is extremely great at Crescent Beach Spring, as well as at many other well-known offshore springs throughout Florida. In addition, ground water can transport nutrients and metals from agricultural and urban areas, not only impacting the quality of drinking water, but also influencing changes in benthic habitats where ground waters discharge into overlying marine waters.

This study has been initiated to link offshore hydrostratigraphy, implementing seismic profiling and geochemical signatures of the discharge waters, with onshore sources. In April 1999, divers inserted a well point to collect representative ground water discharging from Crescent Beach Spring. Ground water was analyzed for nutrients, ions, metals, stable- and radioisotopes, and age dates. Initial results suggest that discharge water is fresher than that observed in 1995 and lower in selective metals and major ions. Ammonia is an order of magnitude greater than overlying surface waters. Groundwater ammonia levels are consistent with reducing environments found in the Floridan Aquifer System. Previous work has been conducted at CBS but not to the extent that has been accomplished in this study. Comparing recent with previous data will help indicate what the possibilities of contaminating the aquifer source waters are as well as ascertaining groundwater-flow rates and residence times.

# Surface Geophysical Investigation of a Chemical Waste Landfill in Northwestern Arkansas

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## Abstract

In May 2000, the U.S. Geological Survey performed a surface geophysical investigation on a site used for disposal of unknown types of chemical waste in the 1960's. The site is located near Fayetteville, Arkansas, on the Springfield Plateau of the Ozark Plateaus physiographic province and is about 100 feet by 110 feet in size. The surface is flat lying and characterized by a 40-foot thickness of cherty clay regolith material underlain by the chert-rich, karst limestone of the Boone Formation. Information available about the site's history indicates that as many as six pits were excavated for the disposal of laboratory chemicals in glass containers that may or may not be intact. The objective of the surface geophysical investigation was to use noninvasive methods to delineate possible buried chemical zones. The information collected in this investigation may be useful to locate possible leachate plumes to optimize subsequent sampling and remediation.

Methods used at the site focused on the electrical insulating properties of the nonmetallic (glass) containers, electrical conducting properties of possible leaking fluids, and electromagnetic properties of the disturbed regolith material in the vicinity of the burial zones. The following geophysical methods were used at the site: 1) electromagnetic conductivity, 2) magnetometer, and 3) 2D-DC electrical resistivity survey. The electromagnetic survey was performed in horizontal co-planar mode, measuring both the quadrature and inphase component. The magnetometer survey was performed using a pair of memory magnetometers: one as a roving instrument and one as a stationary instrument continuously measuring the earth's magnetic field. The 2D-DC electrical resistivity survey consisted of 9 profiles of 28 electrodes at a 6-foot spacing with Wenner, Schlumberger, and dipole-dipole array data collected. This combination of geophysical tools was successful in delineating several types of subsurface anomalies consisting of buried metal and discrete high and low resistivity zones at various depths.

Five geophysical anomaly types were categorized and designated as "types A, B, C, D, and E" for reference when describing areas of concern at the site. Anomaly type A is indicated by a discrete high resistivity zone with an associated low resistivity zone below, possibly suggesting buried nonconductive material and leaking conductive fluid below it. Anomaly type B is indicated by a discrete high resistivity zone with no associated low resistivity zone below, possibly suggesting the presence of buried nonconductive material and no leaking conductive fluid. Anomaly type C is indicated by a medium resistivity zone adjacent to a high resistivity zone possibly suggesting the presence of disturbed regolith associated with burial. Anomaly type D is indicated by a zone of high resistivity material with responses from the magnetometer or the electromagnetic survey suggesting nearby buried metal. Anomaly type E is indicated by a discrete zone of high resistivity material underlying a shallow low resistivity zone with a negative electromagnetic response, which indicates the presence of buried metal. These five anomaly types were used to characterize and identify areas of concern that could be possible locations of buried materials at the site. Sampling and source removal plans are being developed based on areas of concern identified from the anomalies.

## INTRODUCTION

The U.S. Geological Survey performed a surface geophysical investigation at a chemical waste landfill in May 2000. The chemical waste landfill is located at Latitude 36 deg. 07' 16", Longitude 94 deg. 11' 22" which is in the SW ¼ of the NE ¼ of the NW ¼ of section 28, Township 17N, Range 30W, Washington County, Arkansas (fig. 1) and is about 100 feet by 110 feet in size. Anecdotal accounts by persons present during the placement of the waste suggest that as many as six pits possibly were excavated and used at the site during the 1960's for burial of laboratory chemicals in glass containers. The exact locations and extents of the pits are unknown.

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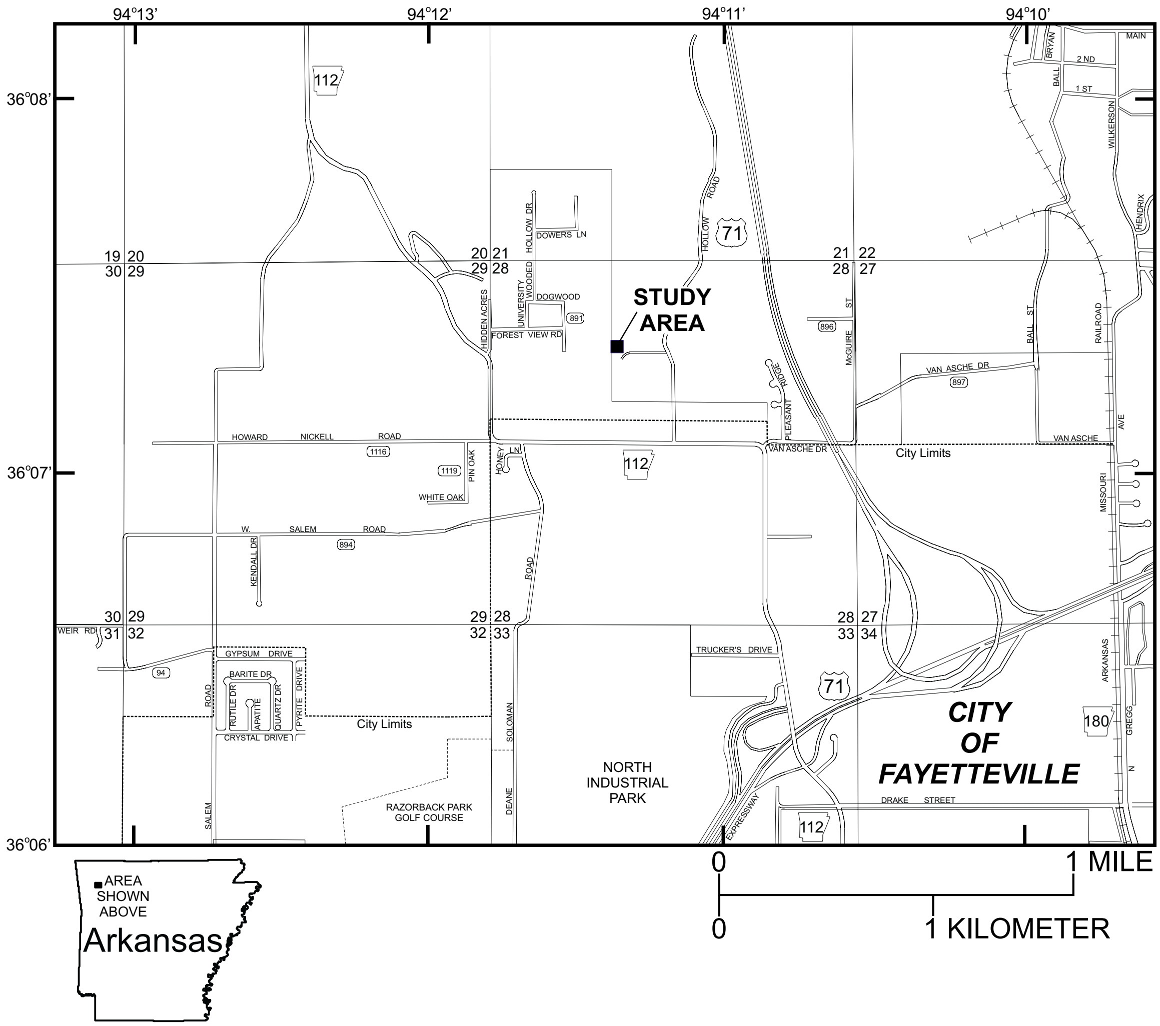



**Figure 1.** Location of the chemical waste landfill, near Fayetteville, Arkansas.

The site is located on the Springfield Plateau of the Ozark Plateaus physiographic province (Fenneman, 1938). The surface is flat lying and characterized by a cherty clay regolith underlain by the chert-rich, karst limestone of the Boone Formation. Small piles of chert cobbles exist on the northern and southern borders of the site. Depth to the top of the limestone is estimated to be approximately 40 feet, and water levels in local shallow domestic wells completed in the Boone Formation range from 20 to 32 feet below land surface. A seismograph station consisting of a concrete block building was constructed on the site in the early 1970’s. According to personnel involved with the construction and operation of the seismograph station, the building’s foundation is constructed of concrete and metal reinforcement with large pilings beneath it penetrating the cherty clay regolith to the top of the limestone.

The objective of the surface geophysical investigation is to use noninvasive methods to delineate possible buried chemical zones within the site boundary. The lack of historical information about the chemicals buried at the site requires that noninvasive methods be used to minimize the possibilities of causing further damage at the site. The information collected in this investigation may be useful to locate possible leachate plumes to optimize subsequent sampling and remediation. This paper summarizes the surface geophysical investigation performed at the chemical waste landfill.

A 100-foot by 110-foot grid was surveyed and staked on the site. The grid extent is considered to be the full areal extent of the chemical waste landfill and is the area of interest for the surface geophysical investigation.

## SURFACE GEOPHYSICS APPROACH AND FINDINGS

Materials buried at the site are believed to be in glass containers that may or may not be intact. Methods used at this site focus on the electrical insulating properties of the nonmetallic (glass) containers, electrical conducting properties of possible leaking fluids, and electromagnetic properties of the disturbed regolith material in the vicinity of the burial zones. Three different types of geophysical surveys were performed on a 10-foot northing/easting grid: 1) electromagnetic conductivity, 2) magnetometer, and 3) 2D-DC electrical resistivity.

References to locations at the site use the surveyed grid as a datum. The grid consists of 12 north-south (N-S) lines at 10-foot intervals designated by alphabetic characters F through Q, and 11 east-west (E-W) lines at 10-foot intervals designated by numbers 6 through 16. Locations outside the grid are designated with alphabetic or numbers characters beyond the limits of the surveyed grid, but retain the established dimensional pattern. Locations between lines are referred to as X.5, such as F.5 being between F and G.

### *Electromagnetic Conductivity Survey*

A Geonics[1] EM-31 unit was used for the electromagnetic conductivity (EM) survey. The EM-31 is a frequency-domain electromagnetic instrument that has a nominal penetration depth of about 18 feet in its horizontal co-planar mode (referring to the orientation of the coils). Disturbed soil generally has a higher porosity than undisturbed soil, with subsequent higher water content and therefore higher conductivity. The soil in a burial zone should appear more conductive unless mixed with less conductive material. Conductivity measurements were recorded for about 423 stations (fig. 2) on the grid with alternating 5-foot and 10-foot spacing using both in-line and broadside orientations (to check for anisotropy) and measuring inphase (to search for metallic debris) and quadrature (to gather stable soil conductivities) components for each orientation (totaling 1,692 measurements). The EM data were contoured using a kriging routine and overlaid on a map of the grid on four separate plots. Kriging is a geostatistical gridding method that produces contour maps that reveal trends in the data.

EM methods have been used effectively elsewhere to map locations of buried metallic objects, disturbed soil, and potential conductive-ion leachate plumes emanating from landfills (Bisdorf and Lucius, 1999). By measuring soil conductivity with the quadrature-phase component, it is possible to detect locations of disturbed moisture-bearing soil and conductive leachate plumes. The inphase component (measured in units of parts per thousand of primary electromagnetic field of the instrument) is primarily used for detection of metal objects, although metal objects also effect the quadrature phase measurements. Negative values usually indicate that the instrument is oriented perpendicular to a highly conductive object (such as iron or steel). Extremely high positive values of conductivity indicate that the metal object is aligned parallel to the orientation of the instrument.

Several anomalous values appear to be associated with buildings and known metal objects at the chemical waste landfill. The quadrature phase (fig. 2) and inphase measurements are affected in some areas by metal objects present at the site, such as two metal signs near F.5-7 and I-7.5, an empty steel drum near K.5-16, and the seismograph station (corners approximately at L.5-11, N-11, N-14, L.5-14). Other extrinsic readings occurred at E-15, associated with a storage building on adjoining property to the west; G.5-4, near some metal strapping on the surface; and a diagonal line of low conductivity extending from the southeast to the northwest from J-14 to H.5-16, probably associated with a buried power cable that surfaces at the west side of the seismograph station. The only conductivity highs that appear on the quadrature phase plot are associated with surface features such as the storage building and the seismograph station.

Several conductivity lows appear to be related to buried materials. A prominent conductivity low appears between I-7.5 and K-7.5 (fig. 2); this is confirmed by a negative inphase response to be caused by buried metal in that area. A similar cause and effect is present at M.5-6 on both quadrature (fig. 2) and inphase plots. There is a large pile of tree and brush debris present at this location that could contain

[1] Any use of trade, product, or firm names is for descriptive purposes only and does not imply endorsement by the U.S. Government.

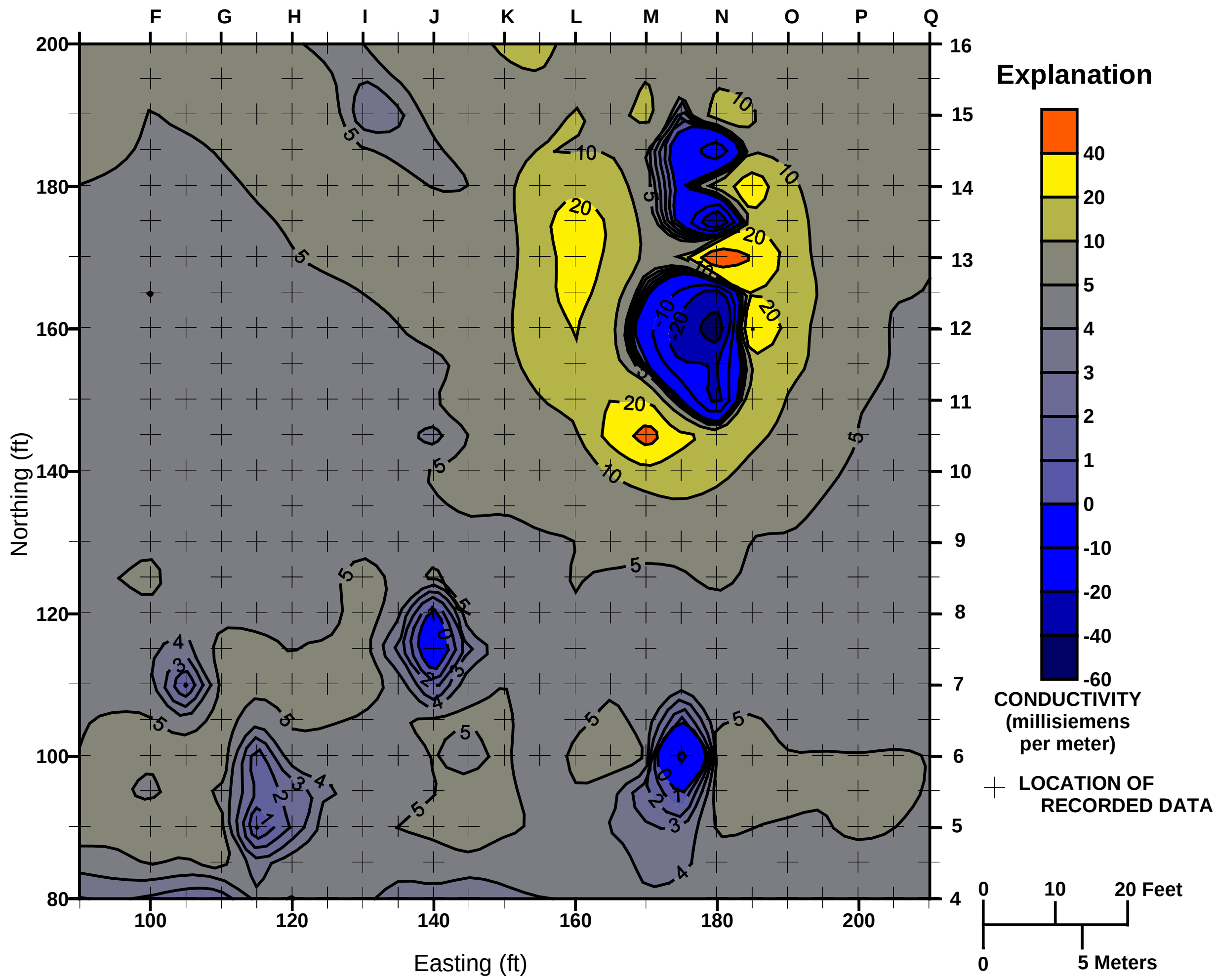



**Figure 2.** Electromagnetic conductivity (north-south orientation) quadrature phase component survey.

some metal. At J-10.5, an anomaly of slightly low conductivity is present on the quadrature phase plot, especially in the N-S orientation (fig. 2), possibly indicating a nonconductive mass aligned east-west. This is supported by the lack of anomalies on the inphase plots at J-10.5 or I.5-11.

## *Magnetometer Survey*

The magnetometer survey was performed on the morning of May 13, 2000. The magnetometer system comprised a pair of Geometrics G-856 memory-magnetometers, which measure magnetic intensity in units of nanoteslas. One was used as a roving instrument which recorded measurements at stations on the surveyed 10-foot grid, and one was placed nearby at an off-site location as a base station to continuously record the earth's magnetic field. The magnetic data were downloaded to a laptop computer and the base station record was used to correct the roving station data for diurnal drift. The measurements were then contoured using a kriging routine and overlaid on the grid (fig. 3). The natural magnetic intensity at the base station location near the chemical waste landfill on the day of the survey was about 52,800 nanoteslas.

Magnetometer readings are expected to be erratic in the presence of disturbed soil and ferrous materials, both of which may be present at the chemical waste landfill. The major feature present on the magnetometer plot between gridlines L and N and 11 and 14 is the magnetic intensity extreme associated with the seismograph station building. Unfortunately, the effect from this building masks any subtle differences in magnetic intensity that could be present within 20 feet of the building, thus making this area of the survey invalid toward the objective of the investigation. Other extrinsic responses include the two metal signs near locations G-7 and I-8 appearing as lower

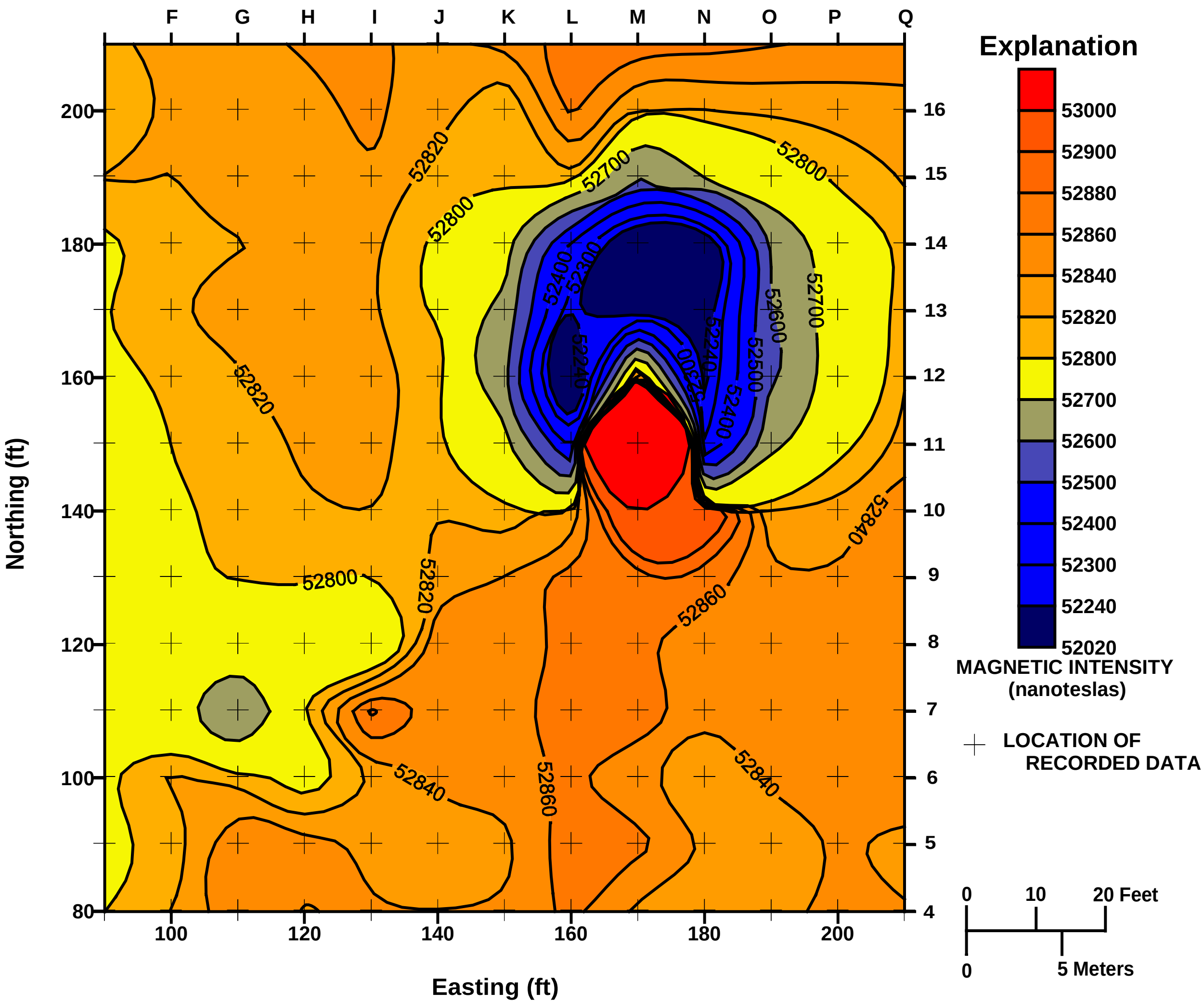



**Figure 3.** Magnetometer survey.

intensity on the magnetometer survey plot (fig. 3) and the empty steel drum near L-16 appearing as higher intensity.

A response of concern is located near I-7 to I.5-7 (fig. 3), which is an increase of magnetic intensity probably associated with the metal detected with the EM at J-7.5. The shifted nature of the magnetic anomaly with respect to the EM anomaly is typical of a magnetic field of a monopole object buried at an angle from horizontal. A monopole has lines of equal magnetic field that point radially in or out from the positive or negative monopole respectively (Briener, 1973). The object is probably shallow (1-3 feet deep) as indicated by the limited extent of the anomaly.

The area of relatively higher magnetic intensity (greater than 52,860 nanoteslas) south of the seismograph station building (fig. 3) does not correlate with any EM anomalies and therefore is probably associated with natural variations of magnetic intensity of the regolith present at the site.

The EM anomaly present at M.5-6 (fig. 2) does not appear in the magnetometer survey, which indicates that the metal present is not of sufficient mass or does not contain enough ferrous metal to change the magnetic intensity. The fact that it affected the EM readings so drastically, combined with the lack of magnetometer response, indicates that a small light metal object was very near the EM instrument, possibly in a nearby brush pile.

In Eve L. Kuniansky, editor, 2001, U.S. Geological Survey Karst Interest Group Proceedings, Water-Resources Investigations Report 01-4011, p. 107-115

## *2D-DC Electrical Resistivity Survey*

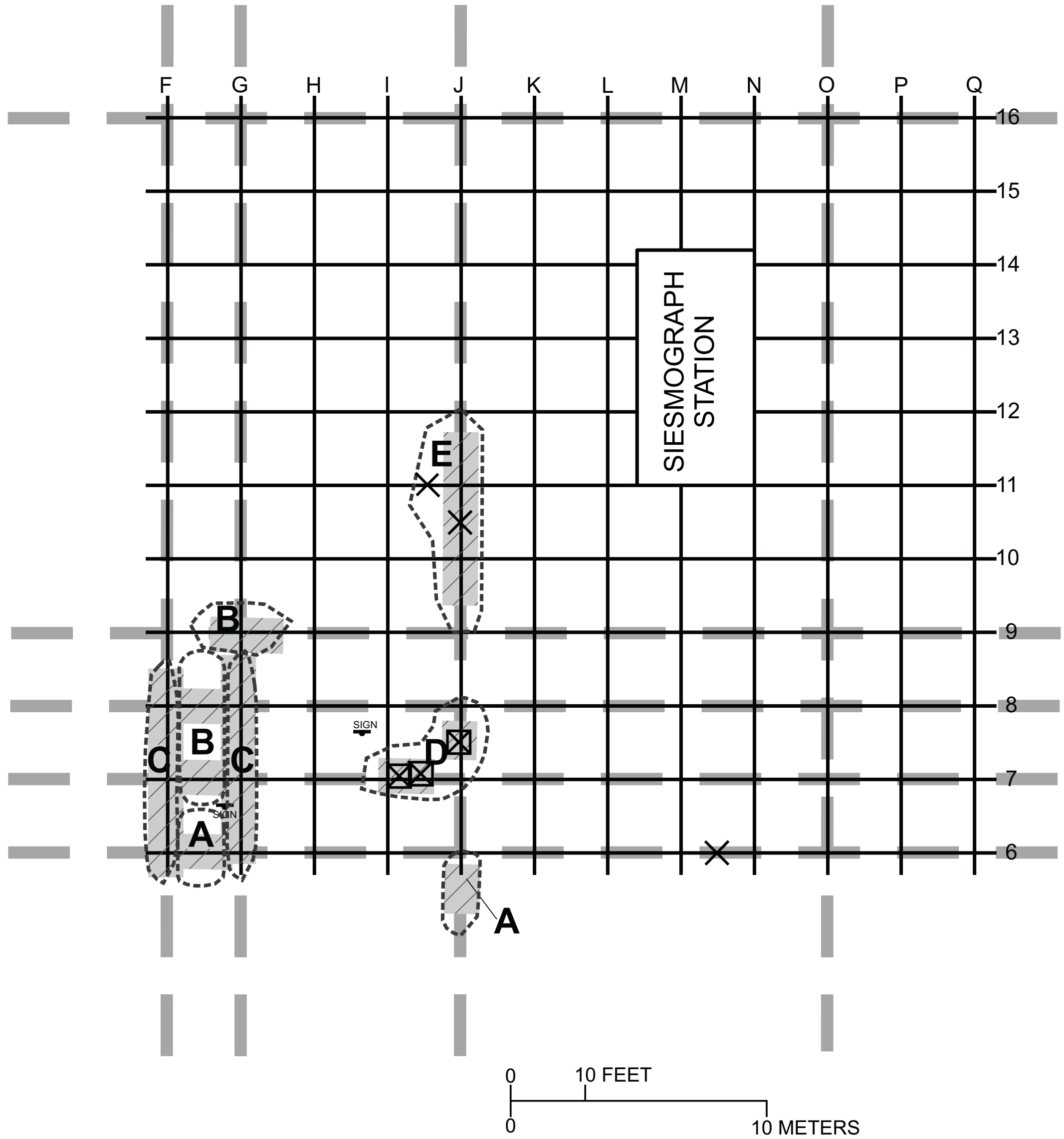
Electrical resistivity surveys are commonly used for hydrogeological, mining, and geotechnical investigations, and environmental surveys (Loke, 1999a). Subsurface electrical resistivity is related to buried materials and various geological and hydrogeological parameters such as the mineral and fluid content, porosity and water saturation.

A 2D-DC (two-dimensional, direct-current) electrical survey was performed at the chemical waste landfill to determine the subsurface resistivity distribution by making measurements on the ground surface. From these measurements, the true resistivity of the subsurface could be estimated. The resistivity directly measured in the field survey is not the true resistivity of the subsurface but an “apparent” resistivity value. The apparent resistivity value equals that of the resistivity of a homogeneous ground measured from the same electrode arrangement. To determine the true subsurface resistivity, an inversion of the measured apparent resistivity values must be performed (Loke, 1999a). The result is a contoured representation of the apparent and inverse modeled “true” resistivity in a cross-sectional format. In this format, the insulating quality of the glass containers thought to be present at the site are expected to appear as higher resistivity values among the lower resistivity values of the disturbed regolith. The heterogeneity of the regolith material made some interpretations difficult.

The 2D-DC resistivity system used at the chemical waste landfill was the Sting/Swift resistivity system by Advanced Geosciences, Inc. Of the three different arrays (Wenner, Schlumberger, and dipole-dipole) that were used initially, the dipole-dipole was the most successful because of its detailed resolution near the land surface. Based on the anomalies observed, the nine dipole-dipole profiles, which were collected at 6-foot electrode spacings, were further evaluated. They were located along the following grid lines: 6, 7, 8, 9, 16, F, G, J, and O (fig. 4). Because of the trapezoidal geometry of the dipole-dipole cross-sectional profiles, the lines were extended beyond the surveyed grid to optimize subsurface coverage.

There are several steps involved in producing a cross-sectional profile of “true” resistivity of the subsurface. The field measurements are spatially related to positions in the cross section and recorded by a datalogger. The field data are subsequently downloaded to a computer and input into the finite-difference model RES2DMOD (Loke, 1999), which inverts the data and creates a contoured profile of field apparent resistivity. This apparent resistivity profile is used to create a simplified integrated model of the cross section, which is subdivided into discrete blocks with specific resistivity values. The finite-difference model is run on the interpreted model grid and the resulting modeled resistivity profile from the inversion is compared to the apparent resistivity profile. Through a trial-and-error process, the interpreted model is improved to create a modeled resistivity profile that closely resembles the field apparent resistivity profile. The interpreted model is then considered to have produced a representation of “true” resistivity of the subsurface when a close match is made between the modeled resistivity profile and the apparent resistivity inversion profile. The nine 2D-DC resistivity dipole-dipole profiles were evaluated in detail and the results were used in conjunction with results from the electromagnetic conductivity and magnetometer surveys to delineate areas of concern.

In the southwest corner of the site (fig. 4), an area of concern was identified through the analysis of 2D-DC resistivity profiles along gridlines 6, 7, 8, 9, F, and G. Three types of anomalies are present in this area. The first type, designated as type A, corresponds to a high resistivity zone above a low resistivity zone. This could be caused by nonconductive containers buried 6 to 9 feet deep with a conductive substance leaking below. The type A anomaly in this area is most prominent on east-west trending gridline 6, between gridlines F and G. Another anomaly type similar to the type A, without the associated conductive zone below the high resistivity zone (designated as type B), is present near grid location F.5-7 to G-9. It is unclear if this is in response to the possible buried nonconductive material present along gridline 6 or if this is a zone of higher resistivity regolith material such as a greater concentration of chert. The type B anomaly possibly suggests the presence of buried nonconductive material without leaking conductive fluid. The third type of anomaly present in this area, referred to as type C, is characterized by a medium resistivity zone adjacent to a type A or B anomaly. This type of anomaly is indicative of disturbed regolith. Type C anomalies in this area are located along gridlines F and G. Affected sections of the profiles are indicated on figure 4.




**EXPLANATION**

- 2D-DC RESISTIVITY PROFILE LINE
- 2D-DC RESISTIVITY ANOMALY
- AREA OF CONCERN
- A — ANOMALY TYPE
- EM-31 ANOMALY
- EM-31 AND MAGNETOMETER ANOMALY

ANOMALY TYPES

TYPE A--INDICATED BY A DISCRETE HIGH RESISTIVITY ZONE WITH AN ASSOCIATED LOW RESISTIVITY ZONE BELOW IT

TYPE B--INDICATED BY A DISCRETE HIGH RESISTIVITY ZONE WITH NO ASSOCIATED LOW RESISTIVITY ZONE BELOW IT

TYPE C--INDICATED BY A MEDIUM RESISTIVITY ZONE ADJACENT TO A HIGH RESISTIVITY ZONE

TYPE D--INDICATED BY A HIGH RESISTIVITY ZONE IN CONJUNCTION WITH RESPONSES FROM THE MAGNETOMETER AND THE ELECTROMAGNETIC SURVEY

TYPE E--INDICATED BY A DISCRETE ZONE OF HIGH RESISTIVITY MATERIAL UNDERLYING A SHALLOW LOW RESISTIVITY ZONE WITH A NEGATIVE ELECTROMAGNETIC RESPONSE

**Figure 4.** Locations of anomalies and areas of concern.

An additional type A anomaly indicated by 2D-DC resistivity is present just south of grid location J-6 on gridline J. This area of concern is areally less extensive than the type A anomaly in the southwest corner of the site, but appears at the same approximate depth. It appears as a discrete zone of high resistivity 6 to 9 feet deep above a zone of low resistivity. This could be indicative of nonconductive containers buried 6 to 9 feet deep with a conductive substance extending below. The affected section of the profile is indicated on figure 4.

At grid location I-7 to J-7, the EM and magnetometer indicated probable buried metal in the area and the 2D-DC resistivity indicated a zone of moderately high resistivity present 3 to 6 feet deep. This type of anomaly is designated as a type D anomaly. It is unclear if the high resistivity is caused by an increased concentration of chert or by buried nonconductive material. Another type D anomaly, present at J-7.5, could be a continuation of the one at I-7 to J-7. Affected sections of the profiles are indicated on figure 4.

At grid locations J-9 to J-12, a discrete zone of high resistivity underlies a low conductivity zone near the surface at J-10, which is corroborated by an indication of buried metal from the EM data. This is designated as a type E anomaly. The affected section of the profile is indicated on figure 4.

The areas described herein represent the possible areas of concern delineated by the surface geophysics investigation within the limits of interpretation of the collected data. Any drilling or excavation at the site may result in exposure of potentially hazardous materials. Additional areas of concern may exist on the site that are beyond the limits of detection of the methods and equipment utilized. Within reasonable limits, every effort was made to delineate possible burial zones applying widely used methods and standards.

## SUMMARY

In May 2000, the U.S. Geological Survey performed a surface geophysical investigation on a site used for disposal of unknown types of chemical waste in the 1960's. The site is located near Fayetteville, Arkansas on the Springfield Plateau of the Ozark Plateaus physiographic province and is about 100 feet by 110 feet in size. The surface is flat lying and characterized by a 40-foot thickness of cherty clay regolith material underlain by the chert-rich, karst limestone of the Boone Formation. Information available about the site's history indicates that as many as six pits were excavated for the disposal of laboratory chemicals in glass containers that may or may not be intact.

The objective of the surface geophysical investigation was to use noninvasive methods to delineate possible buried chemical zones. The information collected in this investigation may be useful to locate possible leachate plumes to optimize subsequent sampling and remediation. Methods used at the site focused on the electrical insulating properties of the nonmetallic (glass) containers, electrical conducting properties of possible leaking fluids, and electromagnetic properties of the disturbed regolith material in the vicinity of the burial zones. Several areas of concern at the chemical waste landfill appear to have been impacted through burial of various types of material.

The areas of concern were discovered though electromagnetic, magnetometer, and 2D-DC resistivity surface geophysical methods and are suspected to contain buried materials. The electromagnetic survey was performed in horizontal co-planar mode, measuring both the quadrature and inphase component. The magnetometer survey was performed using a pair of memory magnetometers, one as a roving instrument and one as a stationary instrument continuously measuring the earth's magnetic field. The 2D-DC electrical resistivity survey consisted of 9 profiles of 28 electrodes at a 6-foot spacing with Wenner, Schlumberger, and dipole-dipole array data collected. This combination of geophysical tools was successful in delineating several types of subsurface anomalies consisting of buried metal and discrete high and low resistivity zones at various depths.

Five geophysical anomaly types were categorized and designated as "types A, B, C, D, and E" and were used to delineate areas of concern at the site. Anomaly type A consists of a discrete high resistivity zone with an associated low resistivity zone below, possibly suggesting buried nonconductive material and leaking conductive fluid below. Two type A anomalies were found, one along gridline 6 between gridlines F and G, and the other south of gridline 6, along gridline J. Anomaly type B consists of a discrete high resistivity zone with no associated low resistivity zone below, possibly suggesting the presence of buried nonconductive material without leaking conductive fluid below. A type B anomaly exists between gridlines F and G at gridlines 7 and 8 and another exists near grid location G-9. Anomaly type C is indicated by a medium resistivity zone adjacent to a high resistivity zone possibly suggesting disturbed regolith associated with burial. Two type C anomalies are evident on the eastern and western sides of the type A and B anomalies mentioned above. Anomaly type D is indicated by a zone of highly resistive material in conjunction with responses from the magnetometer or the electromagnetic survey suggesting buried metal nearby; this anomaly is located near grid locations I-7 to J-7. Anomaly type E is indicated by a discrete zone of high resistivity material underlying a shallow low resistivity zone with a negative electromagnetic response, which indicates the presence of buried metal; this anomaly is located near grid locations J-9 to J-12. These five anomaly types were used to characterize and delineate the possible locations of buried materials at the site. Sampling and source removal plans are being developed based on delineations of the anomalies.

The areas described herein represent the possible areas of concern delineated by the surface geophysics investigation within the limits of interpretation of the collected data. Any drilling or excavation at the site may result in exposure of potentially hazardous materials. Additional areas of concern may exist on the site that are beyond the limits of detection of the methods and equipment utilized. Within reasonable limits, every effort was made to delineate possible burial zones applying widely used methods and standards.

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# Borehole Geophysical Applications in Karst Hydrogeology

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## Abstract

Geophysical measurements in boreholes provide useful information about subsurface aquifers, but the heterogeneity of karst aquifers poses a real challenge in conventional well logging. Borehole geophysics presents several tools that may be applied to the characterization of heterogeneous karst aquifers: (1) Image and flow logs provide detailed information about the nature of hydraulically active zones intersected by boreholes; (2) Geometric correlation of logs between boreholes indicates possible connections between the zones in separate boreholes; (3) Water-quality and hydraulic-head data derived from logs can be used to identify subsurface flow paths; (4) Cross-borehole flow experiments can be used to infer the properties of hydraulic connections among subsurface conduits; and (5) Geophysical measurements can be made at local, intermediate, and large scales to infer the relation between scale and hydraulic conductivity. Examples of these applications are given for karst or karst-like aquifers at sites in Arizona, Illinois, Kentucky, New Hampshire, and Florida, demonstrating the specific contributions that borehole geophysics can make in karst characterization.

## INTRODUCTION

Borehole geophysical logging is a commonly used technique for the in situ characterization of aquifers. However, heterogeneous karst aquifers are difficult to describe using data obtained in a limited number of boreholes. Although the logs provide detailed information about formation properties in the immediate vicinity of boreholes, one can never expect to drill enough boreholes to characterize karst flow systems on the basis of borehole data alone. Boreholes cannot sample enough of the heterogeneous subsurface openings to provide a representative sample of distribution of permeability because of the variability of the fractures, bedding planes, and vugs connected to form flow paths. This paper describes a number of ways in which geophysical measurements in boreholes can be combined with other data to mitigate the severe challenges presented by the need to characterize karst flow systems. Each of these is illustrated by a specific example. Further development of these techniques will be needed to improve the ability to characterize karst aquifers using the available borehole characterization technology.

## IDENTIFYING HYDRAULICALLY ACTIVE ZONES

Most geophysical logs provide precise information about the in situ properties of subsurface formations but in the form of measurements such as gamma activity or electrical conductivity that are only indirectly related to hydraulic parameters of interest. Previous results generally indicate that the transmissivity of bedding planes, fractures, and solution openings cannot be inferred from the appearance of those features on borehole image logs or the apparent aperture of those features on caliper logs (Paillet, 1998). Recently developed high-resolution flow logging equipment such as the heat-pulse (Hess, 1986) and electromagnetic (Molz and others, 1994) flowmeters add the important ability to tie borehole hydraulics to geophysical log data. An example is illustrated in figure 1 for a massive limestone aquifer in northern Arizona. The geophysical logs indicate the precise depths where water exits the borehole during steady injection. The outflow points can be associated with features on the other logs that represent the hydraulically conductive features in the vicinity of the borehole. In figure 1, these features include fractures, bedding planes, and a small cavern. Although the full set of logs provides no information about how far these features extend away from the borehole, the flow log does allow some analysis of the limited set of features where flow actually occurs. This is an important step beyond simply identifying the fractures and solution features that intersect boreholes.

## CORRELATION OF ZONES BETWEEN BOREHOLES

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One possible approach in understanding how hydraulically conductive fractures, bedding planes, and solution openings identified in boreholes are connected to form flow systems is to project these features in the regions between boreholes. This seems simple in principle, but becomes difficult in practice when there are many possibly permeable features in each borehole and boreholes are located far apart. Spatial correlation on the basis of appearance in image logs and occurrence at similar depths is generally not effective. A much more effective approach is to locate permeable openings with respect to sedimentary structure. For example, inflow to or outflow from a series of boreholes in northern Illinois was associated with solutionally enlarged bedding plane openings (fig. 2). Many such bedding planes intersected each borehole, but only a few conducted most of the flow (Paillet and Crowder, 1996). The correlation of these bedding planes was established over borehole separations of about a kilometer by correlating gamma logs. The gamma correlation established the strike and dip of bedding so that borehole elevation and the regional dip could be used to define the precise stratigraphic position of the bedding planes in each borehole. This structural correlation showed that sets of bedding planes served as regional conduits, but that the most transmissive bedding plane within sets of closely-spaced planes varied from one borehole to the next. The combination of structural correlation and flow log analysis can be useful in identifying how solution openings are organized into continuous flow paths even when large-scale hydraulic test data are not available.

## IDENTIFYING FLOW PATHS AND COMPARTMENTS

Although structural correlation of solution features on the basis of aquifer geometry is a useful technique, there are other physical/chemical ways to identify how solution openings identified in individual boreholes might be connected in the regions between boreholes. These techniques include methods that identify the chemical or hydraulic-head signatures of aquifer zones. For example, gross chemical signature measurements of zones can be made in open boreholes. Examples of such open-borehole water chemistry techniques are given by Paillet and Pedler (1996) and Tsang and others (1990). Flowmeter logging provides a technique for determining the hydraulic head of individual zones using the numerical flow model inversion of Paillet (1998, 2000). Figure 3 illustrates an example where the flow modeling technique is used to quantify the hydraulic heads of the three different solution openings contributing flow to a borehole. Proper application of either the water-quality or hydraulic-head techniques involves the measurement of fluid column electrical conductivity or borehole flow under at least two different quasi-steady borehole flow conditions. In each case, correlation of water quality or hydraulic head between boreholes can be used to identify large-scale flow paths in karst aquifers.

## CROSS-BOREHOLE FLOW EXPERIMENTS

Suspected connections along flow paths between boreholes can be characterized using the cross-borehole flow method described by Paillet (1998). The transient flow response at selected positions in an observation borehole can be used to identify the hydraulic properties of the flow path between the depths where that path intersects the two boreholes. A simple example is shown for a shallow, karst-like flow path between boreholes in granitic rock in New Hampshire in figure 4. In this example, the water-producing zone in each of four boreholes can be identified. Although hydraulic-head measurements clearly show that this one zone is connected to all four boreholes, the water-producing zones appear very different in each case. The image log interpretations (fig. 4A) vary from clean, steeply dipping fractures to highly altered and enlarged openings of no particular orientation. Simple spatial correlation suggests a sub-horizontal zone of permeability. A cross-borehole experiment shows that the flow induced in an observation borehole after a pump is turned on in one of the other boreholes precisely matches the type curve for a single infinite planar opening of specified transmissivity and storage coefficient (fig. 4B). This kind of information could not have been derived from the combined analysis of the images and geophysical properties of the individual water-producing features at the four places where they intersect individual boreholes. Such cross-borehole flow experiments provide a useful way to characterize the properties of flow paths connecting boreholes after other analysis has identified the existence of such connections.

## INTEGRATING MULTIPLE-SCALE DATA

Even in the most well-funded karst studies, there is unlikely to be enough drilling to characterize flow paths using the combination of flow logging, stratigraphic correlation, and cross-borehole flow testing. Non-invasive surface geophysical soundings can provide the area-wide coverage needed to fully characterize the subsurface. There are commonly two serious problems with using these soundings: 1) Sounding interpretations can be ambiguous; and 2) Subsurface flow depends on such factors as flow boundary conditions that cannot be derived from images of aquifer geometry alone. The combination of limited fine-scale, definitive data from boreholes and large-scale but ambiguous data from surface geophysics can serve to resolve the ambiguity by eliminating alternative models for geophysical interpretation and by relating subsurface hydraulic conditions to measured aquifer properties. For example, Paillet and others (1999) and Paillet and Reese (2000) used electromagnetic sounding profiles to characterize a shallow carbonate aquifer in south Florida. Aquifer units and regional hydraulic gradients were identified using borehole log data. These local aquifer units identified at drilling sites were then extended into the approximately 10 kilometers separations between boreholes using time domain electromagnetic (TEM; Fitterman and Stewart, 1986) soundings. The expected ambiguity in sounding interpretation was resolved by using soundings near boreholes to relate TEM interpretation models to the subsurface conductivity values (fig. 5). These results could be used to (1) eliminate alternate but otherwise equivalent interpretation model geometry; and (2) identify the spatial resolution in the TEM data set. Further development of the integrated analysis of surface geophysical soundings with geophysical, geologic, and hydrologic data from boreholes appears to be the only effective way to completely characterize karst terrain over the full range from local-borehole (1 meter or less) to site-wide (1 kilometer or more) scales.

## CONCLUSIONS

Geophysical measurements in boreholes provide useful information about subsurface aquifers, but the heterogeneity of karst aquifers poses a real challenge in conventional well logging. Borehole geophysics presents several tools that may be applied to this fundamental problem. Image and flow logs provide detailed information about the nature of hydraulically active zones intersected by boreholes that can be used to infer how these zones fit into regional hydrogeology. Geometric correlation of logs between boreholes indicates possible connections between zones in the regions between boreholes. Water-quality and hydraulic-head data derived from logs can be used to identify possible hydraulic connections between subsurface flow paths. Cross-borehole flow experiments can be used to infer the properties of hydraulic connections among subsurface conduits. Geophysical measurements can be made at local, intermediate, and large scales to infer the relation between scale and hydraulic conductivity. Examples of these applications show how they contribute to the characterization of karst or karst-like aquifers at sites in Arizona, Illinois, Kentucky, New Hampshire, and Florida, demonstrating the specific contributions that borehole geophysics can make in karst characterization.

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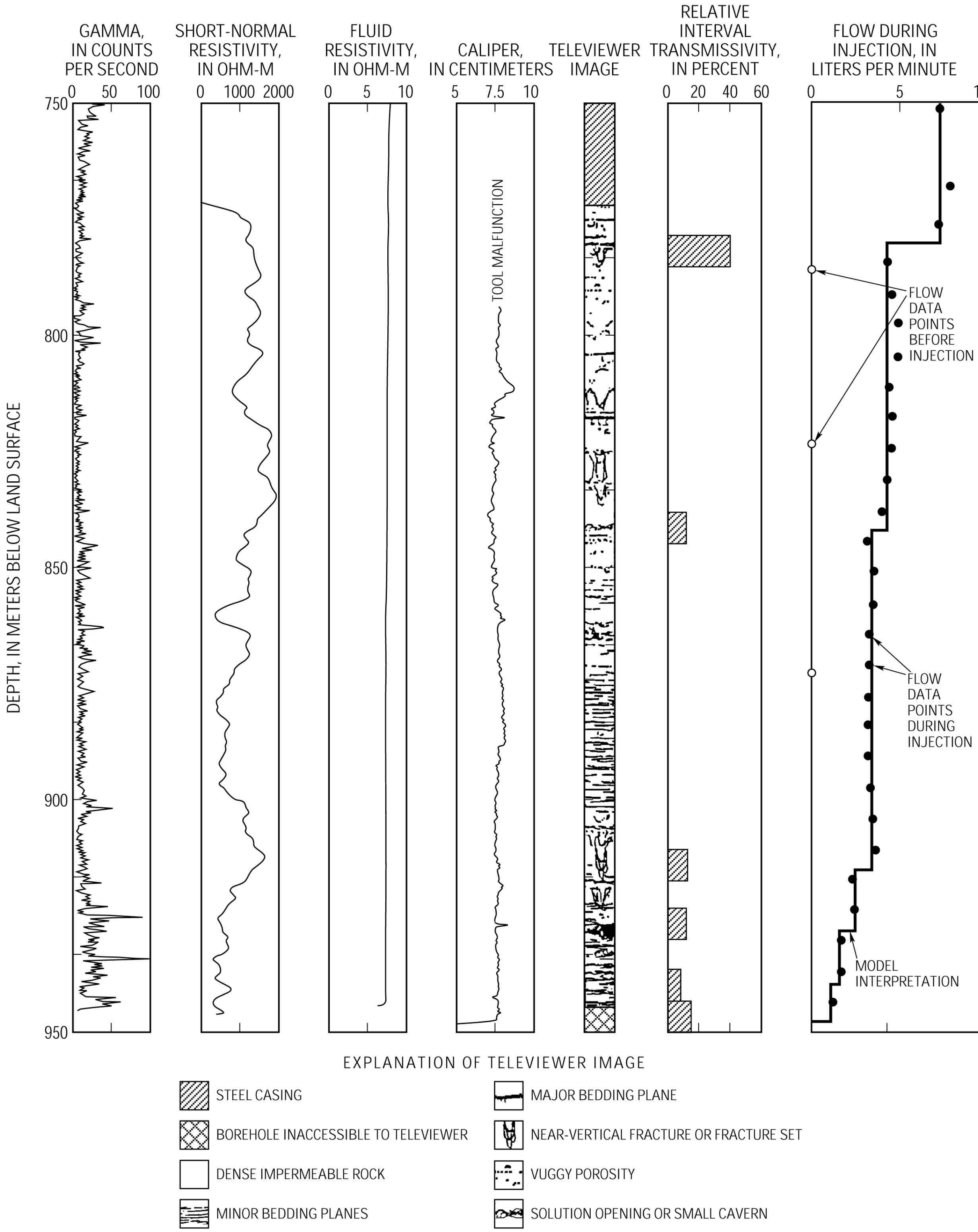



Figure 1–Composite of gamma, short-normal resistivity, fluid column resistivity, caliper, and televiewer logs compared with a borehole flow profile obtained with a heat-pulse flowmeter during injection in a borehole in fractured and bedded limestone in northern Arizona; data from Paillet (1998).

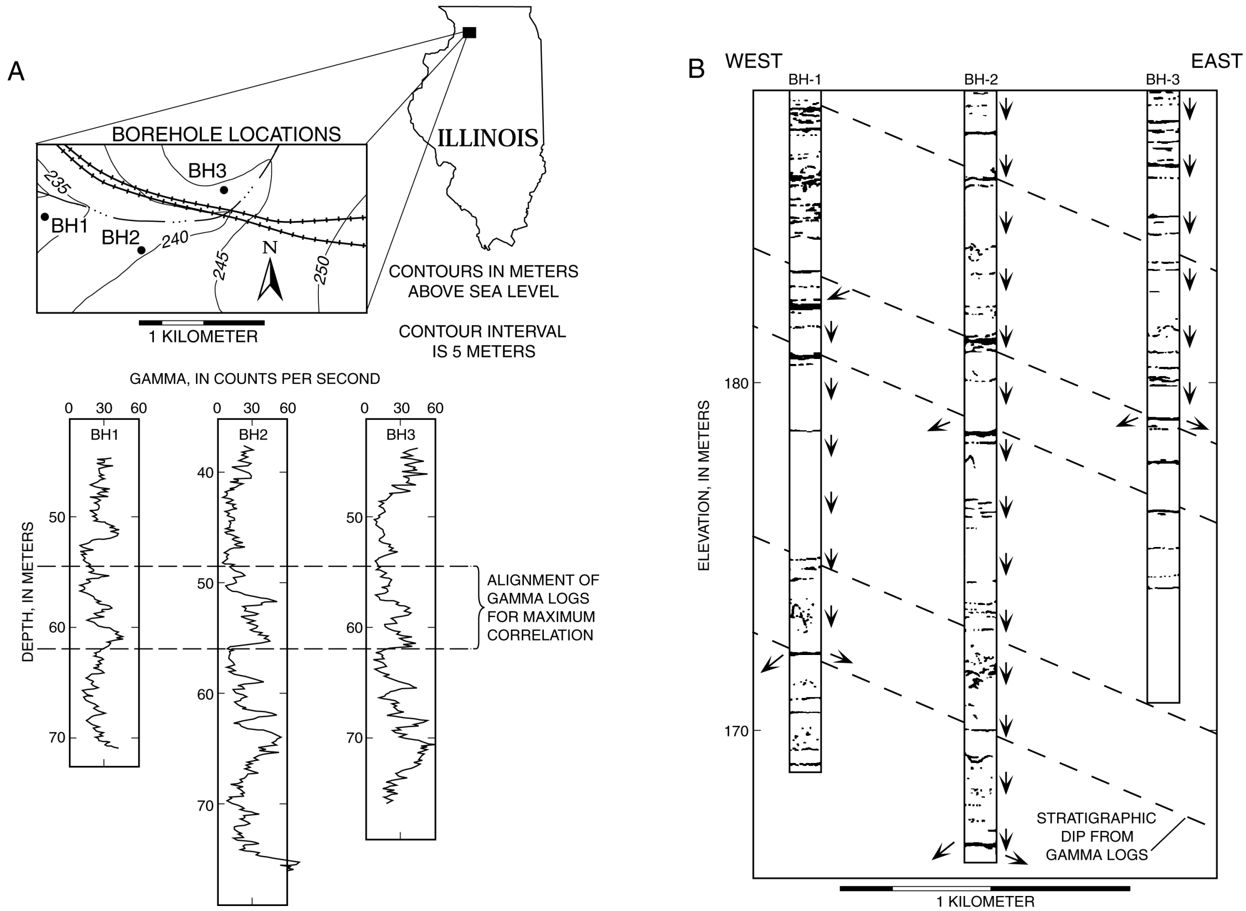
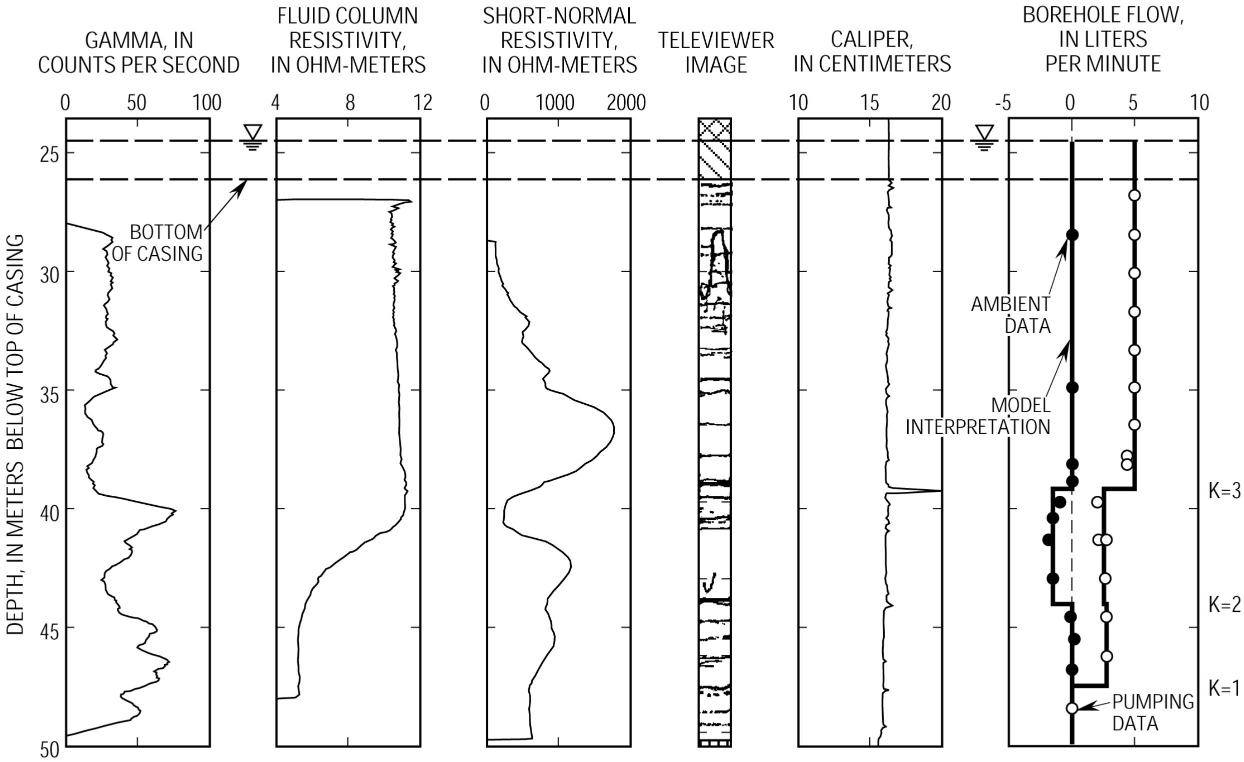



Figure 2–A) Horizontal alignment of gamma logs along an east-west profile used to identify stratigraphic dip in a dolomite aquifer in northern Illinois; and B) Televiewer logs, stratigraphic correlations, and flowmeter information used to identify continuous bedding planes and aquifer flow zones where arrows indicate direction of ambient flows measured in boreholes and locations of inflow and outflow; from Paillet and Crowder (1996).




FLOW MODEL INTERPRETATATION

| ZONE NUMBER | DEPTH (METERS) | TRANSMISSIVITY ($10^{-5} m^2/s$) | HYDRAULIC HEAD (m BELOW REFERENCE) |
|---|---|---|---|
| 3 | 39.0 | 5.0 | 24.6 |
| 2 | 43.5 | 5.0 | 25.8 |
| 1 | 47.0 | 10.0 | 25.5 |

Figure 3–Composite of gamma, fluid column resistivity, short-normal resistivity, televiewer, and caliper logs compared with flow profiles obtained under ambient and pumped conditions in fractured and bedded limestone in south-central Kentucky; data from Paillet (2000).

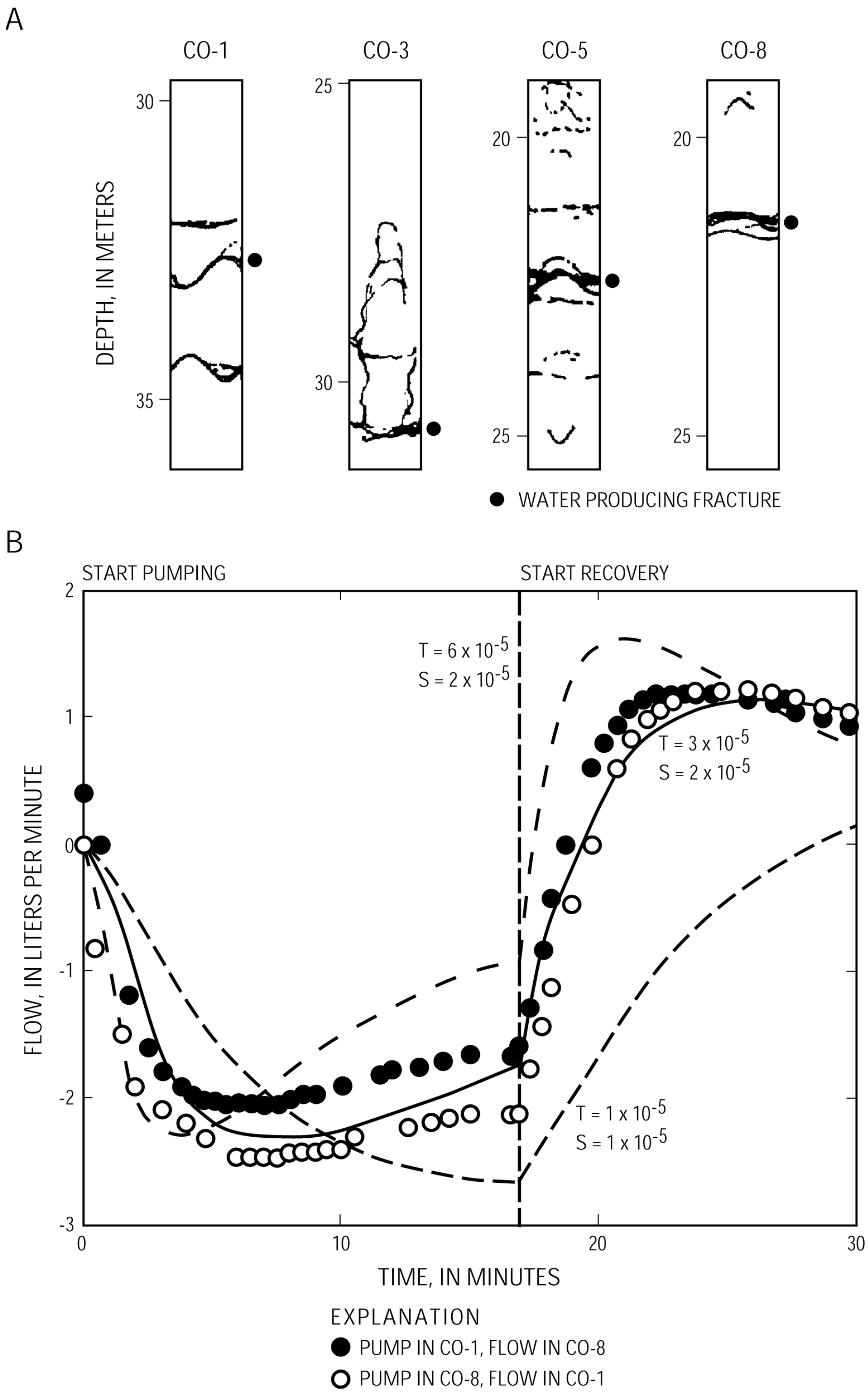



Figure 4–A) Televiewer logs of altered, water-producing fracture zones in granitic schist at a site in central New Hampshire; B) A cross-borehole flow experiment, where flow is measured above a fracture zone in one borehole while a pump is turned on and off in an adjacent borehole, shows that the fracture zone containing these features can be modeled as a single horizontal fracture of uniform transmissivity (T) and storage coefficient (S); modified from Paillet (1998).

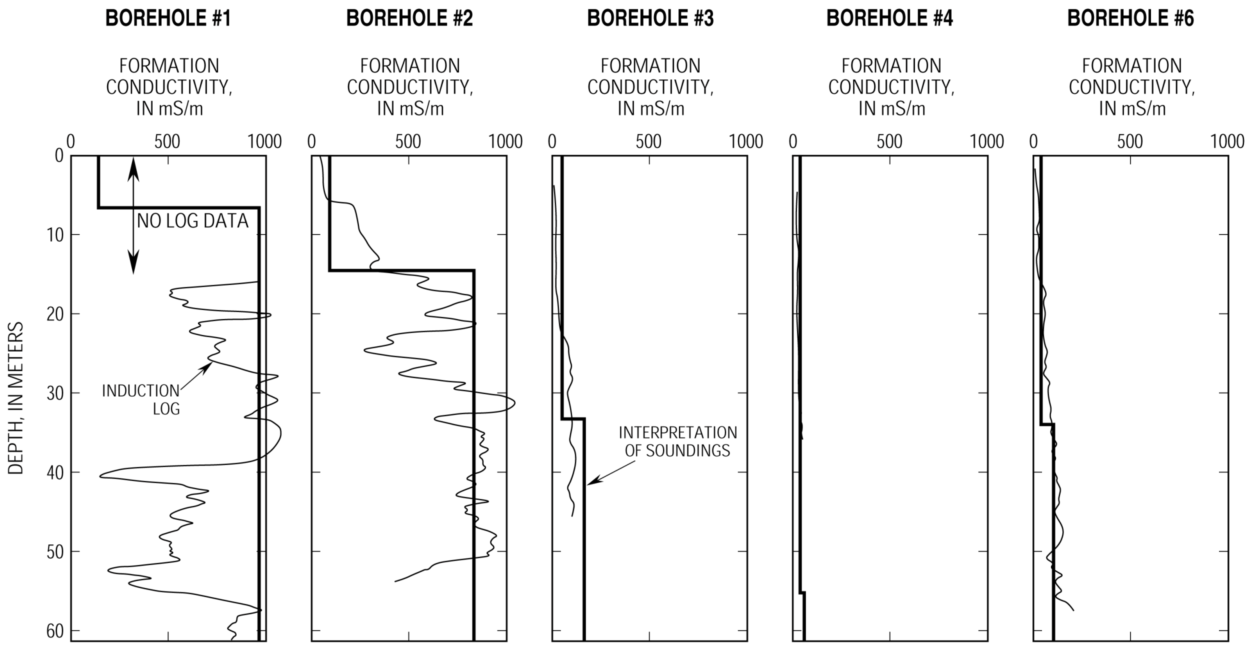



Figure 5–Two-layer interpretations of electromagnetic soundings made adjacent to boreholes with induction logs; these data were used to identify the appropriate inversion model for the projection of aquifers between borehole sites as described by Paillet and others (1999).

# Subsurface Characterization of Selected Water Bodies in the St. Johns River Water Management District, Northeast Florida

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Fluid exchange between surficial waters and groundwater, as well as the processes that control this exchange, are of critical concern to water management districts and planners. Digital high-resolution seismic systems were used to collect geophysical data from more than 40 lakes and rivers of northeastern Florida. Seismic data acquisition in the past has been only partly successful for imaging lake subbottom geology; however, the use of digital seismic technology has greatly enhanced potential applications. Seismic profiles collected from sites in northeastern Florida have demonstrated the potential application of these techniques in understanding the formation of individual lakes. In each case study, geologically controlled solution and/or mechanical processes determined the geomorphology of lakes and evidence of these processes may be seen in seismic profile. Processes that control lake development are twofold: 1) karstification or dissolution of the underlying limestone, and 2) collapse, subsidence, or slumping of overburden to form sinkholes. Initial lake formation is directly related to the karst topography of the underlying host limestone (Fig. 1). Lake size and shape are factors determining the thickness of overburden and size of the collapse or subsidence, and/or the clustering of lake-forming depressions.

Lake evolution follows sequential stages to maturity that creates progression through the following geomorphic types (Kindinger and others, 1999, 2000) (Fig. 2):

- (1) active subsidence or collapse phase (young) - the open to partially filled collapse structures typically associated with sinkholes;
- (2) transitional phase (middle age) - the sinkhole becomes plugged as the voids within the collapse are filled with sediment, periodic reactivation may occur;
- (3) baselevel phase (mature) - active sinkholes are progressively plugged by the continual erosion of material into the basin, and eventually sediment fills the basins;
- (4) polje (drowned prairie) - broad flat-bottom basins located within the epiphreatic zone that are inundated at high stages of the water table and have one or all phases of sinkhole development and many types of karst and karren features.

Most lakes in this study are small (less than1-km diameter) making stratigraphic correlation from lake to lake difficult. Seismic profiles of subsurface features were used to define the lacustrine geologic history and to locate possible breaches in the confining layer that maintains these lakes (Fig. 3). Six types of acoustic signatures were identified from the seismic profiles to describe the structural history of each lake (Fig.4). Using these criteria, Florida lakes can be classified by size, sediment fill, subsurface features, and geomorphology. Classification of lakes utilizing digital seismic technology has led to a better understanding of the interaction between the geology and hydrology of central Florida.

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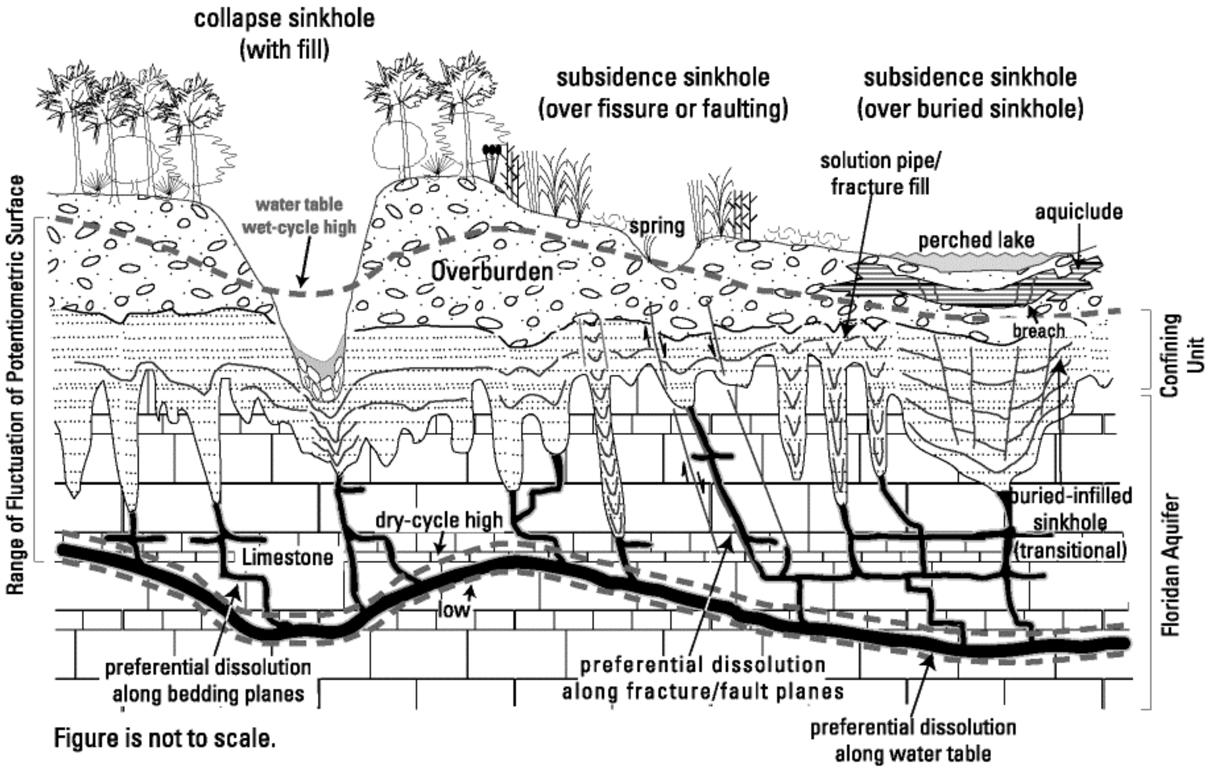


Figure 1. Diagnostic solution and collapse features of karst and karren topography from northeastern Florida. Individual sinkholes range from less than 1 m to more 100 m.

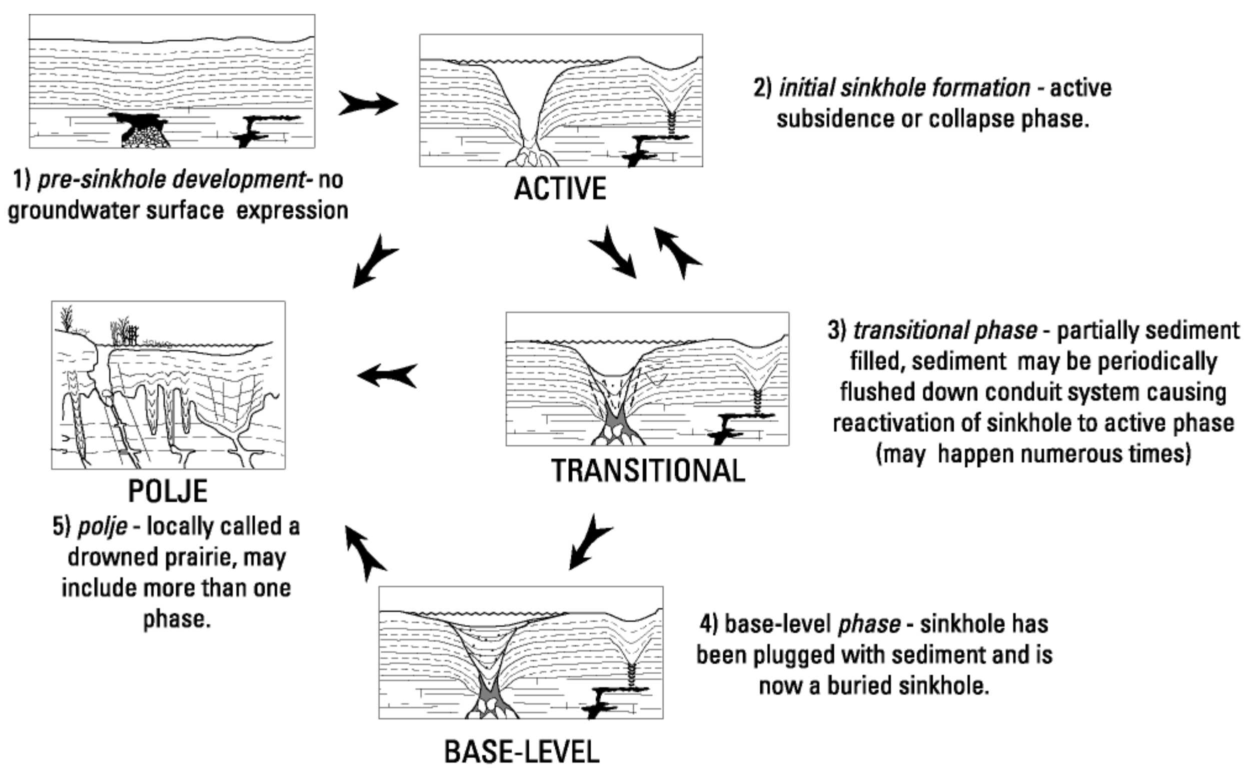


Figure 2. Predicted sinkhole sequential evolution in Florida. Modified from Kindinger and others, 1999.

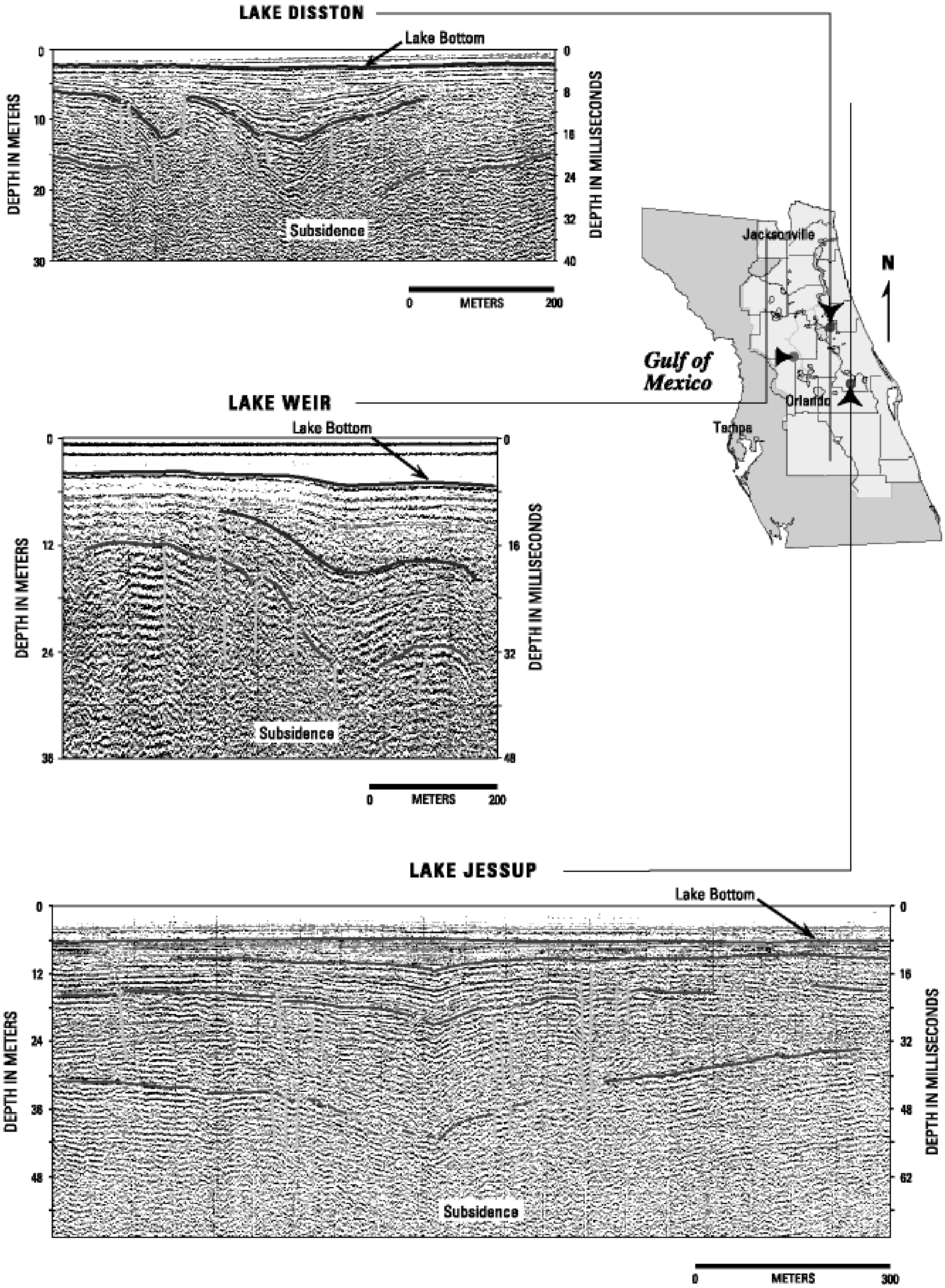


Figure 3. High-resolution seismic profile examples from three lakes located in separate geomorphologic regions of northeastern Florida. Drawn lines are for interpretive purposes and do not indicate correlation between profiles.

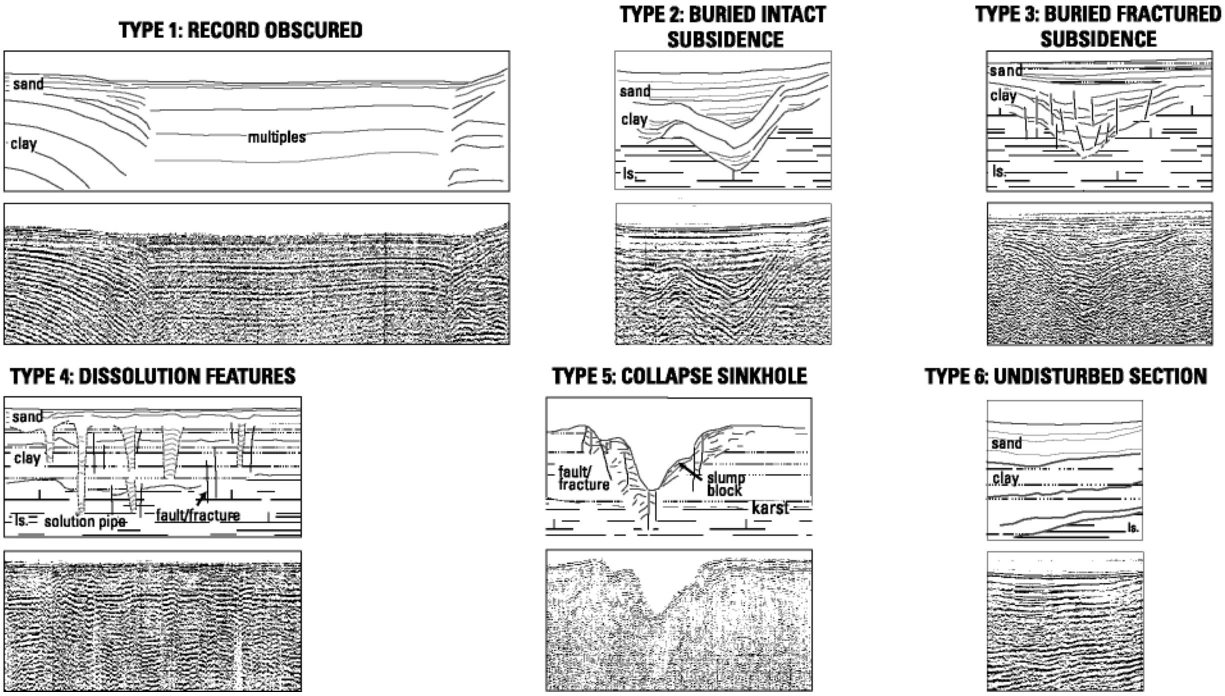

Figure 4. Seismic profiles with line drawing interpretations of six types of features described from the lakes of northeastern Florida.

# Geophysical characterization of pre-Holocene limestone bedrock underlying the Biscayne National Park Reef Tract, Florida

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## Abstract

Shallow seismic investigations of the Pleistocene bedrock beneath the northern Florida Reef tract revealed the presence of a deep channel that is presently infilled with up to18 m of unconsolidated sediments. The channel is located 3 km east of the present shoreline at Elliot Key and cuts across the shelf towards an indented valley-like feature between the offshore bank margin reefs of Long Reef and Triumph Reef. The orientation, slope, and morphology of the channel suggest a fluvial origin, but other explanations such as a collapsed cave system cannot be ruled out. The channel has influenced patch reef formation by limiting reef growth to the elevated margins of the channel and not the channel axis, which is presently covered by sand. It is likely that other similar channels exist, but they can only be identified with further high resolution surveys. The existence of a more extensive channel network in the northern Florida shelf may suggest that fluvial processes during the last glacial were more active than previously thought.

## INTRODUCTION

Surface topography of the Pleistocene bedrock underlying Holocene sediments of the Florida Reef Tract has long been interpreted as primarily depositional in origin with only minor alteration by sub-aerial erosion (Enos, 1977; Lidz and Shinn, 1991). Pleistocene relief features typically run parallel to the modern reef axis and are composed of reef corals (Hoffmeister and Multer, 1968; Shinn et al., 1977). Hoffmeister and Multer (1968) initially suggested that shelf-margin highs were erosional remnants of a once more extensive barrier, but Enos (1977) argued that a depositional origin was more likely based on the shape and orientation of the buttresses.

Incised channels are fairly common in mixed siliciclastic shelf areas such as Belize (Purdy, 1974; Esker et al., 1996) and Australia (Johnson et al., 1982). These channels are often interpreted as erosional in origin when oriented cross shelf. Incised channels are less common in purely carbonate shelf settings. Small channels have been found in south Florida Pleistocene bedrock. Turmel and Swanson (1976) found a 2-3 m deep channel north of Rodriguez Key. Channels between the Keys, such as Caesar's Creek and Angelfish Creek, extend 2-5 m below the surrounding bedrock. In general, these channels have been interpreted as original tidal channels that were located between topographically high reefs during the last interglacial (Warzeski, 1976).

Topographic features in the Key Largo limestone interpreted to have a karst or dissolution origin are mainly isolated small depressions. Enos (1977) identified numerous depressions that extend 2-8 m below the surrounding rock floor surface, which were interpreted as dolines or sinkholes formed by ground-water solution. He noted concentrations of these depressions in certain areas and an orientation to the northeast. Lidz et al. (1997) identified a single, large infilled sinkhole east of Key Largo that measured approximately 50 m wide and over 55 m deep. The sinkhole was interpreted to have been produced by ground water solution and ceiling collapse.

This paper briefly describes topographical features of pre-Holocene limestone bedrock underlying the Biscayne National Park (BNP) reef tract. Biscayne National Park is located on the southeast tip of the Florida Peninsula and encompasses an area that includes the shallow coastal areas of Biscayne Bay and the offshore reef tract. The outer shelf extends from a series of mid-shelf Pleistocene islands, the largest of which is Elliot Key, out to the shelf margin reefs of Pacific, Ajax, Long, and Triumph (Figure 1). Modern shelf communities consist of extensive seagrass beds, sediment banks, sand, and reefs, some of which can reach sea level.

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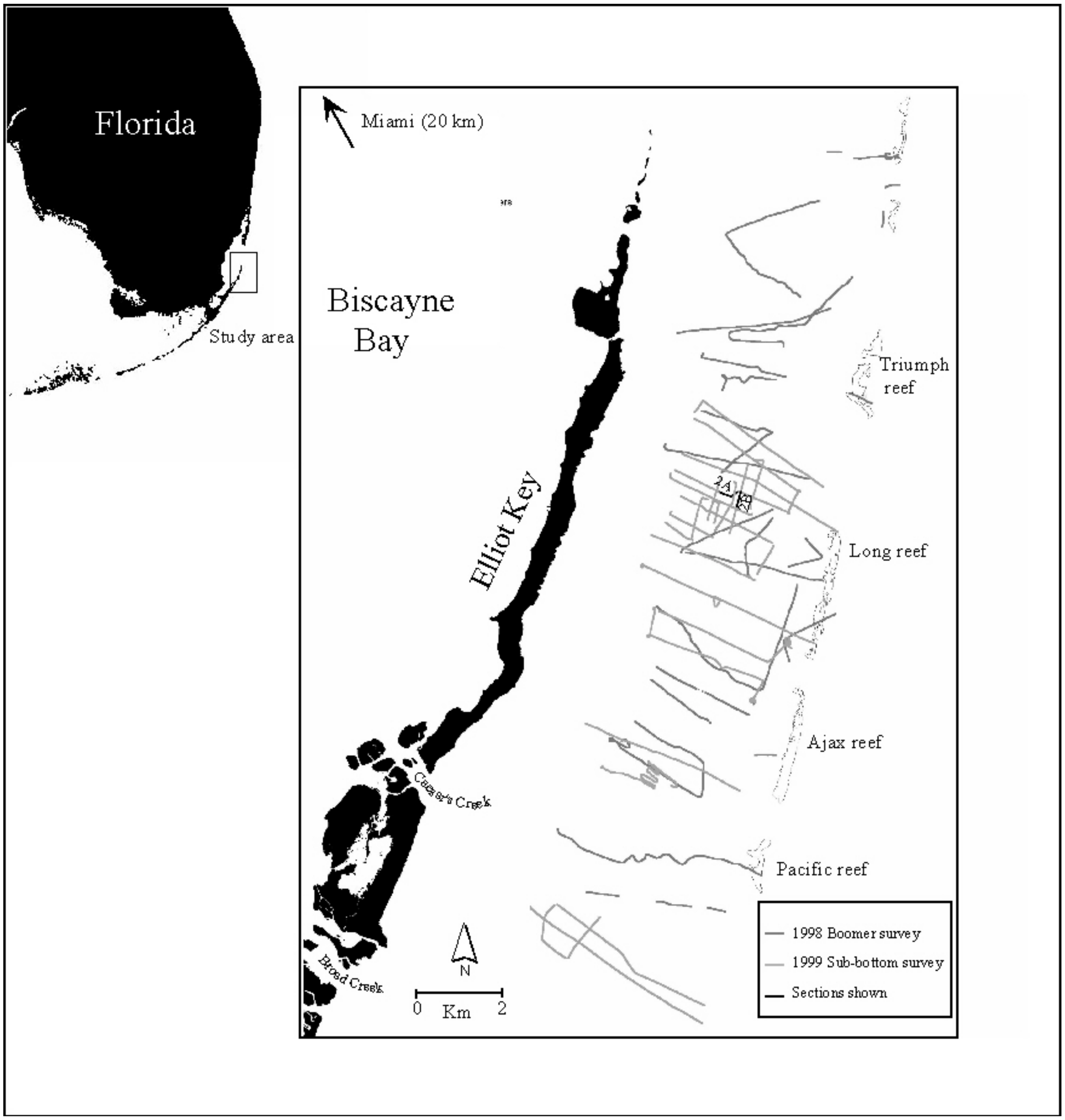


Figure 1 : Location map of south Florida showing seismic data sets from 1998 and 1999 used in this study and position of section profiles shown in figure 2.

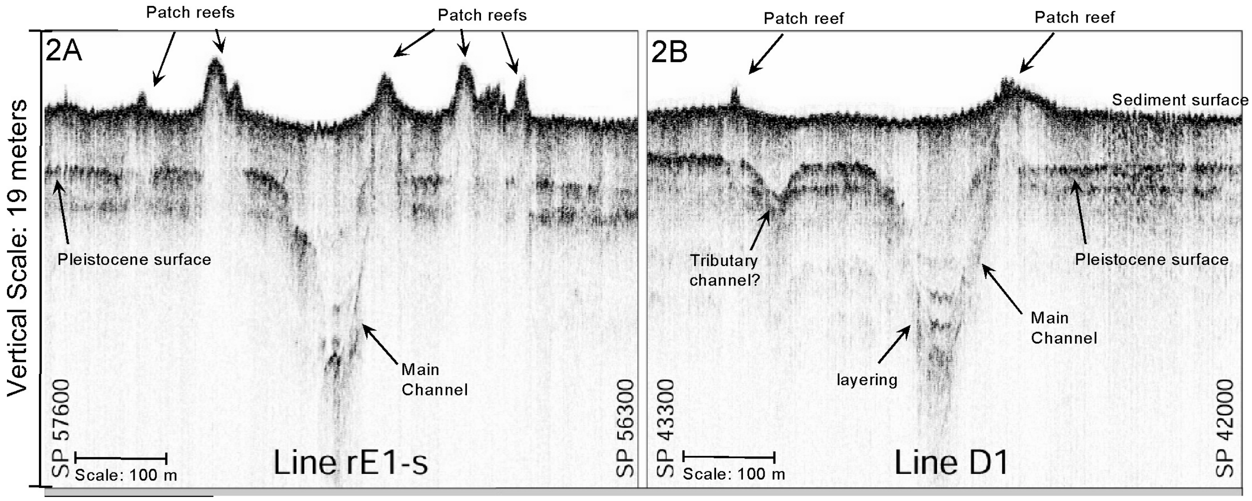


Figure 2: Sub-bottom profiles across two sections of the channel and control on patch reef growth.

The abundance of patch reefs is particularly high in this section of the Florida Reef tract (Marszalek et al., 1977; Shinn et al., 1989). Carbonate sediments deposited during the Holocene transgression cover most of the Pleistocene surface to a thickness up to 14 m (Enos, 1977).

## METHODS

Seismic reflection profiles were recorded during field surveys in Biscayne National Park in September, 1998 and November, 1999. For the first survey, a Geoacoustic boomer source and 4-element streamer were mounted on a shallow draft boat equipped with a 220 volt generator. Lines were shot at 200 joules with a shot interval of 300 ms. These settings allowed features greater than 50 cm to be resolved and penetration to the Holocene/Pleistocene interface, except over well cemented hardbottoms and reefs. A total of 80 km of lines were shot with a spacing between 3 and 5 km covering most of the study area (Figure 1). Raw data were recorded in SEG-Y format and later resampled and post-processed with bandpass (300-2500 Hz) filter and deconvolution filter to further enhance features.

The second survey utilized a sub-bottom profiler (Edgetech model SB-0512) set with a FM frequency pulse between 0.5 and 8 kHz. Resolution of sedimentary and bedrock features was typically 10 cm or better. The sharp impedance contrast between unconsolidated sediments and Pleistocene bedrock was clearly visible but could rarely be seen beneath well-cemented patch reefs and bank-margin reefs. Surveys were focused on areas with the greatest topographical relief features based on the boomer survey. A total of 73 km of lines were shot centered near Elliot Key with line spacing generally less than <1 km (Figure 1). The close spacing of lines was necessary to distinguish a single depression indicative of a doline or sinkhole from networks of connected depressions indicative of a channel.

Positioning for both surveys was achieved using a Trimble GPS receiver with real-time differential correction to +- 5m. The two datasets were merged and the Pleistocene/Holocene interface identified in travel time units. Travel times were converted to approximate depths using a mean acoustic velocity of 1500 m/s. The offset produced by changing water depth, waves, and the depth at which the instruments were being towed was determined by comparing intersections between seismic shot points and NOAA bathymetry point data. Calculated Pleistocene depths were verified by hand probing and overall accuracy was estimated at +/- 0.5m.

## SUMMARY

Surface topography of the Pleistocene surface beneath the BNP reef tract is dominated by small-scale depressions of various sizes with very few positive relief features. Depressions are concentrated on the shelf between -8 and -14 m below mean sea level. They vary from small V-shaped depressions less than 10 m across and 1 m deep, to large U-shaped features more than 200 m across and 10 m deep. Larger scale surveys by Enos (1977) also found that large depressions were more concentrated in Biscayne National Park in comparison to the rest of the Florida Reef Tract. Nearly all depressions are infilled with sediments to the level of the seafloor (Figure 2). Closely spaced reflection profiles around one of the larger depressions detected during the first boomer survey revealed adjacent depressions of similar width and depth oriented in an east-west cross shelf direction. Correlation between depressions suggests they are connected as part of a channel system rather than a network of isolated sink holes. Examination of 1:48,000 color aerial photographs shows traces of the channel expressed as a sand gap that winds between modern patch reefs.

The channel begins near the seaward edge of present day Hawk Channel about 3 km from Elliot Key and winds for at least 3.6 km across the shelf towards an indented valley-like feature between the offshore bank margin reefs of Long Reef and Triumph Reef (Figure 3). The average depth of the main channel is 10 m below the surrounding Pleistocene bedrock and about 22 m below present day sea level. A gradual seaward gradient of 1:1000 in the channel floor is similar to the off-shelf gradient in the adjacent rock floor. Midway up the main channel axis a distinct split occurs into two channels of similar depth and width which diverge to the northeast and southeast, respectively (Figure 3). The two channels appear to end abruptly near Hawk Channel in a near vertical step of nearly 8 meters. It is possible that the channel continues beyond what we have traced, but no deep (>8 m) depressions were identified further inland. Shallow (<2 m deep) depressions detected north and south from of where the main channel ends may be feeder tributaries, but more profiling is necessary to verify a connection to the main channel. The morphology of channel is similar in appearance to modern tidal channels such as Caesar's Creek and the Safety Valves but considerably deeper.

The origin of the channel is somewhat enigmatic and several possibilities will be discussed. A primary depositional origin is not likely since sub-aerial exposure would have rounded and flattened the original channel margins. It is possible that the channel represents a collapsed underground cave system formed by groundwater solution during the last glacial. Groundwater sources would be numerous with Biscayne Bay and the adjacent mainland providing a potential drainage basin. A collapsed cave would result in large limestone blocks infilling the base of the channel and these were not observed in the profiles. Furthermore, the gradual seaward gradient of the channel and winding, branching morphology suggest a fluvial origin, although a sub-surface karst origin can not be ruled out.

The main argument against a fluvial explanation is that the high porosity of the Miami Limestone, lack of tectonic faulting, and flat topography of South Florida all favor slow diffuse flow rather than the channelized flow necessary for fluvial cutting. Development of surface flow in porous limestone requires either cemented calcrete caps or impermeable soil to be present (Ford and Williams, 1989). Laminated calcrete crusts are found capping most Pleistocene sequences in south Florida (Multer and Hoffmeister, 1968). Since these crusts require prolonged periods of sub-aerial exposure to form, it is likely that channelized flow was not developed until the later part of last glacial period.

A fluvial explanation would still require a mechanisms to incise through cemented limestone. It is unlikely that dissolution from meteoric water alone could remove enough limestone to account for the channel. A more likely explanation is that some of the down cutting was caused by mechanical erosion from either small amounts of siliciclastic sand or some other abrasive. An obvious source for such an abrasive sand is not evident at this time. Modern analogues of large coastal rivers in south Florida include the Miami River and Oleta River which drain the eastern Everglades. The fact that the channels of these rivers do cut as deep into the bedrock may simply result from the less pronounced slope in bedrock compared to the outer shelf.

The outer Florida shelf was flooded by rising sea-level approximately 8 kyr ago (Lidz and Shinn, 1991). Patch reef growth is estimated to have started soon after the platform was flooded. The channel influenced the development of patch reefs by limiting where they could nucleate. Seismic sections (Figure 2) clearly show how patch reef growth around the channel is more limited to the margins of the channel and not in the sediment-filled center of the channel. Patch reefs are found along both sides of the channel but are most abundant on the northern bank. This is in agreement with previous studies (Purdy, 1974; Shinn et al.. 1977; Halley et al. 1977) that suggest that reefs cannot form on preexisting topographic depressions, but are favored to grow on topographic highs or irregular surfaces. The complex shelf topography created by incised channels and numerous depressions may also play a role in explaining the astounding abundance of patch reefs in this part of the Florida reef tract.

The presence of a large incised channel in the outer Florida shelf raises the possibility that pre-Holocene fluvial processes may be more important than previously thought. It is likely that other similar channels exist, but they can only be identified with further high resolution surveys. Core drilling is needed to better understand the origin of the channel, timing of formation, and infilling history during the Holocene transgression.

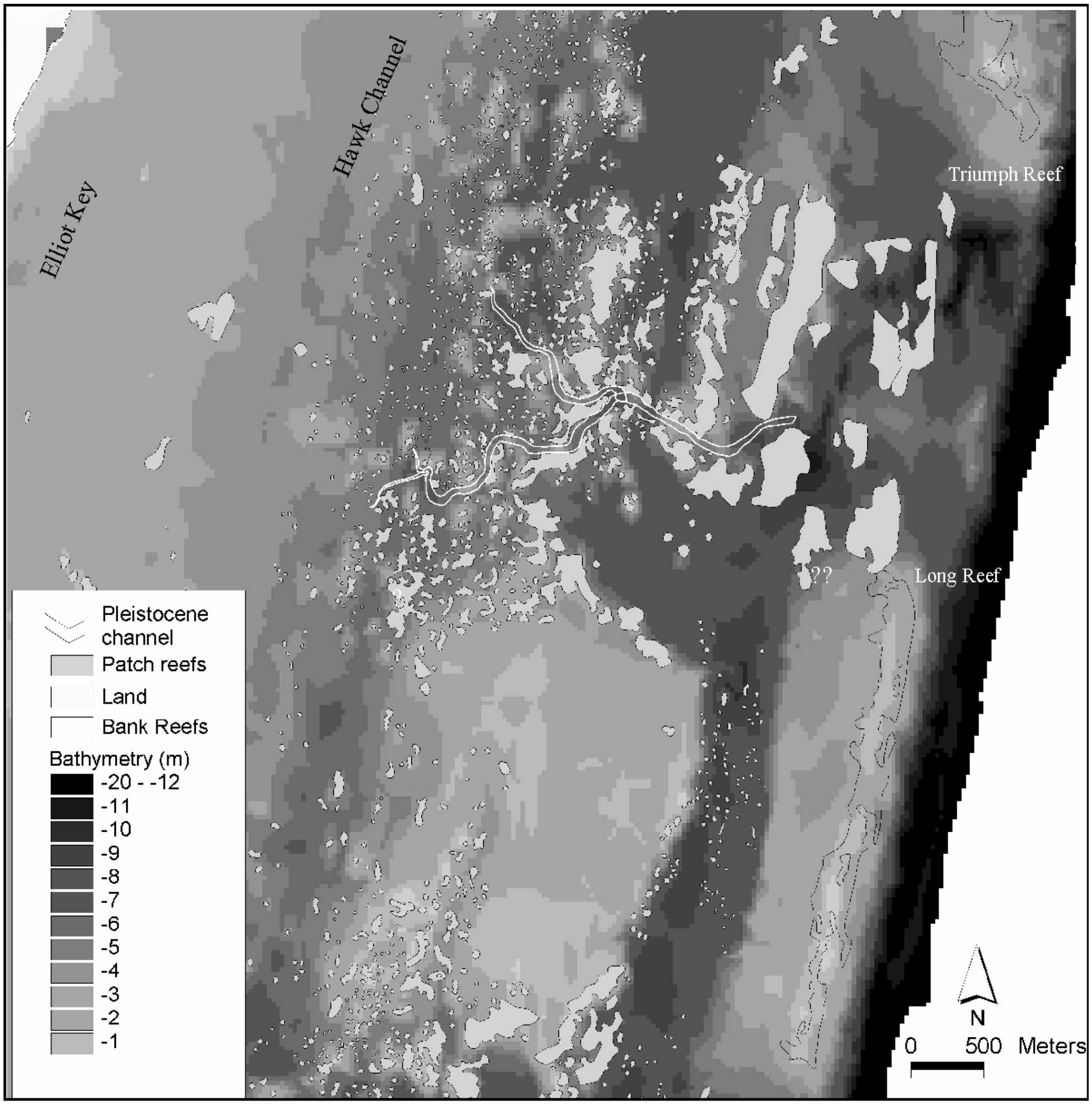


Figure 3: Map showing the location and morphology of the channel superimposed on position of mapped patch reef. Bathymetric data illustrate the location of larger shallow sediment banks and bank-margin reefs.

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# Characterization of the Karstic Biscayne Aquifer in Southeastern Florida Using Ground-Penetrating Radar, Digital Optical Borehole Images and Cores

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## Abstract

Ground-penetrating radar is a useful tool in the characterization of shallow carbonate aquifers. This technique was applied to karstic carbonate rocks of the upper Biscayne aquifer in southeastern Florida. Resultant ground-penetrating radar profiles showed the following geologic features: (1) paleo-sinkholes, (2) paleo-outliers, (3) paleotopographic relief on subaerial exposure surfaces, (4) low-angle accretion bedding and (5) vertical stacking of upward-shallowing cycles. These features were then ground-truthed by comparing them to features recognized in digital optical borehole images, cores, and outcrops. Ground-penetrating radar was used to map a hydrologically important low-permeability layer which, at least in local areas, retards vertical movement of groundwater. This hydrologic layer spans the geologic contact between the Miami Limestone and the underlying Fort Thompson Formation, and includes multiple subaerial-exposure surfaces.

Digital optical borehole images were utilized to orient core samples, “reconstruct” missing core samples, and define vug geometry, bedding, depositional features and grain size in the carbonate rocks of the upper Biscayne aquifer. Features identified on log-derived images were correlated or calibrated to existing core samples; when core samples were absent, images served as a substitute. Other common applications for the digital opitical borehole images include identification and characterization of fractures, depositional features, aquifer architecture, faults, and evaluation of borehole stress.

Using a pixel-counting technique, vuggy porosity was measured for 290 feet of digital optical borehole images. These images were recorded in the upper Biscayne aquifer from 17 coreholes in an approximately 110-square mile area contiguous to the eastern boundary of the Everglades. Quantification of vuggy porosity in another 24 borehole imaging logs is in progress. Analysis of the 17 borehole imaging logs suggests that geologic depositional cycles, rock fabric, and quantity and type of vuggy porosity are all interrelated and that karst-related conduit flow is the principal mechanism of ground-water movement in the upper Biscayne aquifer.

Findings indicate that conduit-flow paths within the Fort Thompson Formation are produced by well-connected, solution-enlarged pore space. Characteristics of the solution-enlarged pore space vary as a result of depositional textures, diagenesis in a meteoric-water system, and vertical position within stacked lithofacies that combine to form each upward-shallowing cycle. Thin, vertical solution pipes are usually associated with tidal-flat deposits that commonly cap cycles and contain a low permeability matrix. These pipes provide a network of passageways for vertical ground-water flow across the low-permeability cycle caps. Middle portions of cycles are relatively non-vuggy. Well-connected pelecypod molds and irregular vugs are mainly in the lower portion of cycles. Horizontal conduit flow appears to be largely within the vuggy porosity at the base of each cycle.

Dividing the Biscayne aquifer into small-scale, time-bounded cycles has facilitated accurate assessment of depositional and diagenetic facies and will form the basis of a high-resolution aquifer conceptual model. This conceptual model, which is suitable for numerical modeling, will show the distribution of porosity and hydraulic conductivity. Because of the direct hydraulic connection between surface-water features and the aquifer, a thorough understanding of the Biscayne aquifer and its characteristics is critical to properly managing water levels within existing and planned surface-water reservoirs in the Everglades.

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# Geophysical Investigations of Upward Migrating Saline Water from the Lower to Upper Floridan Aquifer, Central Indian River Region, Florida

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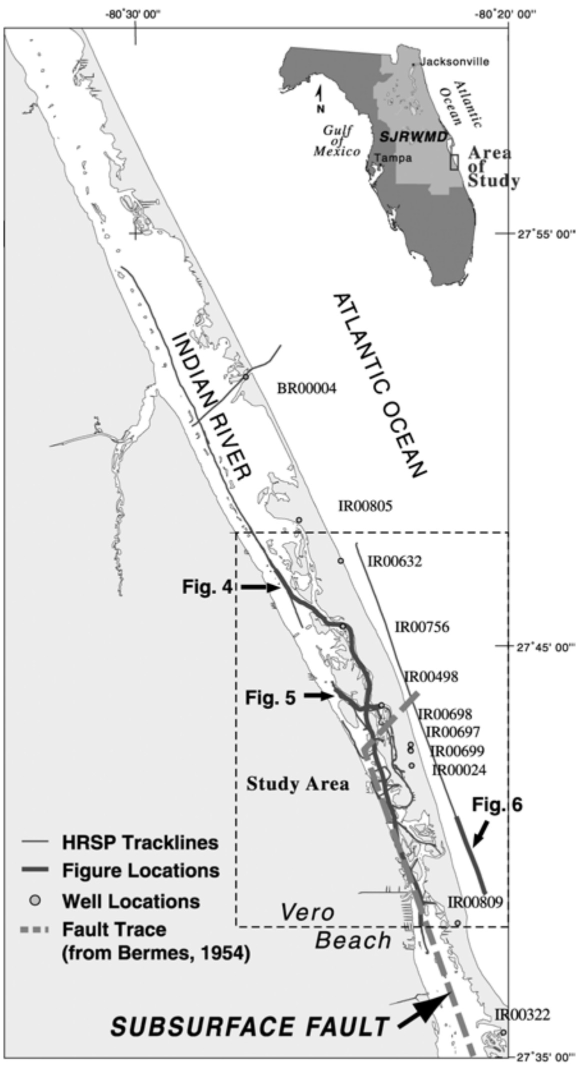


Fig. 1. Location of survey area.

Indian River Lagoon (IRL) extends approximately 250 km along the east central coast of Florida and consists of three interconnected basins: Mosquito, Banana River, and Indian River. The study area for this report is 75 square-km in the middle section of IRL (Fig. 1). The geology below the lagoon is of interest to researchers. Subsurface features are believed to control the hydrology, and the concern that prompted this investigation is that possible faulting at depth has provided a pathway for the upward migration of saline water from the Lower Floridan to the potable supply within the Upper Floridan Aquifer. Data collected for the study includes remote sensing of subsurface features using High-Resolution Single-Channel Seismic Profiles (HRSP). The HRSP was correlated to gamma-ray intensity profiles from well logs. The subsurface features identified in this study are believed to control the hydrology and have geomorphologic expression on the surface.

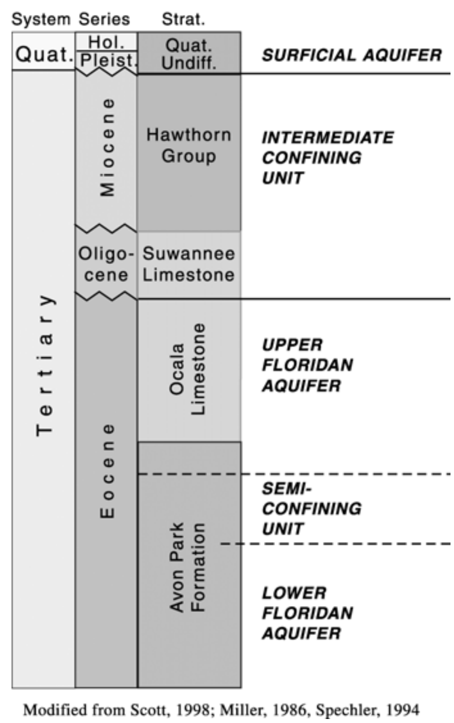


Fig. 2. Generalized stratigraphic column for the area.

The nearsurface geology can be divided into three main lithologic units: (1) Quaternary sands and clay of the surficial units; (2) variable clays, sands, and carbonates within the Hawthorn Group (Miocene), and; (3) Eocene limestones, which include the Ocala and Avon Park (Fig. 2). These units constrain the hydrology. The interbedded sands and clays of the

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Hawthorn Group provide the Intermediate-confining unit, separating the Upper Floridan Aquifer from the Surficial Aquifer (Fig. 2). At depth, the upper Avon Park Formation acts as a semi-confining unit, separating the fresh water Upper Floridan from the more saline Lower Floridan Aquifer. Discontinuities within these confining units may affect the regional hydrology by allowing leakance between the aquifers.

Profiles of gamma-logs provide a trend of the lithologic units roughly parallel to the barrier island system (Fig. 3). Gamma radiation is a product of naturally occurring radioactive material in the sediments, and is prevalent in the clays and silts of the Hawthorn Group. Peaks in the gamma intensity show a thickening of the Hawthorn Group southward, with a relatively consistent upper horizon and the lower horizon dropping away. This drop led to the postulation by Bermes [1958] that a displacement fault exists through the region. Subsequently, Schiner and others [1988], included water quality parameters to delineate the fault zone. The fault was reported to strike parallel with the lagoon in a NNW direction and turns NE towards the Atlantic Ocean in the center of the study area (Fig. 1). Water samples taken from Upper Floridan aquifer wells located east of the fault had chloride concentrations between 1,400 to 2,900 parts-per-million (ppm). Upper Floridan aquifer wells that were sampled to the north and west of the fault had chloride concentrations of < 700 ppm. The abrupt change in chloride concentration between wells suggest that the fault may provide a pathway for upward migration of saline water. The drop in elevation of the Ocala Limestone (Fig. 3) may delineate the location of the fault. The relatively consistent upper horizon of the Hawthorn Group indicates that any displacement that may have occurred had terminated during the late Tertiary.

Apart from well logs, subsurface data in the Indian River region is scarce due to the difficulty of sampling, both directly and remotely, in terrain that includes shallow water bodies, unconsolidated sediments and rock. HRSP was originally developed for offshore work, but is becoming increasingly more common in coastal and inland areas due to recent developments in the technology. Complete descriptions of the methods used in this study can be found in Kindinger and others [2000]. In 1997, approximately 82 km of seismic profiles were collected from the central IRL region and adjacent offshore areas (Fig. 1). Elsewhere on the Florida Platform, single-plate HRSP has proved adequate in imaging the subsurface, down to the uppermost Ocala Limestone [Kindinger and others, 2000, Wolansky and others, 1983]. Since the limestone lacks sufficient velocity contrasts such as bedding planes or lithologic variability, few internal reflectors

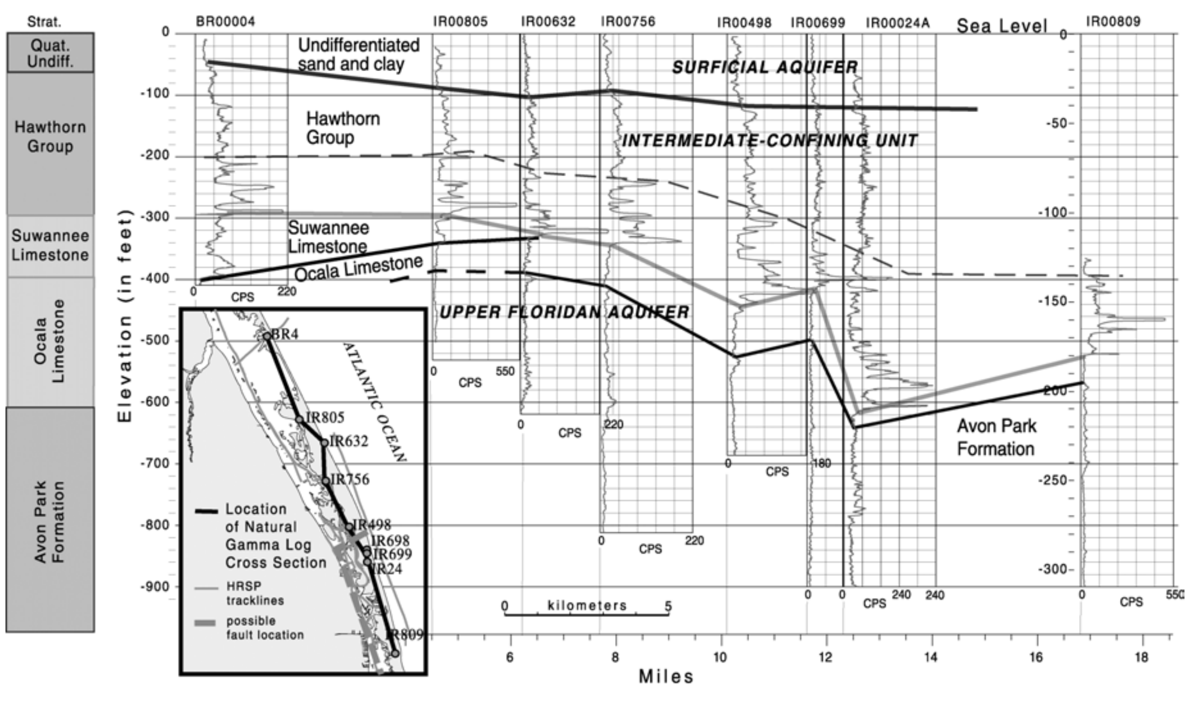


Fig. 3. Natural gamma log profiles from the study area (see inset map).

can be identified. At depth, increased acoustic noise obscures geologic structure. In the Indian River region, signal quality was generally good, even below depths interpreted by gamma logs to be within the Ocala and Avon Park limestones. Correlation between HRSP and gamma logs was accomplished by comparing time to depth of a distinctive horizon in both profiles, in this case the top of the Ocala Limestone (Figs. 3 & 4). The resulting velocity was calculated to be 1,955 meters per second between 100 and 200 meters depth. This velocity compares to similar velocities obtained elsewhere in Florida [Sacks and others, 1991, Weiner, 1982].

Once correlation of HRSP to the geology was accomplished, interpretation of the data could proceed. Figures 4, 5 and 6 are HRSP which show strong consistent reflectors across the profile. These reflectors are interpreted to be the bounding reflectors of the Hawthorn Group, Ocala Limestone and Avon Park Formation, with a few internal or intermediate reflectors. These reflectors are in close agreement with depth and attitude to the gamma log profiles, especially when considering figures 3 and 4, which run parallel to each other throughout the study area. Smaller, localized features are also present (Figs. 4 & 6) and are interpreted to be solution or collapse features. These features are important because they may represent localized breaches in the confining units, providing pathways for fluid migration between aquifers. Other features noted in the HRSP include intermediate reflectors within the Hawthorn Group separating to the south. Also, truncation of reflectors and small infilled channel-like features along the top of the Hawthorn suggest of a flooding surface (Figs. 4 & 5).

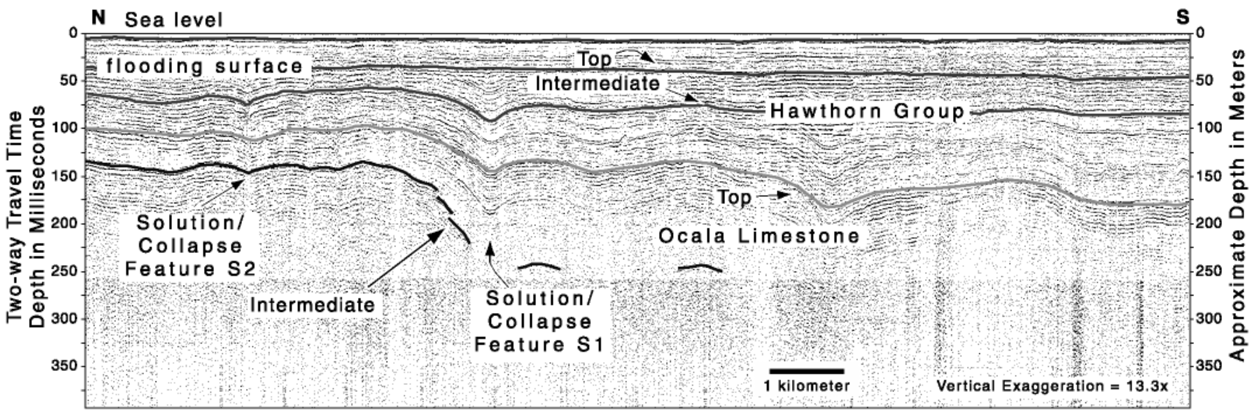


Fig. 4. HRSP with interpretations, see Figure 1 for location.

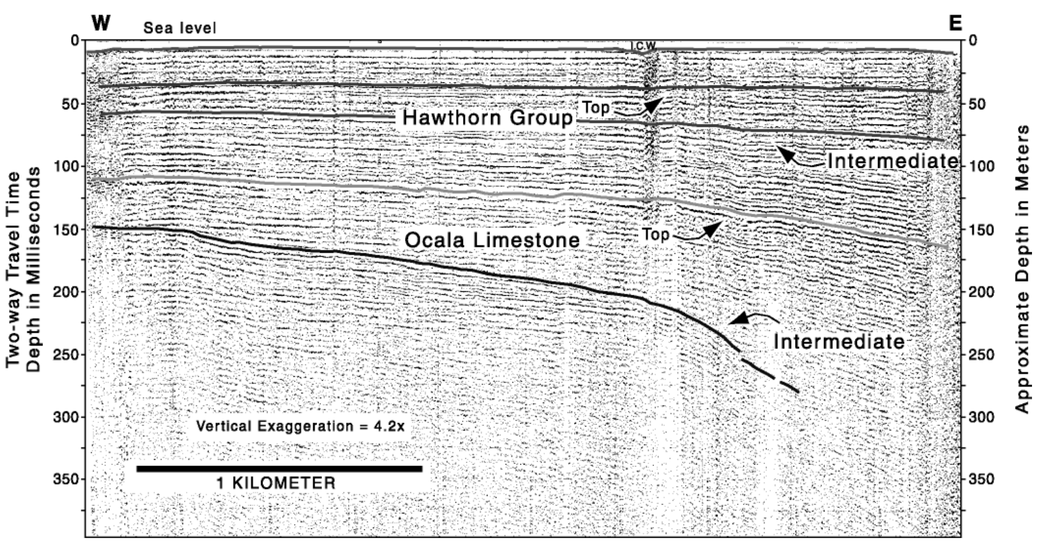


Fig. 5. HRSP with interpretation, see Figure 1 for location.

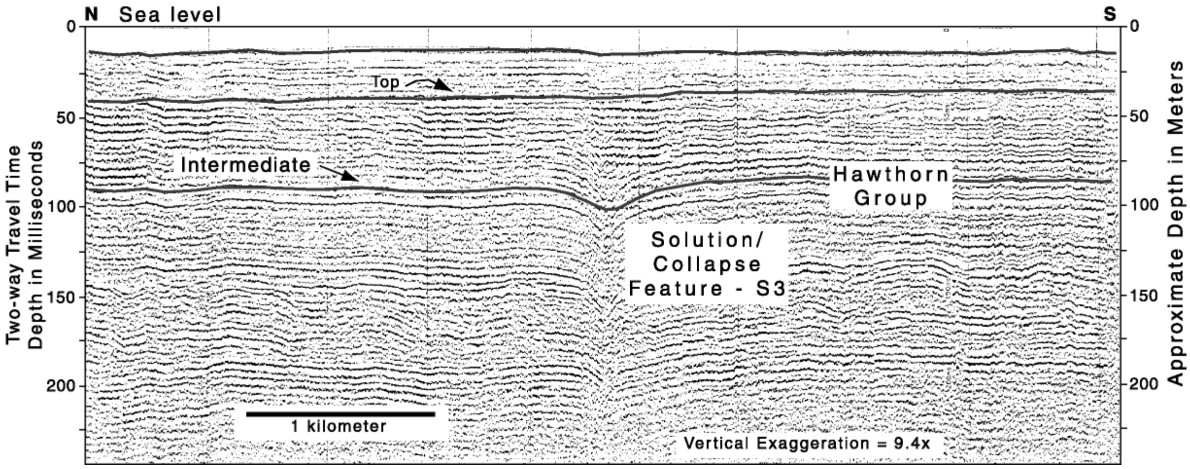


Fig. 6. HRSP with interpretation, see Figure 1 for location.

As with the gamma profiles, the most notable, large-scale feature noticed from the HRSP is the dramatic drop in the Ocala and Avon Park (?) reflectors, that eventually descend beyond resolution depth. Locating this drop across the HRSP tracklines (Fig. 7) indicates the feature strikes along the IRL, throughout the length of the study area.

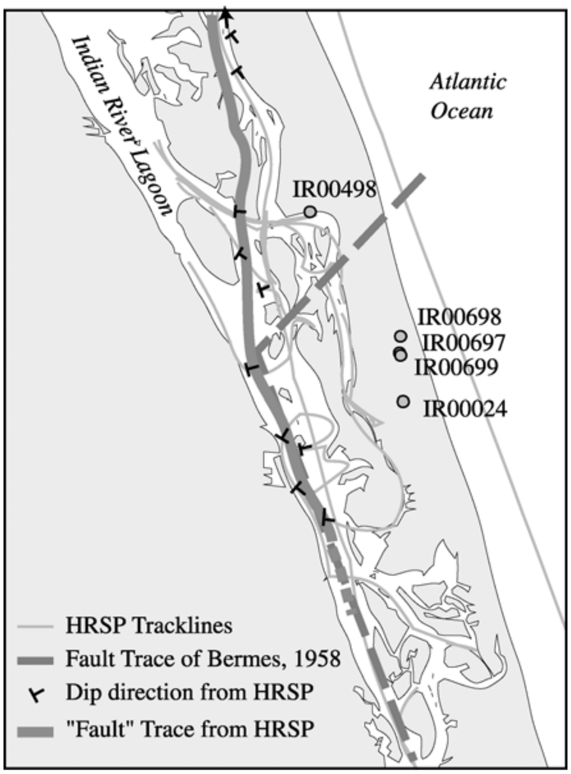


Fig. 7. Location of displacement features.

This strike is parallel to the fault trace postulated in the earlier studies, although the turn offshore is not apparent. The reflectors appear to downwarp (Figs. 4 & 5), and no obvious displacement along reflectors is visible. Unless the actual fault is too deep to resolve by the available HRSP, the acoustic pattern is more characteristic of common karst systems, such as dissolution and subsidence. Characterization of subsurface geology in karst systems elsewhere in Florida shows that these features are also capable of developing breaches in confining units and do not necessarily need to be related to large scale faulting [Kindinger and others, 2000].

Since the HRSP is capable of high-density regional coverage, reflectors can be digitized relative to geographic position, and contour structure maps can be produced (Fig. 8). The maps confirm that the most dramatic relief is in the deepest reflectors, which corresponds to the tops of the Ocala Limestone and Avon Park Formation. Although the Hawthorn Group thickens to accommodate this drop, the upper horizon remains relatively level. An isopach map of the Hawthorn Group shows that the unit widens from northwest to southeast (Fig 8).

A stratigraphic model to describe the data is suggested in the following four steps: 1.) pre-Miocene subsidence in the area; 2.) erosion of a headland to the northwest and deposition into the area occurred during Miocene times (Hawthorn Group) and possibly later; 3.) pre-Quaternary sea-level rise levels the upper Hawthorn Group and infills any channeling or small depressions, and; 4.) continued dissolution at depth from Tertiary to present has created localized subsidence features throughout the Hawthorn and possibly in more recent sediments. The large subsidence potentially controls the regional

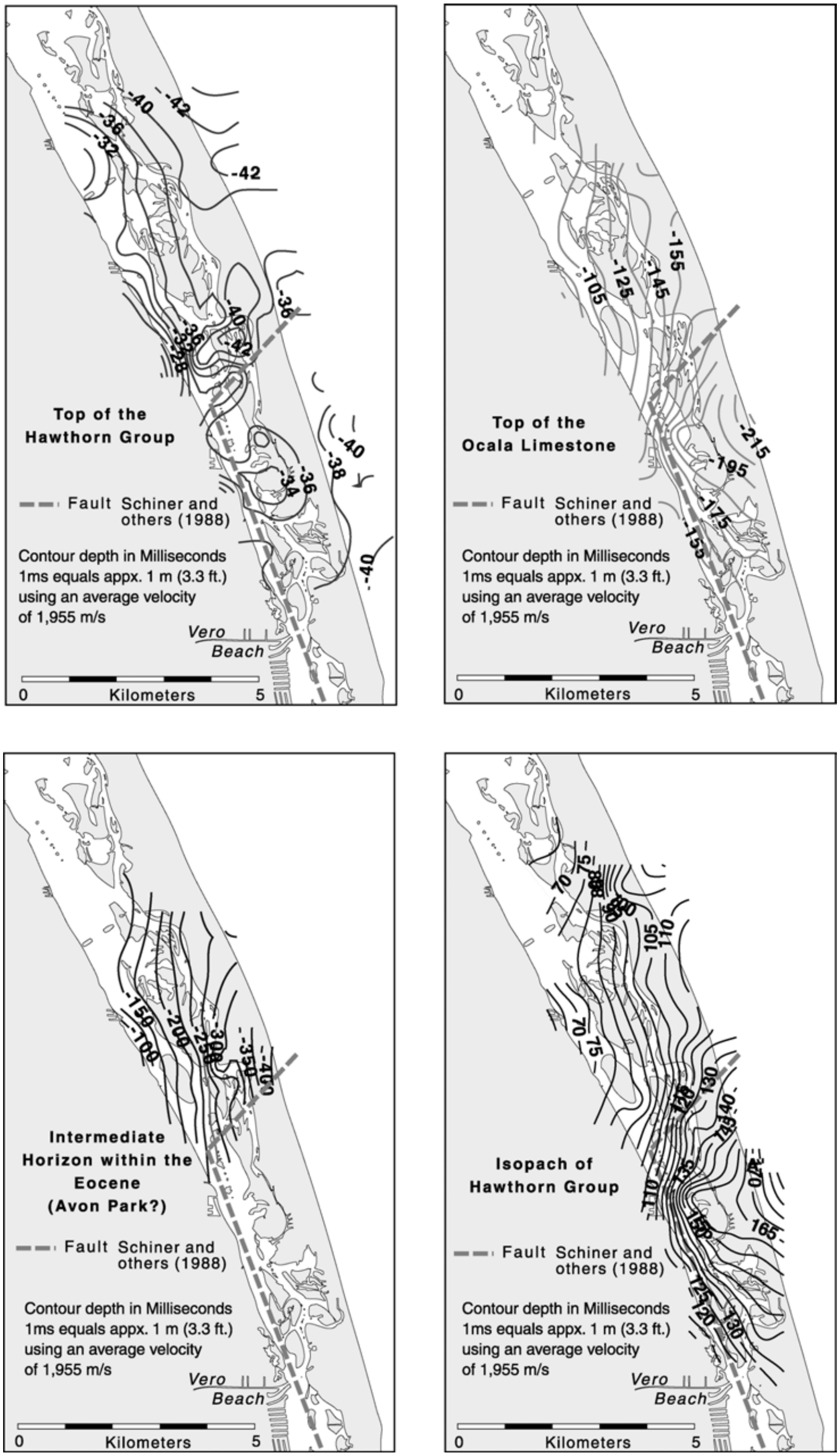


Figure 8. Structure contour and Isopach maps of horizons interpreted from HRSP.

hydrology. Possible stress fractures or slumping along the periphery of the subsidence could produce breaches in confining layers and pathways for fluid migration. The smaller subsidence (piping) features associated with continued dissolution at depth could provide similar conduits.

Migration of saline waters from the Lower to Upper Floridan Aquifer can occur within the Avon Park Formation. Since the available HRSP is not resolvable into the Avon Park Formation, the presence of a fault at that depth cannot be ruled out. Since no displacement along reflectors is evident in the shallower reflectors, the influence of a deep fault on shallow stratigraphy would be through structural deformation of the geology as it accommodates the drop across the fault trace. However, the structural deformation reflected in the HRSP can also be created by karst-related dissolution and subsidence within the limestone units, without regional faulting. Subsequent dissolution can further compromise the confining units and continue to enhance migratory pathways.

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# Seasonal and Short-Term Variability in Chlorinated Solvent Concentrations in Two Karst Springs in Middle Tennessee: Implications for Sampling Design

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## Abstract

The U.S Geological Survey in cooperation with the Tennessee Department of Environment and Conservation, Division of Superfund is evaluating volatile organic compound sampling strategies for karst springs. Water-quality signatures for two contaminated karst springs in Middle Tennessee were determined by continuously monitoring selected water-quality properties through one complete hydrologic cycle. High-frequency flow-controlled sampling was used to characterize chlorinated solvent concentrations. The results indicate that discharge from and contaminant concentrations in Wilson Spring near Lewisburg, Tennessee, are flashy as indicated by rapid fluctuations in water quality and discharge in response to storm events. During one storm event, chloroform concentrations in discharge from this spring increased to 33 milligrams per liter, which represents a 60-fold increase over the concentration before the storm. The increased chloroform concentrations were correlated with a decrease in specific conductance and peak discharge. The maximum chloroform concentrations detected during this storm, represents a sixfold increase over the maximum concentrations measured at any other time during the study period. Big Spring at Rutledge Falls shows little change in water-quality properties indicating a large recharge area and diffuse-flow aquifer conditions. Results from the study demonstrate the need to develop site-specific sampling strategies for karst springs. Results also indicate that evaluating water-quality signatures of karst springs is helpful in designing a site-specific sampling strategy.

## INTRODUCTION

Karst springs are important sampling points that are commonly used as monitoring locations to collect water-quality data at contaminated sites in karst hydrogeologic regimes (Quinlan and Ewers, 1985). Abrupt changes in water quality in karst springs in response to rainfall events are well documented (Hess and White, 1988, Quinlan and Alexander, 1987; Dreiss, 1989; Brown and Ewers, 1991; Ryan and Meiman, 1996). Quinlan and Alexander (1987) published guidelines for water-quality sampling frequency in karst terranes. The importance of high-frequency flow-dependent sampling in accurately determining concentrations of fecal bacteria, chloride, and pesticides in water from a karst spring has been documented (Meiman, 1991). Cretella (1985) showed that volatile organic compound (VOC) concentrations in a karst spring could vary as much as tenfold throughout a storm cycle. Despite these findings, little consideration has been given to the design of sampling programs for monitoring water quality in karst springs. Quarterly and semi-annual samplings are still conducted at many contaminated karst sites.

Depending on the maturity of karst development and the nature of a rainfall event, each spring has a distinctive water-quality and discharge signature (Quinlan and Ewers, 1985). Hydrographs of continuous water-quality data are needed to reveal these signatures. The U.S. Geological Survey, in cooperation with the Tennessee Department of Environment and Conservation, Division of Superfund, is studying three contaminated springs in karst areas of Middle Tennessee to evaluate VOC sampling strategies. The springs were continually monitored for specific water-quality properties; and during selected storm events, high-frequency sampling was conducted for VOC analysis. This report presents preliminary data from two of these springs, Wilson Spring near Lewisburg, Tennessee, and Big Spring at Rutledge Falls near Tullahoma, Tennessee.

## DESCRIPTION AND HISTORY OF WILSON SPRING

Wilson Spring is located in the Central Basin karst region of Tennessee (fig. 1), as described by Wolfe and others (1997). The hydrogeology of the Central Basin is characterized by ground-water flow in massive Ordovician limestones. This type of massive limestone alternates through out the geologic section with thin-bedded, shaly, limestones (Farmer and Hollyday, 1999). Uplift of the Nashville Dome has resulted in the development of extensive fracturing in this karst region. Dissolution of the limestone has enlarged these fractures resulting in karst development in the massive

In Eve L. Kuniansky, editor, 2001, U.S. Geological Survey Karst Interest Group Proceedings, Water-Resources Investigations Report 01-4011, p. 141-149

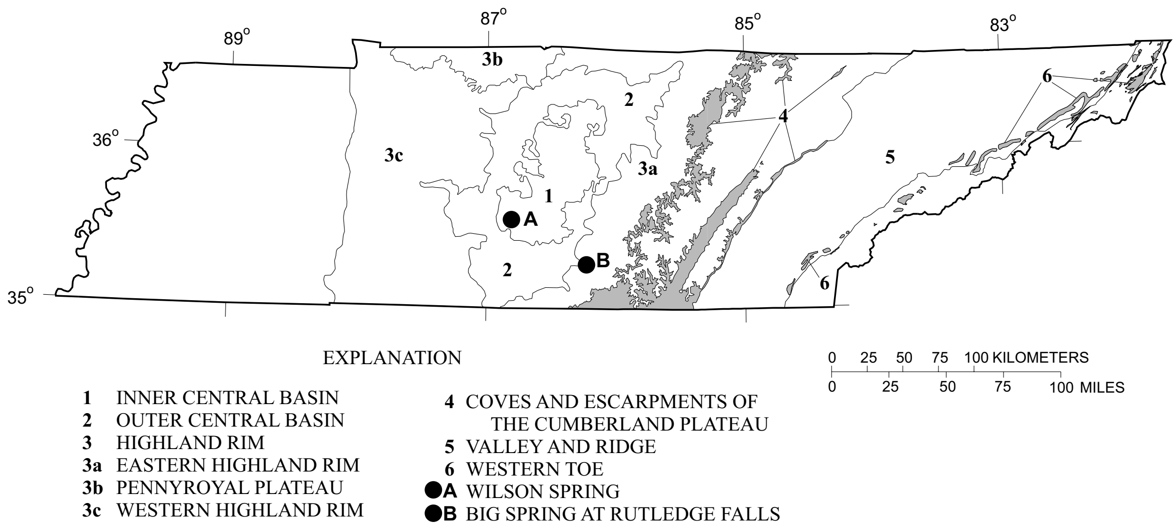


**Figure 1.** Location of study springs and karst regions of Tennessee. (Modified from Wolfe and others, 1997.)

formations. The thin-bedded shaly formations generally act as confining units. The thin-bedded Ordovician-age Lebanon Limestone caps the hills in the study area and retards the downward movement of water. Surface streams that run off the Lebanon Limestone onto the massive Ridley Limestone (Ordovician age) can sink into the upper Ridley karst aquifer as described by Crawford and Ulmer (1994). A 10-foot-thick thin-bedded unit is present within the Ridley Limestone approximately 100 feet below the stratigraphic top of this unit (Wilson, 1990). The thin-bedded unit restricts downward flow, and cave streams are developed on the top of this unit. Wilson Spring is the surface discharge point of one of these cave streams (Crawford and Ulmer, 1994). Land use in the area is predominantly agricultural.

In October 1990, a train derailment near Wilson Spring resulted in the release of more than 15,000 gallons of chloroform, a dense nonaqueous phase liquid. The chloroform sank into the upper Ridley aquifer at the site. According to Crawford and Ulmer (1994), the chloroform pooled on top of the thin-bedded unit of the Ridley Limestone, and then moved southwest downdip along weathered bedding planes until being trapped by low permeability, less weathered rock. Water containing chloroform is transported along the strike of the bedding planes through a cave stream southeast to Wilson Spring. Since 1992, Ogden Environmental and Energy Services (a consultant for the railroad) have been continuously monitoring discharge and rainfall at the site and analyzing samples on a monthly bases for VOC's. Based on this monitoring, chloroform concentrations typically range from 1 to 5 milligrams per liter (mg/L) seasonally. Water from the spring is impounded and treated before being released into Big Rock Creek.

## DESCRIPTION AND HISTORY OF BIG SPRING

Big Spring at Rutledge Falls is located in Middle Tennessee on the escarpment of the Highland Rim (fig. 1). The spring discharges approximately 3.5 cubic feet per second ($ft^3/s$) from the Manchester aquifer into Crumpton Creek. The discharge occurs near the contact between the Late Devonian-age and Early Mississippian-age Chattanooga Shale and the overlying Mississippian-age Fort Payne Formation. The Chattanooga Shale ranges from 20 to 30 feet in thickness and is generally considered to be a major confining unit in Tennessee (Burchett, 1977). The Fort Payne Formation ranges from 20 to 230 feet in thickness and is predominantly soluble dolomitic limestone. The Manchester aquifer is a regional aquifer comprised of gravel in the residuum of the upper part of the Fort Payne Formation and solution openings in the bedrock of the Fort Payne Formation (Burchett and Hollyday, 1974). Numerous springs and seeps are present along the Highland Rim where the contact between the Fort Payne Formation and the Chattanooga Shale crops out. Land use in the area of the spring is mixed residential and agriculture. Chlorinated solvents including perchloroethylene (PCE) and trichloroethylene (TCE) have been detected in water samples collected from the spring at concentrations of approximately 2 and 9 micrograms per liter (μg/L), respectively. The general direction of ground-water

In Eve L. Kuniansky, editor, 2001, U.S. Geological Survey Karst Interest Group Proceedings, Water-Resources Investigations Report 01-4011, p. 141-149

flow in the area near the spring is from the southeast (Mahoney and Robinson, 1993).

## METHODS

In February 2000, gaging stations were established at both springs according to the procedures described by Carter and Davidian (1982). Continuous stage-recorders described by Buchanan and Somers (1968) were used to collect stage data in 0.01-foot increments at 15-minute intervals. On October 1, 2000, the recording interval was changed to 10 minutes at Wilson Spring to correspond with frequency of data collection by the consultant for the railroad. Discharge measurements were made at both springs for a range of stages according to procedures described by Buchanan and Somers (1969), and discharge ratings were developed following procedures described by Kennedy (1983). These ratings were applied to the continuous-stage data to produce continuous-discharge records using techniques described by Kennedy (1983).

Field measurements of water quality were made at both springs according to the general procedures described by Wood (1976) and Wilde and Radtke (1998). Field measurements of specific conductance, pH, dissolved oxygen (DO), and water temperature were made using a Hydrolab DataSonde 4. At Wilson Spring, the instrument was placed in a tub just below the lip of a flume because of the shallow water depth inside the flume. During field visits, a specific conductance and temperature meter was used to compare measurements made in the tub with measurements in the flume. Measurements were made at 15-minute intervals at both springs until October 1, 2000, when Wilson Spring measurements were changed to 10-minute intervals to coincide with the consultant's data-collection interval. Instruments were calibrated before deployment by using standard reference solutions and following the procedures recommended by the manufacturer (Hydrolab Corporation, 1997). At approximately 3-week intervals, data were downloaded from the instruments and calibration was checked. Recalibration was performed as necessary. Rainfall data collected at Wilson Spring were obtained from the consultants for the railroad.

VOC samples were collected and processed using methods described by Wilde and others (1998a and 1998b). Forty-milliliter VOC vials were dipped in the water body by hand to collect grab samples at both springs. At Wilson Spring during a storm event in November 2000, an ISCO model 6100 automatic VOC sampler was used to collect samples at 1-hour intervals from the tub. Several grab samples were collected during this storm event to verify the accuracy of the results obtained from the automatically collected samples.

Samples were analyzed for VOC's using a Scentograph PLUS II portable gas chromatograph (GC) equipped with a purge-and-trap unit. Purge-and-trap procedures similar to those described by U.S. Environmental Protection Agency (U.S. EPA) Method 5030B (1992) were used. Periodic duplicate samples were collected and sent to the U.S. Geological Survey National Water Quality Laboratory (NWQL) in Denver, Colorado, to validate the results obtained using the portable GC.

## DISCUSSION

Continuous water-quality data collected at Big Spring show little variation from February through October 2000 (fig. 2). Storm events that occurred during the study period appeared to have only slight effects on measured water-quality properties. Gage height showed little variation, therefore, a rating curve was not established. Temperature, DO, and pH remained constant with mean values of approximately 14 $^{o}$C, 8 mg/L, and 7.0, respectively. Specific conductance averaged about 150 microsiemens per centimeter (µS/cm), with a slight increase in the summer and fall months (fig. 2). TCE concentrations reflect a similar pattern of little variation, with values ranging between 7 and 10 µg/L (fig. 3). The flashy water-quality signature (characterized by large variations in concentrations of chemical constituents and values of physical properties) normally associated with conduit-type karst springs is not present at Big Spring. The constant discharge implies that this spring discharges from an aquifer with a large recharge area, a large amount of storage, and in which diffuse flow is dominant-all characteristics that correlate with the Manchester aquifer.

In contrast, Wilson Spring has a flashy water-quality signature. During spring the 2000 storm season, storm events resulted in dramatic changes in discharge and specific conductance, and smaller fluctuations in pH and temperature. During storms, discharge increased from less than 0.1 ft$^3$/s to greater than 5 ft$^3$/s and specific conductance commonly decreased from a high of about 550 to as low as 100 µS/cm (fig. 4). The higher pre-storm specific conductance may be attributed to a long residence time of water in the aquifer. This long residence time allows for increased dissolution of the limestone resulting in an increased ionic content and resultant specific conductance of the water. The lower specific conductance during the storm event is attributed to the rapid influx of water with ionic content that has had only a short period of

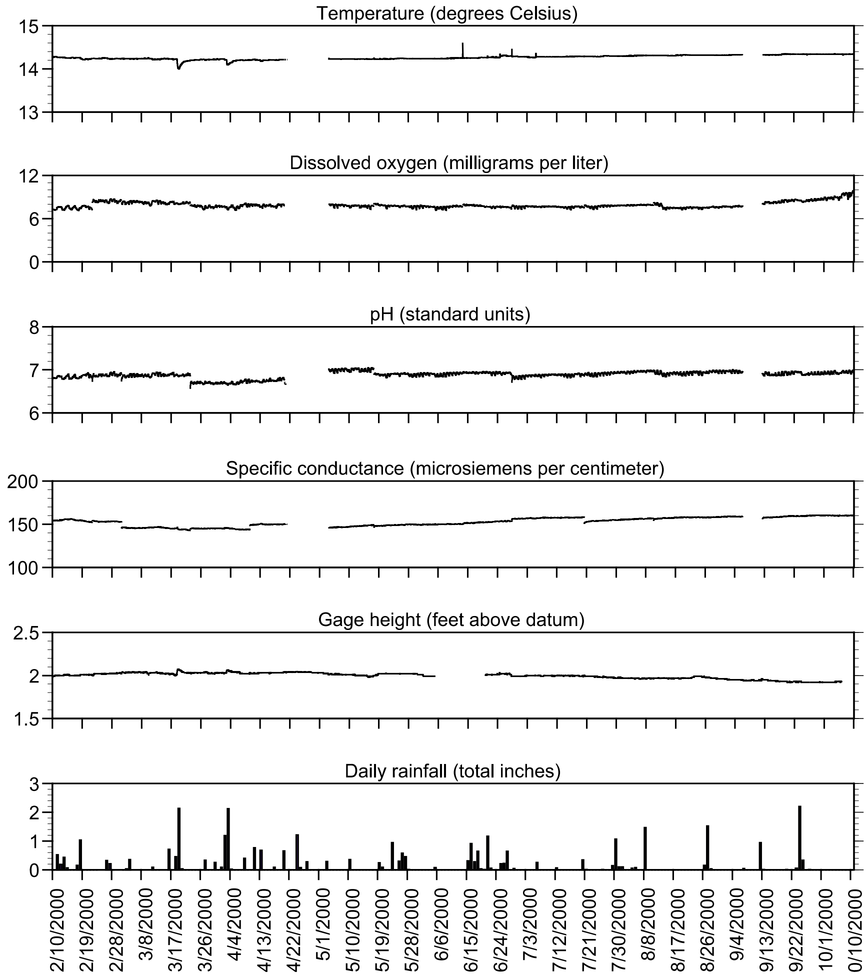


**Figure 2.** Continuous water-quality and gage-height data from Big Spring at Rutledge Falls at 15-minute intervals, Feb. 10 through Oct. 10, 2000. Daily rainfall data from U.S. Geological Survey gaging station located 19 miles west-northwest of Big Spring. (Line gaps indicate missing data.)

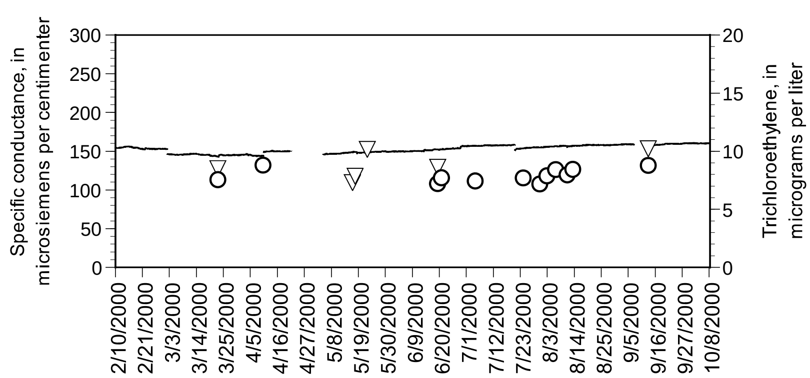


**Figure 3.** Trichloroethylene and specific conductance data from Big Spring at Rutledge Falls, Feb. 10 through Oct. 8, 2000. Water samples were analyzed by the U.S. Geological Survey National Water Quality Laboratory (USGS NWQL) or using a portable gas chromatograph (GC). (Line gaps indicate missing data.)

In Eve L. Kuniansky, editor, 2001, U.S. Geological Survey Karst Interest Group Proceedings, Water-Resources Investigations Report 01-4011, p. 141-149

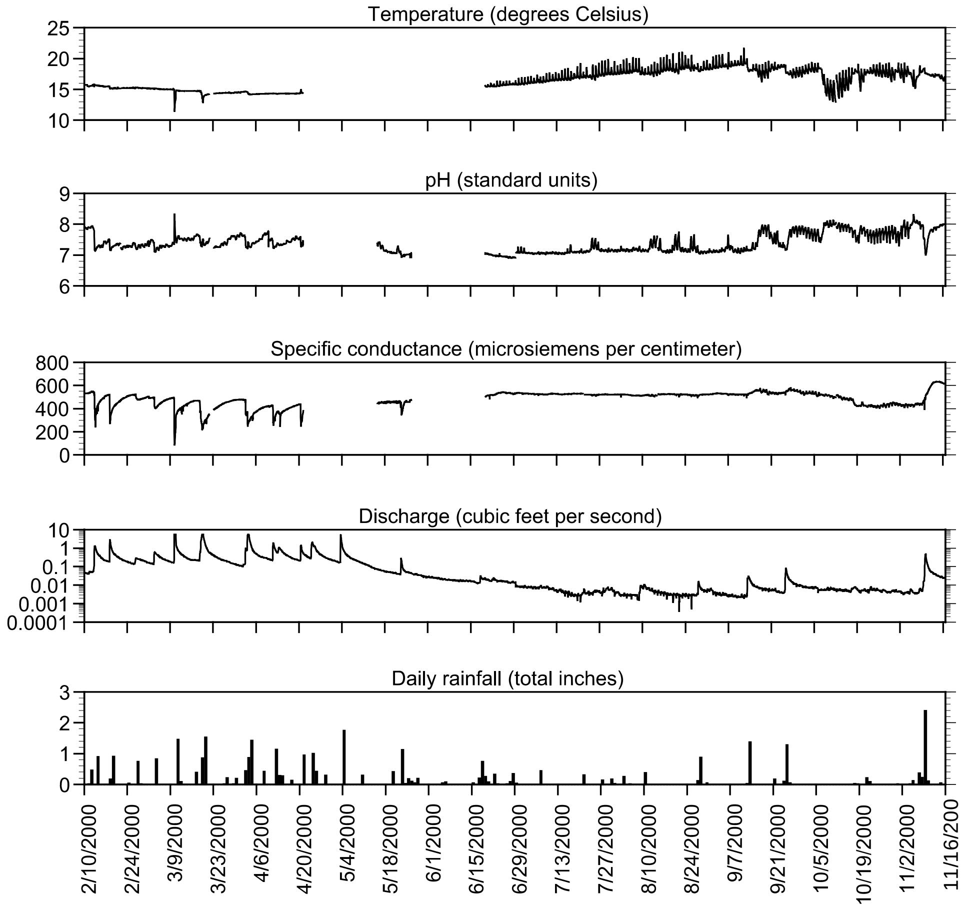


**Figure 4.** Water-quality, discharge, and rainfall data collected from Wilson Spring, Feb. 10 through Nov. 16, 2000 (discharge and rainfall data collected by Ogden Environmental and Energy Services). (Line gaps indicate missing data.)

time to interact with the limestone. During summer 2000, discharge declined and specific conductance remained stable at approximately 580 µS/cm. Daily cycles in values of pH and temperature in the summer may be the consequence of biological activity in the tub and solar radiation (fig. 4).

During the sampling period, six grab samples were collected from the tub that corresponded with samples collected by the automatic sampler. Chloroform concentrations detected in the samples collected by the automatic sampler were consistently within 15 percent of the concentrations detected in the grab samples.

On November 8, 2000, the first major storm occurred following a dry summer. The data from this storm elucidated some important characteristics about the hydrology of the spring. The storm produced 2.61 inches of rain beginning on November 8 and lasting until 1800 hours Greenwich Mean Time (GMT) on November 9 (fig. 5). At approximately 1500 hours GMT on November 8, discharge began to increase from a base-flow level of about 0.005 $ft^3/s$ to a peak of 0.5 $ft^3/s$ at 1350 hours GMT on November 9 (fig. 5). As discharge increased, specific conductance started to rise, reaching a maximum value of 630 µs/cm at 1500 hours GMT on November 13, approximately 4 days after the peak discharge occurred. This peak is 205 µS/cm higher than the value of 425 µS/cm observed on November 8, 2000 (fig. 4).

Dreiss (1989) concluded that fluctuations in dissolved solids and other water-quality properties, as well as contaminant concentrations that do not coincide with peak discharge, are common in mature karst systems. Temporary increases in specific conductance after the start of storm events are believed to be the result of more mineralized water that is driven from the diffuse-flow parts of the aquifer by increases in head in the conduits (Hess and White, 1988). During the period of increasing

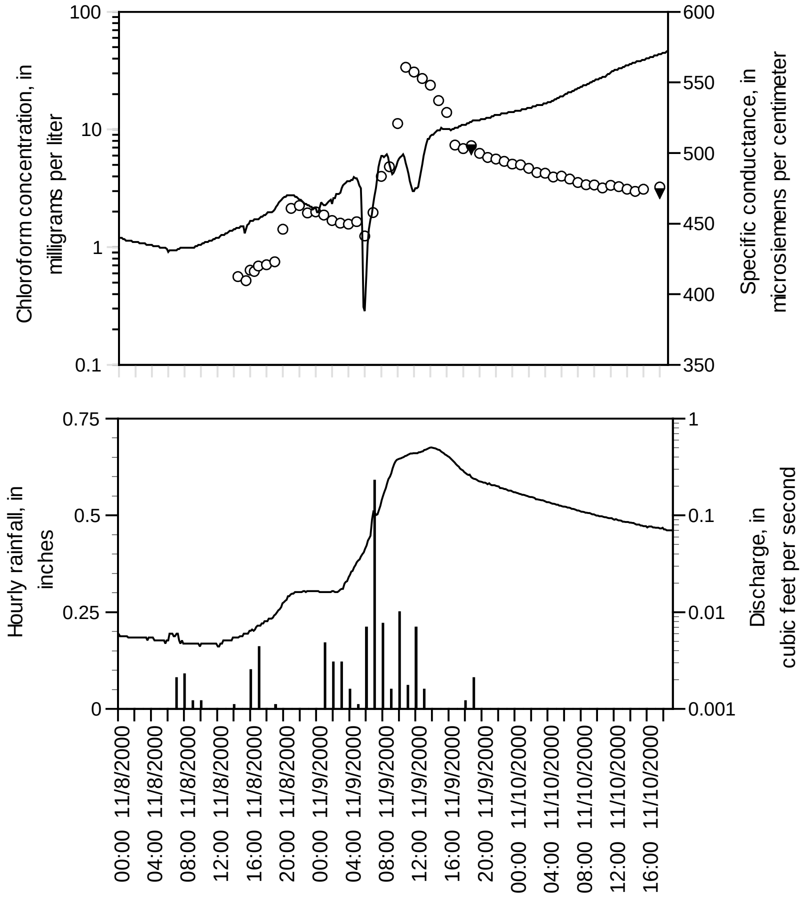


**Figure 5.** Chloroform, specific conductance, rainfall, and discharge data collected from Wilson Spring during a November 2000 storm (discharge and rainfall data collected by Ogden Environmental and Energy Services). Water samples were analyzed by the U.S. Geological Survey National Water Quality Laboratory (USGS NWQL) or using a portable gas chromatograph (GC).

discharge in Wilson Spring on November 9, 2000, an abrupt decline in specific conductance was observed and correlated with a decrease in chloroform concentrations (fig. 5). During a minor storm on September 12, 2000, a similar decrease in specific conductance occurred. During the latter part of this storm as specific conductance increased, VOC concentrations increased from 0.76 to 1.27 mg/L (fig. 6). VOC data were not collected during the first part of the storm. On September 21 and on August 14, 2000, chloroform concentrations of 1.38 and 0.86 mg/L, respectively, were detected in water samples from the spring. The decrease in specific conductance on September 12 and November 9, 2000, are both interpreted to be the results of direct recharge of water to the aquifer from a nearby sinkhole or sinkholes. Hess and White (1988), in studies of other karst springs, interpreted these types of events to indicate arrival of water from specific solution channels to the main spring.

During the November 2000 storm, a second distinct specific conductance decrease occurred, coinciding with peak discharge. During this part of the storm, chloroform concentrations increased to 33 mg/L (fig. 5), which represents a 60-fold increase over the concentration before the storm. Dye traces performed

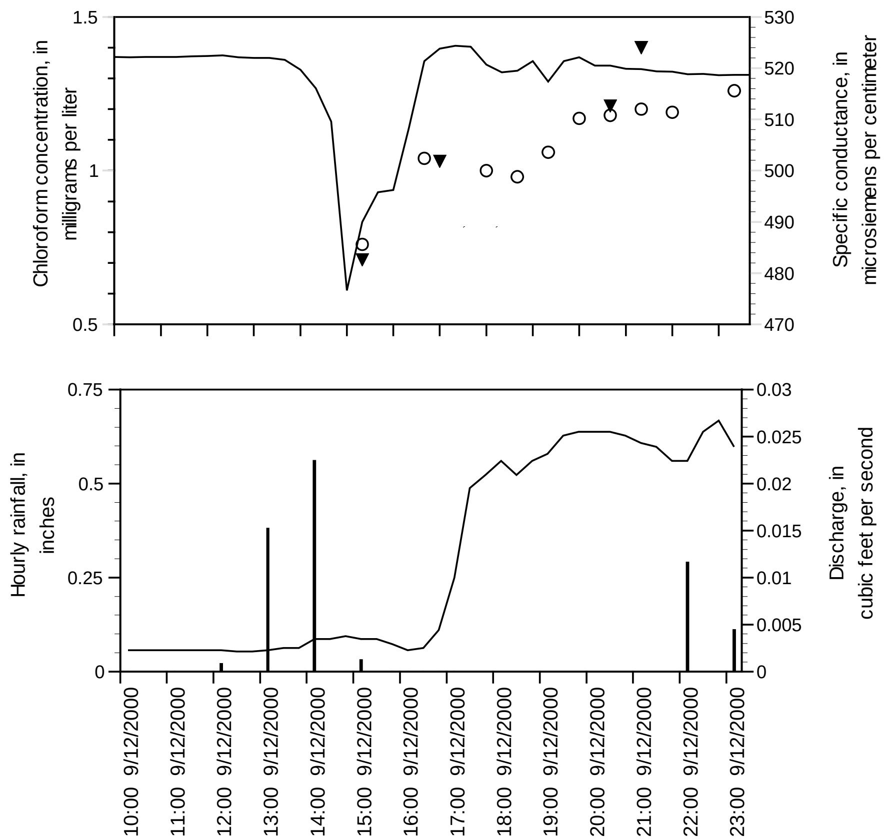


**Figure 6.** Chloroform, specific conductance, rainfall, and discharge data collected at Wilson Spring during a September 2000 storm (discharge and rainfall data collected by Ogden Environmental and Energy Services). Water samples were analyzed by the U.S. Geological Survey National Water Quality Laboratory (USGS NWQL) or using a portable gas chromatograph (GC).

by Crawford and Ulmer (1994) indicate the main source of water to Wilson Spring to be a cave stream that flows approximately 2,200 feet from near the contaminant source area to the spring. This cave stream appears to be transporting high concentrations of chloroform to the spring under certain hydrologic conditions. The maximum chloroform concentrations detected during the November 2000 storm, represents a sixfold increase over the maximum concentrations measured at any other time during the study period. Data collected during subsequent storms will determine if increased chloroform concentrations occur during storms throughout the year or only during specific hydrologic events, such as the first major storm after extended dry periods.

## CONCLUSIONS

Because of the low probability of successfully drilling wells that intersect karst conduits, springs are commonly used to monitor ground-water contamination in karst terrane. Literature is replete with studies documenting large water-quality variations in karst springs in response to storm events. Studies have addressed the need for specially designed sampling methodologies to detect peak contaminant concentrations in karst springs. Knowledge of the complete range in contaminant concentrations is necessary to calculate accurate contaminant loads in discharge from karst springs. In addition, pulses of high contaminant concentrations could exceed toxicity levels for aquatic organisms. Lack of awareness of the occurrence of pulses could result in undetected risk to the aquatic community in and downstream of these springs. Despite the previous work, sampling programs for monitoring karst springs rarely implement the special techniques needed to accurately characterize contaminant concentrations and loads. The wide range in concentrations present in the storm samples from Wilson Spring indicates the need for sampling strategies to be developed for karst springs on a site-specific basis. The water-quality signature of a given spring can be determined by continuously monitoring selected water-quality properties through one complete hydrologic cycle. This signature can be used to develop long-term sampling strategies that best characterize contaminant levels. In springs such as Big Spring, quarterly or semi-annual sampling may be adequate, but in flashy springs such as Wilson Spring, a more comprehensive program including the collection of high-frequency flow-controlled VOC samples is necessary.

The springs evaluated in this study contain VOC's. Only one other study was found in the literature that addressed sampling for these types of compounds in a karst system. Additional research is needed to evaluate the differences in contaminant signatures that may exist for the various classes of chemicals in karst springs. This study further emphasizes the need to re-evaluate sampling methodology in karst aquifer systems.

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# Use of Oligonucleotide Hybridization Probes and Polymerase Chain Reaction to Determine the Source of Fecal Contamination in Karst Terranes

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## ABSTRACT

Fecal contamination of surface and ground water remains a serious health problem in the U.S. and the world. This problem cannot be adequately addressed until the source of the contamination is known and remediated. At present, no standard monitoring technique exists that can identify the source of the bacteria. We have been working to modify and apply the molecular technique known as "oligonucleotide hybridization probes" to identify bacteria sources in the karst terrane of Middle Tennessee. The oligonucleotide hybridization probes can be used to target ribosomal RNA and DNA. Fecal bacteria unique to a host species can be identified using this molecular technique. Tests have been conducted using RNA hybridization probes that target universal sequences (all bacteria), *E. faecalis* (warm blooded animals), *Lachnospira multiparus* (ruminants), *Fibrobacter succinogenes* (ruminants), *Fibrobacter succinogenes* (ruminants), *Fibrobacter intestinales* (ruminants), *Bacteroides distasonis* (humans), *Bacteroides vulgates* (primarily human), *Bacteroides fragilis* (human) and *Salmonella* sp. (poultry & human pathogen). We were able to differentiate between these organisms in blind studies, but the concentration of bacteria had to be equal to or greater than 1,000 bacteria per liter before the cells were visually detected. The sensitivity was increased by using polymerase chain reaction (PCR) to amplify nucleic acid sequences on DNA (instead of RNA). PCR increased the sensitivity so that only 10 cells per liter were required for positive detection. Using PCR, we identified *Bacteroides fragilis,* a bacterium unique to humans, and *E. coli* in a water and biofilm sample collected from a karst spring-fed stream in Middle Tennessee. These preliminary results indicate this technique has potential to identify sources of fecal bacteria in hydrologically complicated karst terranes.

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# Geochemical and Microbial Evidence of Fuel Biodegradation in a Contaminated Karst Aquifer in Southern Kentucky, June 1999

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## Abstract

Complex hydrogeologic conditions coupled with poorly understood biodegradation processes in karst aquifers have led many to believe that the potential for natural attenuation of petroleum fuel hydrocarbons is limited. This research addressed the capacity for biodegradation processes in karst. Ground-water samples were collected for bacteria and geochemical analysis from several contaminated monitoring wells in an unconsolidated regolith and karst aquifer that had varying concentrations of dissolved fuel. Bacteria concentrations were greatest in ground-water samples containing the greatest fuel contamination. Additionally, bacteria isolated from fuel-contaminated ground-water samples readily grew in Petri dishes with dissolved gasoline fuel as the only source of food. The wells with screens intersecting less contaminated sections of the aquifer had greater dissolved oxygen concentrations (6.3 milligrams per liter) than those intersecting more contaminated sections (dissolved oxygen less than 0.1 milligrams per liter). Also, where the oxygen concentrations were diminished, geochemical evidence indicated that anaerobic processes were active. This evidence includes elevated levels of ammonia and ferrous iron in the fuel-contaminated ground-water samples. Based on these results, biodegradation of fuel constituents in the karst aquifer is indicated, and therefore, natural attenuation should not be disregarded because of preconceptions about low microbial activity in karst aquifers.

## INTRODUCTION

Approximately 40 percent of the United States east of the Mississippi River is underlain by various types of karst aquifers (Quinlan, 1989), and more than two-thirds of the State of Tennessee is underlain by carbonate rocks that are classified as karst (Wolfe and others, 1997). Potential sources of ground-water contamination are numerous in karst regions; however, the fate of ground-water contaminants such as petroleum hydrocarbons in karst aquifers is poorly understood because of the complex hydrology (Field, 1993) and the lack of biodegradation studies. Furthermore, ground-water models that simulate flow or predict the fate and transport of contaminants in unconsolidated, porous-media aquifers have limited application to karst aquifers. Most natural attenuation and bioremediation guidelines specify that these models are not applicable in fractured rock or karst aquifers (U.S. Environmental Protection Agency, 1997).

The U.S. Geological Survey (USGS) conducted a preliminary study to evaluate whether microbes in a karst aquifer have a capacity to biodegrade the fuel-related compounds benzene, toluene, ethyl-benzene, and xylenes (BTEX). This report presents field and laboratory data collected to determine the potential for biodegradation of jet-fuel contamination in a karst aquifer in Southern Kentucky. The study site was selected because of the known presence of BTEX in the ground water and the availability of hydrologic and historic water-quality data. The objective of the study was to characterize the microbial population and geochemical conditions present in ground water at the study site to determine if BTEX biodegradation in the karst aquifer was significant enough to warrant further investigation.

## METHODS AND MATERIALS

Ground-water samples were collected from seven monitoring wells located within a 5-hectare study area with a known jet-fuel spill, north of Clarksville, Tennessee, in southern Kentucky (fig.1). The wells were selected based on depths of screens and recent reports of fuel contaminant concentrations. The goal was to examine the geochemistry and bacteria present under a broad range of hydrogeologic (fig. 1) and contaminant conditions. The monitoring wells selected were screened in regolith, epikarst, and bedrock.

Water-quality samples were collected by the USGS during June 1999. Ground-water samples were collected from the wells using a submersible pump. Wells were micropurged as according to Puls and Paul (1995). The pump was lowered to the middle of the screen before starting. Water was pumped at a rate of approximately 1 liter per minute. An electronic tape measure was used to concurrently monitor the water level in the well casing. Water recharging the well was considered to be equal to the water being pumped when the water level became steady. At the point when water recharge equaled pump discharge and the

In Eve L. Kuniansky, editor, 2001, U.S. Geological Survey Karst Interest Group Proceedings, Water-Resources Investigations Report 01-4011, p. 151-156

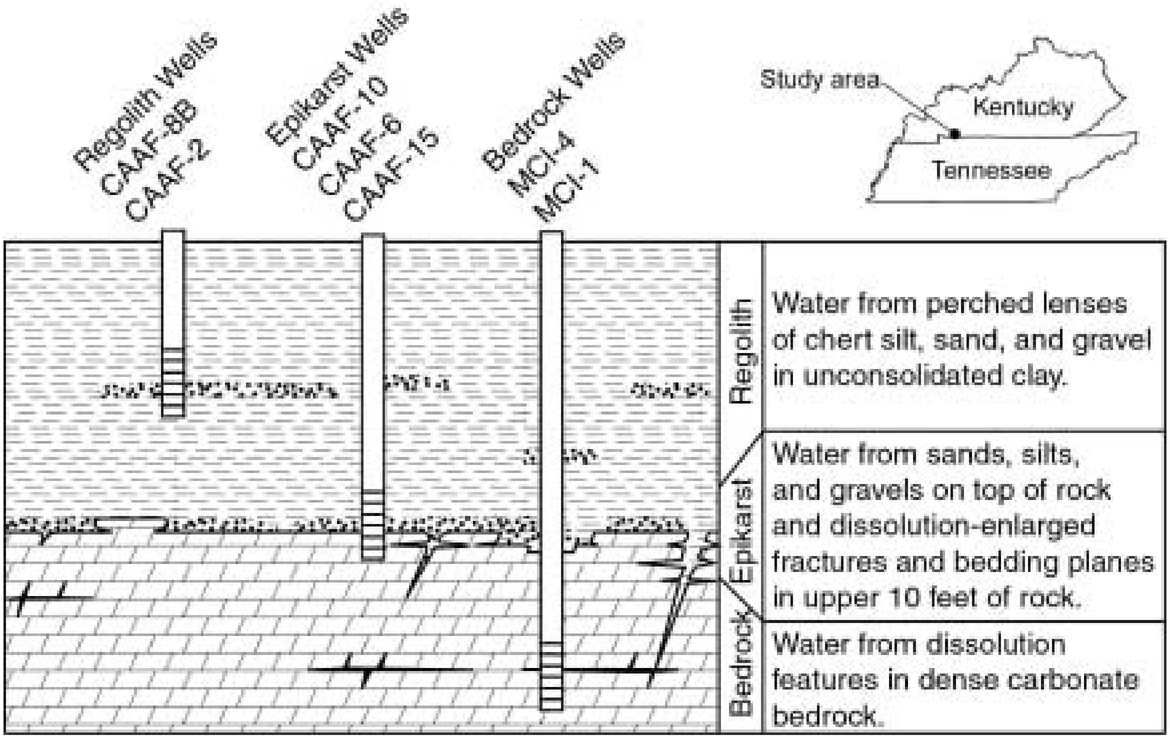


Figure 1. Generalized hydrogeologic setting of wells sampled for the biodegradation assessment of a karst aquifer in southern Kentucky.

specific conductance, turbidity, and fluorescence values in the pumped water no longer changed over a 1-minute period, water samples were collected. The first ground-water samples collected were for volatile organic compounds, followed by bacteria, and geochemical analyses respectively. Bacteria growth tests were begun within 6 hours of sample collection. Dissolved oxygen, alkalinity, carbon dioxide, pH, specific conductance, and temperature were measured immediately upon sample collection (Wood, 1981). The other ground-water samples were preserved according to Wood (1981) and the Hach Company (1992), and were analyzed within the holding period. The water sample for BTEX analysis was collected in three 40-milliliter (mL) vials and analyzed by gas chromatography at the USGS, National Water Quality Laboratory in Arvada, Colo. Analyses for nutrients and geochemical constituents were conducted using spectrophotometric methods described by Hach Company (1992), and, by Byl and Williams (2000).

Bacteria were enumerated and identified in ground-water samples from the monitoring wells to supplement the interpretation of the geochemical information. Bacteria were enumerated from four of the seven monitoring wells using two methods (data from the remaining three wells were not used because the samples were not processed within 24 hours of collection). The facultative and aerobic heterotrophic bacteria were enumerated using dilution buffer and tryptic soy agar plate counts as described by Eaton and others (1995). Aerobic and facultative bacteria capable of using gasoline as their sole food source also were enumerated using sterile media pads soaked with dilution buffer containing dissolved gasoline. Sterile filters with 0.45-micrometer pore sizes were used to inoculate the growth media. A 0.01- or 1-mL aliquot of sample water was transferred to 20 mL of sterile dilution buffer and drawn onto the filter as described in the membrane-filtration method (Britton and Greeson, 1989). The filters were placed on the growth media using sterile forceps, and the plates were placed in an incubator at 35 °C for 5 days. Bacteria colony-forming units were counted after 24, 48, and 120 hours. Results are reported as colony-forming units per milliliter of sample water using data from the 48-hour count (table 1).

Bacteria types present in the samples were identified by the use of the ribonucleic (RNA)-oligonucleotide hybridization method (Amann and others, 1995; Byl and others, 1997). The RNA-oligonucleotide hybridization method is a technique that exploits unique nucleotide sequences in the ribosomal RNA (rRNA) to identify bacteria (Amann and others, 1995). The method can be used to identify groups of bacteria such as sulfate-reducing bacteria or specific genera such as *Nitrosomonas* sp. The level of identification used in this study was general bacteria groups such as iron oxidizers, methanotrophs, ammonia oxidizers, and sulfate reducers (Byl and others, 1997; Farmer and others, 1998).

## RESULTS AND DISCUSSION

Results of the analyses demonstrated geochemical conditions that are indicative of fuel biodegradation in contaminated wells. A strong inverse pattern was found between dissolved oxygen (DO) levels and BTEX concentrations in the water samples, implying a biological oxygen demand resulting from BTEX consumption (tables 1 and 2). In addition, a great quantity of bacteria was present in each of the samples, with the largest populations associated with the most contaminated wells. Subsurface bacteria known to influence BTEX degradation processes were identified using RNA-hybridization probes (Farmer and others, 1998, Chapelle, 1993).

Table 1. Summary of benzene, toluene, ethylbenzene, and xylenes (BTEX) and bacteria analysis for wells in study area, June 9-11, 1999. Descending down the chart, wells are arranged in relative order of shallow to deepest and least to most fuel contaminated. The blank was collected from equipment (pump) rinse water between sampling at wells CAAF-2 and CAAF-6. Bacteria were enumerated using media pads soaked with dilution buffer containing gasoline as a food source. No bacteria grew in the control treatments when the gasoline was absent. Bacteria samples also were tested with different RNA-hybridization probes to identify bacteria. Bacteria types identified were: *Pseudomonas aureginosa* (known to efficiently biodegrade BTEX in soils), *Pseudomonas* species (a broader, more encompassing group of Pseudomonas bacteria), ammonia-oxidizing bacteria, and iron-oxidizing bacteria. The intensity of the bacteria identified by RNA-hybridization is indicated by "+" symbols. Three +++ symbols means greater than 100 bacteria present per milliliter, two ++ indicate a moderate number of bacteria (10 - 100 bacteria / mL), one + indicates there were only a few bacteria present (1 - 10 bacteria/mL).

[Bnz, benzene; Tol, toluene; Eth, ethyl-benzene; Xyl, xylenes; CFU, colony-forming units per 1 milliliter; µg/L, microgram per liter; mL, milliliter; <, less than; RNA, ribonucleic acid; BD = below detection; ND = No data; Pseudo., *Pseudomonas*; sp., species; aure, *aureginosa*; Ammonia ox., ammonia oxidizers]

| Well and date samples were collected | Hydro-geologic setting | Bnz µg/L | Tol µg/L | Eth µg/L | Xyl µg/L | CFU per mL | RNA hybridization results |
|---|---|---|---|---|---|---|---|
| CAAF-8B 6/9/99 | Regolith | <0.5 | 1.00 | <0.5 | 1.6 | 500 | Pseudo. aure.+, Pseudo. sp. + |
| CAAF-10 6/10/99 | Epikarst | BD | 0.63 | <0.5 | 0.6 | 325 | Pseudo. aure.+, Pseudo. sp. + |
| MCI-4 6/9/99 | Bedrock | <0.5 | 0.75 | <0.5 | 0.6 | 93 | Pseudo. sp. ++ |
| CAAF-2 6/11/99 | Regolith | 2.4 | BD | 11.0 | 5.5 | ND | Pseudo. sp. ++, Ammonia ox. + |
| CAAF-6 6/11/99 | Epikarst | 16.0 | 3.50 | 8.8 | 25.0 | ND | Pseudo. aure.+, Pseudo. sp. ++, Ammonia ox. ++ |
| CAAF-15 6/11/99 | Epikarst | 220. | BD | 3.1 | 360.0 | ND | Pseudo. aure.+++, Pseudo. sp. +++, Iron oxidizer +++, Ammonia ox. ++ |
| MCI-1 6/10/99 | Bedrock | 75.0 | 14.0 | 26.0 | 38.0 | 2550 | Pseudo. aure.+++, Pseudo. sp. +++, Ammonia ox. + |
| BLANK | | 0.71 | BD | 4.3 | 2.9 | ND | (None) |

Table 2. Summary of selected geochemical data for wells in study area, June 9-11, 1999.

[µS/cm, microsiemen per centimeter; mg/L, milligram per liter; $CaCO_3$, calcium carbonate; DO, dissolved oxygen; $NO_3$, nitrate; $NH_3$, ammonia; Aq $Fe^{2+}$, aqueous ferrous iron; <, less than)

| Well and date samples were collected | Hydro-geologic setting | pH | Specific conduct-ance [µS/cm] | Alka-linity [mg/L $CaCO_3$] | DO [mg/L] | $NO_3$ [mg/L] | $NH_3$ [mg/L] | Aq $Fe^{2+}$ [mg/L] |
|---|---|---|---|---|---|---|---|---|
| CAAF-8B 6/9/99 | Regolith | 5.0 | 31 | 9 | 6.3 | 3.5 | <0.01 | 0.01 |
| CAAF-10 6/10/99 | Epikarst | 7.3 | 266 | 143 | 5.5 | 2.6 | <0.01 | 0.06 |
| MCI-4 6/9/99 | Bedrock | 6.9 | 373 | 145 | 3.6 | 14.5 | <0.01 | 0.01 |
| CAAF-2 6/11/99 | Regolith | 6.0 | 548 | 305 | <0.1 | 4.4 | 2.59 | 16.20 |
| CAAF-6 6/11/99 | Epikarst | 7.4 | 233 | 121 | 0.3 | 0.9 | <0.01 | 0.06 |
| CAAF-15 6/11/99 | Epikarst | 7.3 | 257 | 120 | 0.5 | 1.3 | 0.19 | 0.14 |
| MCI-1 6/10/99 | Bedrock | 6.3 | 549 | 293 | <0.1 | 4.4 | 0.80 | 5.80 |

The bacteria identified were *Pseudomonas* sp., *Pseudomonas aureginosa*, and two groups of bacteria known as ammonia-oxidizing and iron-oxidizing bacteria. *Pseudomonas* bacteria, one of the most common soil bacteria, contain a variety of enzymes that degrade BTEX, such as toluene di-oxygenase and catechol di-oxygenase (Chapelle, 1993). *Pseudomonas aureginosa* is a specific species in the *Pseudomonas* family that has been shown to be important in BTEX biodegradation (Houghton and Shanley, 1994). The number of *Pseudomonas aureginosa* generally increased with increasing BTEX concentrations in the samples, indicating they were flourishing in the contaminated waters.

Ammonia-oxidizing bacteria also were identified in samples from four of the wells. These bacteria types have an enzyme called ammonia mono-oxygenase that can degrade BTEX in the presence of DO (Bedard and Knowles, 1989). Bacteria associated with iron geochemistry also have been shown to degrade BTEX (Chapelle, 1993). Only one highly contaminated well (CAAF-15) tested positive for iron-oxidizing bacteria. Elevated concentrations of dissolved iron in wells CAAF-2 and MCI-1 indicated that iron-reducing bacteria were active in other parts of the aquifer. In general, a consortium of aerobic and anaerobic bacteria appeared to be present in water samples from the contaminated wells implying that environmental conditions change in the aquifer or that heterogeneous niches exist within the karst aquifer with regard to redox conditions.

Perhaps the most direct evidence that bacteria in the karst aquifer at this site are capable of degrading fuel came when bacteria were grown directly on gasoline as the sole food source. Filters containing indigenous bacteria from the aquifers were placed on sterile media pads soaked with dilution buffer (Eaton and others, 1995) with gasoline dissolved in it. The control tests contained sterile media pads soaked with dilution buffer only, and no gasoline. Results of these tests are shown by the number of colony-forming units (CFU) per milliliter that grew on the pads (table 1). Bacteria in water samples from well MCI-1, a karst bedrock aquifer well, grew rapidly on the gasoline-amended media, while nothing grew on the filters placed on the control pads containing dilution buffer only, indicating the bacteria used the gasoline as a food source to grow and replicate. Bacteria from three other well samples, CAAF-8B, CAAF-10, and MCI-4, also grew on gasoline-amended media, but not on the control pads soaked with only dilution buffer. Samples from wells CAAF-2, CAAF-6, and CAAF-15 were not tested for growth on gasoline because of time constraints.

Results from these tests indicate several preliminary lines of evidence exist that bacteria are actively degrading fuels in the karst aquifer at the site. The geochemical evidence includes such indicators as oxygen consumption, ammonia production, iron dissolution, and sulfur reduction as bacteria use these constituents in their metabolic processes (table 2). The bacteria evidence includes identification of bacteria known to consume BTEX, high bacteria counts, and growth of indigenous bacteria using fuel as their only food source. These data demonstrate that bacteria thrive in water from both the deep and shallow wells at the site. Based on these results, biological degradation of BTEX is believed to occur in both unconsolidated regolith and bedrock parts of the fuel-contaminated aquifer.

## *Implications of the findings*

The concentrations of bacteria present in the bedrock aquifer samples were high enough to be comparable with concentrations reported for contaminated sand aquifers (Ghiorse and Wilson, 1988). Thus, some of the fuel biodegradation utilization capacity derived for sand aquifers may apply to ground water in regolith, epikarst, and karst aquifers. A stoichiometric balance of terminal electron acceptors utilized in the consumption of BTEX by bacteria is used to derive a utilization capacity (Newell and others, 1996). Comparing geochemical constituents in relatively clean wells to concentrations in more contaminated wells derives the estimated concentration of terminal electron acceptors in an aquifer. Using the methods and stoichiometric equations found in the U.S. Environmental Protection Agency model BIOSCREEN: Natural Attenuation Decision Support System (Newell and others, 1996), the biodegradation capacity would be nearly 4 and 3 milligrams BTEX per liter of water based on the concentrations of DO, nitrate ($NO_3$), ferrous iron ($Fe^{2+}$) and sulfate ($SO_4^{2-}$) found in samples from the regolith and bedrock aquifers, respectively (table 3). These results imply that BTEX dissolved in the regolith and karst aquifers are likely to be consumed by bacteria given sufficient time (the highest BTEX concentrations identified in the karst aquifer have been less than 1 mg/L). Note that the biodegradation capacity does not provide a rate of biodegradation, so factors such as ground-water flow velocity and contaminant retention time in the aquifer are important unknown variables. A study that integrates a more detailed assessment of the hydrogeology, along with further geochemistry and biology would be required to determine if BTEX biodegradation occurs at a rate sufficient to protect target receptors.

Table 3. Concentration of terminal electron acceptors used to determine BTEX biodegradation capacity as described in Newell and others (1996)

[BTEX, benzene, toluene, ethyle-benzene, xylenes; mg, milligram; mg/L, milligrams per liter]

| Terminal electron acceptor molecule (TEA) or byproduct | BTEX Utilization Factor (mg TEA consumed / mg of BTEX degraded) | Estimated concentra-tions of TEA in regolith water | Biodegradation capacity in regolith (mg of BTEX degraded per liter of water) | Estimated concentra-tions of TEA in bedrock water | Biodegradation capacity in bedrock aquifer (mg of BTEX degraded per liter of water) |
|---|---|---|---|---|---|
| Dissolved oxygen (DO) | 3.14 mg DO / mg BTEX | 6.3 mg / L | 2.00 | 3.6 mg /L | 1.15 |
| Nitrate ($NO_3$) | 4.9 mg $NO_3$ / mg BTEX | 3.5 mg / L | 0.61 | 5.0 mg/L | 1.00 |
| Ferrous ($Fe^{2+}$) iron from ferric ($Fe^{3+}$) | 21.8 mg $Fe^{3+}$ transformed / mg BTEX | 0.05 mg / L | 0.73 | 16 mg/L | 0.27 |
| Sulfate ($SO_4^{2-}$) | 4.7 mg $SO_4^{2-}$ / mg BTEX | 4.0 mg / L | 0.43 | 2.0 mg/L | 0.43 |
| **Estimated total biodegradation capacity** | | | **= 3.77 mg** | | **2.85 mg** |

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# Preliminary Conceptual Models Of Chlorinated-Solvent Accumulation in Karst Aquifers

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## Abstract

Conceptual models are needed to assist regulators and site managers in characterizing chlorinated-solvent contamination in karst settings and evaluating clean-up alternatives. Five preliminary conceptual models were developed, emphasizing accumulation sites for chlorinated DNAPL in karst aquifers. The models were developed for the karst regions of Tennessee, but are intended to be transferable to similar karst settings elsewhere.The five models of DNAPL accumulation in karst settings are: (1) trapping in regolith; (2) pooling at the top of bedrock; (3) pooling in karst conduits; (4) pooling in bedrock diffuse-flow zones; and (5) pooling in isolation from active ground-water flow.

More than one conceptual model of DNAPL accumulation may be applicable to a given site depending on details of the contaminant release and geologic setting. Trapping in regolith is most likely where the regolith is thick and relatively impermeable with few large cracks and fissures. Accumulation at the top of rock is favored by flat-lying strata with few fractures or karst features near the bedrock surface. Fractures or karst features near the bedrock surface encourage migration of chlorinated DNAPL into karst conduits or diffuse-flow zones in bedrock. DNAPL can migrate through one type of bedrock aquifer into an underlying aquifer of a different type or into openings that are isolated from significant ground-water flow.

## INTRODUCTION

Chlorinated solvents are widely used in many industrial operations. High density, low viscosity, and low interfacial tension relative to water make chlorinated solvents mobile contaminants that are difficult to find or remove when released into the ground-water system. Because karst conduits are commonly too large to develop significant capillary pressures, chlorinated solvents can migrate to considerable depth in karst aquifers as dense non-aqueous-phase liquids (DNAPL's). Within the context of this report, the term DNAPL is used to describe the immiscible or non-aqueous phase of chlorinated solvents and applies only to liquids with high density and low viscosity.

Chlorinated solvents generally enter the subsurface environment as DNAPL and migrate downward and laterally until local conditions favor their accumulation (Schwille, 1988; Cohen and Mercer, 1993; Pankow and Cherry, 1996). Major controls of the movement and ultimate fate of chlorinated DNAPL in the subsurface are: (a) the physical and chemical properties and mass of specific contaminants, (b) the areal extent and rate of contaminant release, and (c) the nature of the hydrogeologic environment into which the contaminant migrates (Mercer and Cohen, 1993; Pankow and Cherry, 1996). For a chlorinated solvent release in karst, specific factors that control the residence time of contaminant accumulations and the concentration and movement of related dissolved-phase contamination include:
(1) the mass of the bulk contaminant source and its location relative to the water table;
(2) the sorption properties of the material through which the contaminant migrates and in which it accumulates;
(3) the surface-area to volume ratio of DNAPL ganglia, blobs, pools, and residual accumulations;
(4) the local ground-water flow regime in areas of chlorinated-solvent accumulation;
(5) the degree of hydraulic connection between areas of chlorinated-solvent accumulation and karst conduit systems; and
(6) the overall ground-water flow regime of the area surrounding the site and the location of the contaminant source relative to recharge and discharge boundaries.

A necessary first step toward improving site characterization for chlorinated-solvent spills in karst settings is the development of conceptual models of where the contaminants are likely to accumulate and where and how they may be moving. A good conceptual model provides numerous working hypotheses that can be evaluated and refined as new information becomes available. The absence of an adequate conceptual model commonly leads to wasted effort and expense as data are collected which do little to illuminate the problem at hand. Standard techniques of site characterization developed for aqueous-phase

In Eve L. Kuniansky, editor, 2001, U.S. Geological Survey Karst Interest Group Proceedings, Water-Resources Investigations Report 01-4011, p. 157-162

contaminants or for porous granular media may provide irrelevant or erroneous results at DNAPL sites in karst settings (Quinlan and Ray, 1991; Cohen and Mercer, 1993; Barner and Uhlman, 1995).

This report presents five conceptual models of DNAPL accumulation in karst settings, developed by the U.S. Geological Survey in cooperation with the Tennessee Department of Environment and Conservation, Division of Superfund. The models emphasize DNAPL accumulation in different compartments of the subsurface environment. The models were developed for the karst environments of Tennessee, but the concepts presented in this section are intended to be transferable to similar karst settings in adjacent states and applicable in other areas.

The five conceptual models of DNAPL accumulation in karst settings are:
1) DNAPL trapping in regolith;
2) DNAPL pooling at the top of bedrock;
3) DNAPL pooling in bedrock diffuse-flow zones;
4) DNAPL pooling in karst conduits; and
5) DNAPL pooling in isolation from active ground-water flow (fig. 1).

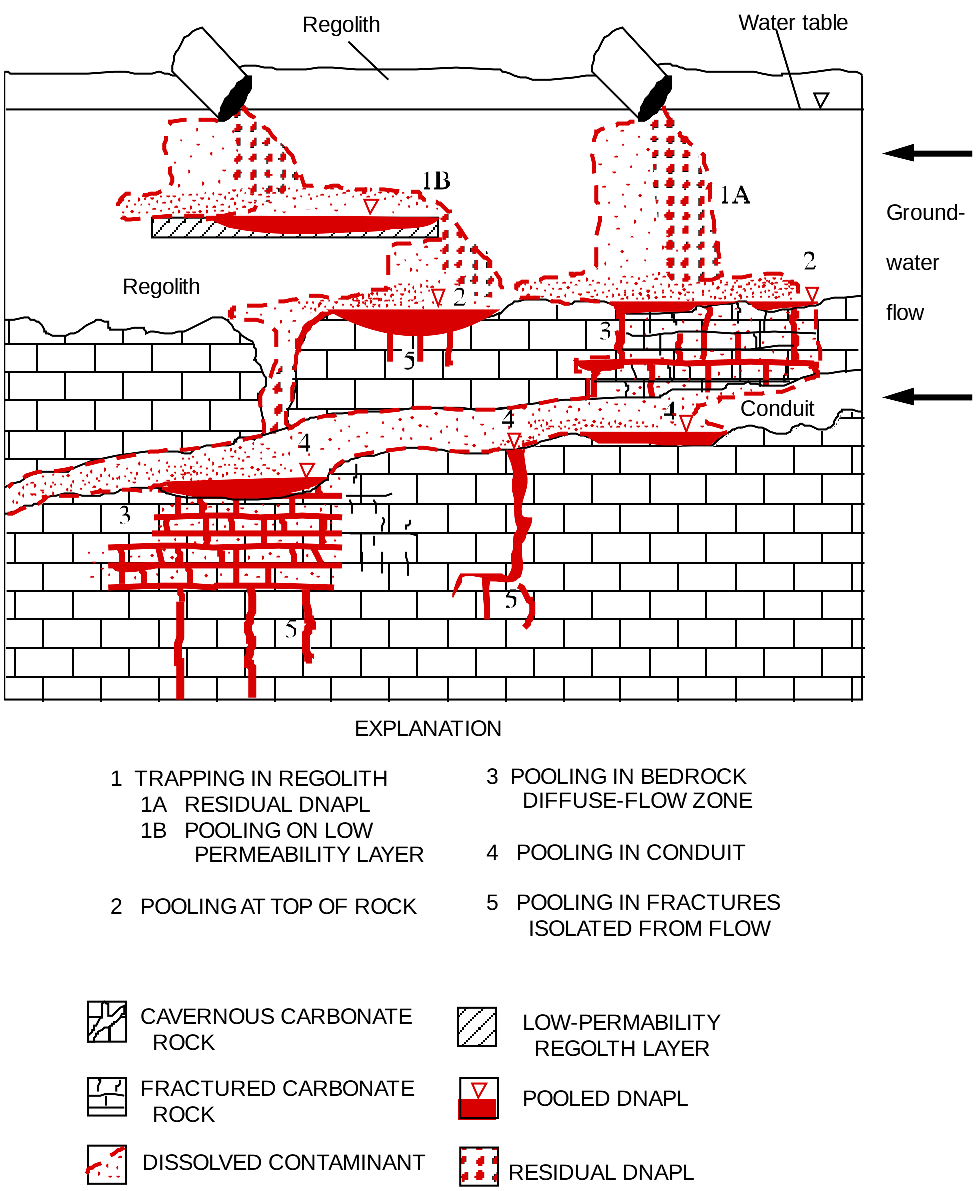


**Figure 1.** Distribution of potential DNAPL-accumulation sites in a hypothetical karst setting. (Modified from Wolfe and others, 1997.)

In Eve L. Kuniansky, editor, 2001, U.S. Geological Survey Karst Interest Group Proceedings, Water-Resources Investigations Report 01-4011, p. 157-162

The conceptual models presented in this report are preliminary in nature. They are intended to be starting points for analysis of chlorinated-solvent contamination in karst settings and do not reduce the critical importance of careful characterization of the environmental settings and contaminant distributions at specific sites. These preliminary conceptual models are scale neutral. There is no minimum amount of DNAPL that could be stored in any of these environmental compartments, and the maximum amounts are a function of the size and nature of the release and the hydrogeologic character of specific sites. The models are mutually compatible in that more than one model may be applicable to a given site.

## TRAPPING IN THE REGOLITH

In the regolith, DNAPL will migrate through macropores, fractures, and intergranular pores. Flow through macropores and fractures is important in fine-grained layers where DNAPL may not be able to enter the intergranular pores (Helton, 1987; Cherry, 1989). Several mechanisms work to trap DNAPL in the regolith. DNAPL will be retained by capillary forces as residual DNAPL under both unsaturated and saturated conditions. DNAPL can pool on top of layers that are lower in permeability relative to the over-lying layer and which provide a capillary barrier to further downward movement of DNAPL. Low permeability layers can be perennially saturated, perennially drained, or alternately saturated and drained.

Low permeability layers must be free of erosional or depositional gaps and free of fractures for pooling of DNAPL to occur. In many cases, the low permeability layers will deflect the downward movement of DNAPL, but will not be of sufficient lateral extent and composition to serve as a significant barrier to downward migration of DNAPL. Where these layers are discontinuous laterally, they can cause horizontal spreading of DNAPL then continued downward movement once the edge of the restrictive layer is reached, resulting in a complex DNPAL distribution. Typically, many small pools in discontinuous lenses and fractures would be expected to form. Contrasts in permeability can be important because even small variations play a major role in determining the DNAPL distribution in the regolith.

Capillary forces in any porous medium that DNAPL can enter will trap a certain amount as residual DNAPL. In the vadose zone, under dry conditions, DNAPL will be the wetting fluid and will be retained as films and wetting rings coating the media. More commonly, the vadose zone will be partly saturated with water and DNAPL will be retained as non-wetting ganglia in the pore throats and bodies of the media. In the saturated zone, DNAPL will be retained as isolated ganglia in the large pore body spaces. Residual saturation will be less in the unsaturated zone because DNAPL drains more easily in the presence of air than in a water-saturated system.

Macropores and fractures may be important pathways for DNAPL movement in clay-rich regolith. As with porous media, residual DNAPL will be held in these fractures by capillary forces as disconnected blobs and ganglia. Values of residual saturation in fractured clays will be less than in porous media (Pankow and Cherry, 1996) and will increase with decreasing fracture aperture (Schwille, 1988). DNAPL pools will accumulate in fractures that pinch out sufficiently with depth to provide a capillary barrier.

## POOLING NEAR TOP OF ROCK

DNAPL contamination can migrate down through the regolith to the top of the underlying bedrock. In this report, "top of rock" refers to the surface between the regolith and bedrock. In karst settings, this surface is commonly irregular, highly weathered, and variable in depth. The top of rock is part of a transitional zone (the subcutaneous zone or epikarst) that includes weathered rock fragments in the regolith and dissolution openings within the upper part of the bedrock. The transitional epikarst zone is commonly 3-10 m thick and extends above and below the top of rock (Quinlan, 1989).

For DNAPL to accumulate at the top of rock, the DNAPL must pass through the regolith and encounter a low-permeability pooling site at the bedrock surface. Even small volumes of DNAPL will in general have the potential to migrate down to the top of rock (Mercer and Cohen, 1993; Pankow and Cherry, 1996). At top of rock, DNAPL will accumulate in pools where differential weathering or structure has created irregularities in the bedrock surface.

For a given DNAPL release, the relative importance of pooling at top of rock will be influenced by the thickness and physical properties of the regolith, the bedrock lithology, and the geologic structure. Thin, permeable regolith will allow DNAPL to reach top of rock more easily than thick regolith with high residual saturation or numerous impermeable layers. Rocks with low secondary porosity or in which dissolution openings in epikarst pinch out with depth (Williams, 1983, 1985) will trap DNAPL more effectively than rocks with efficient hydraulic connections between their surface and underlying bedrock aquifers. Flat-lying or gently dipping rocks are more likely to trap DNAPL than steeply dipping rocks, especially in cases where secondary porosity develops preferentially along bedding planes.

## POOLING IN BEDROCK DIFFUSE-FLOW ZONE

A diffuse-flow zone in bedrock occurs where many small fractures are present, but dissolution is minor or where dissolution-enlarged fractures are filled with granular material. A typical environment for this situation is fractured shales and carbonates or conduits filled with sediment washed in from land surface. Ground water moves through a diffuse network of small fractures or intergranular pores rather than through discrete conduits. Flow through the diffuse network may converge on larger conduits with more active flow. In this situation, DNAPL present in a diffuse-flow zone could provide a source of aqueous-phase contamination to the more active flow in the conduit.

DNAPL will migrate down through the network of diffuse fractures until the capillary resistance becomes too high for continued downward movement. DNAPL will then pool in the fractures. Within a network of small fractures, large vertical accumulations of DNAPL are possible. The residence time of DNAPL pooled in zones of diffuse flow will be determined by dissolution and matrix diffusion. Dissolution into the actively flowing ground-water will be slow because of the small surface area available for DNAPL pooled in fractures. The rate of ground-water movement also will limit the amount of DNAPL that will be depleted by dissolution. In cases where DNAPL is pooled in a network of fractures, a much larger surface area is available for matrix diffusion. Matrix diffusion can be an important process if the rock matrix has significant primary porosity. In the dense, Paleozoic carbonates of Tennessee and similar settings, significant primary porosity is likely only in a zone surrounding fractures and bedding planes where a significant width of dissolution has left a broad band of insoluble residue within an impure soluble rock. DNAPL depletion by matrix diffusion from all but the smallest (< 0.1 mm) aperture fractures may take years or longer (Pankow and Cherry, 1996). After DNAPL has been depleted from fractures, solvent dissolved in the matrix pore water will diffuse back into the fracture, serving as a persistent source of dissolved-phase contamination.

## POOLING IN KARST CONDUITS

The characteristic size range of karst conduits, typically on the order of millimeters to tens of meters (Ford and Williams, 1989, ch. 7), is too large for capillary forces to significantly restrict DNAPL movement (Wolfe and others, 1997). Thus, DNAPL will freely flow into most open conduits it encounters. Once it enters a conduit, DNAPL will flow along the conduit floor, collecting in cracks, pits, or other depressions. The movement, transformation, and persistence of a given DNAPL mass in a karst conduit depends on such case-specific factors as the size and shape of the conduit, the topography of the conduit floor, the degree of residual or sedimentary fill, and the position of the conduit relative to water-table fluctuations. All of these factors exhibit enormous variation in karst terranes (White, 1988; Ford and Williams, 1989).

Karst conduits develop along preferential pathways between areas of ground-water recharge and discharge, and are enlarged by dissolution (White, 1988). Karst landforms, such as sinkholes, that concentrate recharge may be closely integrated with conduits and thus provide direct routes for contaminants to conduit flow networks (Quinlan and others, 1992; Field, 1993, Crawford and Ulmer, 1994).

Once DNAPL enters a conduit, any irregularity or obstruction in the conduit floor or inflection in conduit orientation will provide a place for the DNAPL to pool. Studies of springs show that karst conduit systems retain significant volumes of easily displaced vadose storage (water) in pools (Joseph Meiman, U.S. National Park Service, oral commun., 1997). The same pools will also hold DNAPL and may have enough volume to contain large DNAPL spills. Low spots along the floor of a conduit where DNAPL can pool (sumps) may be without capillary cracks, leaving all the pooled DNAPL exposed to the overlying flow. Such conduits can develop along the tops of aquicludes, or through massive limestone. Other conduits have sumps coinciding with cracks in the conduit floor through which DNAPL could migrate downward and out of the conduit-flow system.

DNAPL pools in karst conduits can be perennially submerged in water or periodically exposed to air. DNAPL pools exposed to air will be depleted through volatilization. The fate of the resulting vapor will depend on the air-flow characteristics of the conduit. In many cases, vapor-phase chlorinated solvents may be as persistent in karst conduits as in other parts of the ground-water system. On the other hand, cave systems with high air flows (Bruce Zerr, Oak Ridge National Laboratory, oral commun., 1996) may efficiently route chlorinated-solvent vapors to the surface.

Flow velocities in conduits are high relative to ground-water flow rates in other settings (White, 1988; Quinlan and others, 1992). Recurrent inputs of fresh water tend to flush aqueous-phase contaminants and maintain a high concentration gradient close to the DNAPL pool. Dissolution is more likely governed by the maximum rate of dissolution into pure water than by the replacement of saturated solution in contact with the DNAPL. Frequent flushing of the DNAPL/water boundary layer would encourage relatively rapid dissolution and a short residence time (from weeks to years). However, Field (1993) notes that karst conduits can rapidly deliver significant quantities of contaminant to a discharge point yet still retain enough contaminant in storage to result in long-term ground-water contamination.

## POOLING IN FRACTURES ISOLATED FROM MAJOR ZONES OF GROUND-WATER FLOW

The high specific gravities and low viscosities of chlorinated solvents cause these compounds to migrate downwards until they encounter openings too small to enter. Under certain conditions, this downward migration can take DNAPL to fractures that are relatively isolated from major ground-water flow zones. In contrast to the previous cases discussed in this section, pools of DNAPL in isolated fractures have minimal interaction with flowing water. Reduced exposure to flowing water has major implications for DNAPL residence time, mitigation and delivery to drinking-water supplies.

Every karst aquifer has a lower boundary below which flow is greatly reduced. In general, smooth, abrupt lower boundaries are probably much less common than rough, gradational ones. Karst develops through the interaction of atmospheric water with soluble rock (White, 1988). In many cases, dissolution and the resulting secondary porosity are concentrated in the upper parts of a carbonate rock unit and decrease with depth (Ford and William, 1989, p. 158-162). The base of karstification is typically a zone in which the karst-conduit system propagates downward through the progressive enlargement and integration of discrete voids which initially have only poor interconnection. This zone may be at considerable depth below the zone of major flow within the aquifer (Ford and Williams, 1989, p. 177-178). The network of conduits and fractures above the base of karstification provides potential flowpaths for DNAPL through the major flow zone to the smaller, more isolated voids below.

A DNAPL pool isolated from ground-water flow will have a long residence time (on the order of decades or longer). The major mechanism for removal will be diffusion into adjacent fractures and primary pores. The low rate of local flow will limit flushing of aqueous phase, resulting in a relatively low concentration gradient near the DNAPL mass and a correspondingly low rate of diffusion and dissolution. Migration of aqueous phase to zones of higher ground-water flow will occur through diffusion. Depending on the flow system, the rate of diffusion may be small or large relative to the flow, resulting in greater or lesser attenuation by dilution. Whatever the attenuation achieved by dilution, aqueous-phase contamination of ground water from DNAPL pools isolated from ground-water flow is likely to persist for many decades.

## CONCLUSION

Two of the most problematic topics in contaminant hydrogeology are chlorinated solvents and karst. Chlorinated solvents have physical and chemical properties that make this class of compounds particularly likely to cause groundwater contamination. The high densities and low viscosities of chlorinated solvents allow them to move readily downward as a DNAPL through the subsurface due to gravity. The same properties that make chlorinated solvents potent ground-water contaminants make them difficult to locate or remove once they enter the ground-water system. Nowhere is this more true than in karst settings.

The extensiveness of karst aquifers and their distinctive hydraulic properties makes these aquifers vulnerable to contamination by chlorinated solvents. DNAPL accumulation areas within an aquifer are important because they are source zones for dissolved-phase contamination. The conceptual models developed emphasize DNAPL accumulation in five compartments of the subsurface environment: (1) the regolith, (2) the top of bedrock, (3) karst conduits, (4) bedrock diffuse-flow zones, and (5) in openings isolated from active ground-water flow. These conceptual models are intended to be starting points for site-specific studies of chlorinated-solvent contamination in karst settings.

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# Emergency Response Sampling of Air from Caves Following a Diesel Fuel Release, Chickamauga-Chattanooga National Military Park, Lookout Mountain, Tennessee

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In February 1996, an interstate fuel pipeline rupture released an estimated 65,000 gallons of diesel fuel into a small ravine on the east side of Lookout Mountain at the Chickamauga-Chattanooga National Military Park in Hamilton County, Tennessee. A relatively small amount (about 1,500 gallons) of the fuel was recovered; the remainder infiltrated the ground and the cave system under Lookout Mountain. A cooperative study was conducted between the National Park Service Emergency Response Team and the U.S. Geological Survey in an effort to determine the spatial extent of the effect of the diesel fuel release on the cave system. As part of the emergency response, air flowing from cave entrances and cave vents at various locations and elevations on Lookout Mountain was sampled and analyzed using a portable gas chromatograph. Low concentrations of volatile hydrocarbons, including benzene, toluene, and xylene, were detected in several of the cave air samples. Chromatograms of the cave air samples were compared to a chromatogram of a "diesel fuel standard" in an attempt to "fingerprint" the air contaminants. Although "fingerprinting" the hydrocarbon contaminants detected in the cave air samples as a close match to the "diesel standard" was not possible, several of the hydrocarbons detected in the cave air samples also were present in the "diesel standard."

# Quantitative Approaches in Characterizing Karst Aquifers

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## Abstract

Karst aquifers are an important ground-water resource and are highly vulnerable to contamination due to relatively fast transport and limited attenuation processes. Quantitative understanding of karst hydrologic functions is integral to managing water resources and developing protection or remediation strategies. However, traditional methods of aquifer characterization and testing, based on Darcian approaches, provide misleading or inadequate quantitative data when applied to karst settings. This difficulty is partly a problem of scale (volume of aquifer tested), and partly a problem of the complex nature of the typical karst aquifer system.

Early approaches to studying karst concentrated on describing geomorphic features and their hydrologic functions, or understanding singular elements of karst flow dynamics, such as spring discharge, or hydraulic properties of solutional conduits. However, proper understanding of karst aquifers requires a systems approach in which the hydrologic function of each primary component—vadose zone, epikarst, and conduit network—is considered separately and as an integrated part of the whole system. The major difficulty facing the hydrologist is that karst aquifers typically exhibit dual ground-water flow regimes, that is, fast (conduit-dominated) flow and slow (diffuse) flow. In selecting investigative techniques to characterize properties of a karst aquifer, it is therefore important to determine how the data obtained by a particular test method are influenced by the fast-flow regime, slow-flow regime, or both. With this point in mind, several quantitative methods that are particularly useful in investigating the hydraulic parameters of the karst aquifer system are briefly discussed here.

Quantitative water-tracing tests, conducted with fluorescent dyes, are among the most useful types of field methods that can be employed in the investigation of a karst aquifer. A common misconception is that dye-tracing methods are too expensive, difficult, or unreliable, to use in many karst investigations. This is simply not the case. Like any other type of aquifer-testing technique, dye-tracing tests require careful planning and implementation, and a proper understanding of the applicability and limitations of the techniques and the data. One great advantage is that tracer tests can be designed and implemented to any field scale, and another is that the movement of the dye tracer almost exactly replicates the movement of water (and many dissolved solutes) through the aquifer.

Quantitative water-tracing tests require careful measurement of dye concentration at frequent sampling intervals and discharge through the sampled part of the aquifer, usually a spring. It is advisable that qualitative, or point-to-point tracer tests, using passive dye detectors and less frequent sampling, be conducted first to delineate ground-water flow paths, ensure that all potential dye-resurgence sites are known and sampled, aid in selecting the proper sampling frequency and duration, and assess the possible interference of ambient fluorescent solutes with detection and measurement of the tracer dye. The use of dye-tracing tests to delineate ground-water flowpaths and basin boundaries in karst aquifers is demonstrated by Bayless, Taylor, and Hopkins (1994) and Taylor and McCombs (1998).

The principal tool for analysis of quantitative water-tracing tests conducted with fluorescent dyes is the dye hydrograph, a specialized tracer breakthrough curve. The time of travel of the dye, indicated by the first detection of dye at concentrations above background fluorescence levels, provides a direct measurement of average ground-water velocity. The shape of the dye-hydrograph curve provides an indication of the dispersion of the dye as it migrates through the aquifer (Greene, 1999). For example, multiple peaks on a dye hydrograph may indicate splitting of the dye along multiple flow paths (conduits), or intermittent flushing of dye from hydraulic dead zones. Dye hydrograph analysis is particularly useful in contaminant transport investigations, because the tracer dye can be used as a surrogate pollutant. Estimates of peak contaminant concentration, persistence at concentrations that exceed quality criteria (such as maximum contaminant level, or MCL), and contaminant loading can be easily calculated for karst springs (Mull and others, 1988).

In Eve L. Kuniansky, editor, 2001, U.S. Geological Survey Karst Interest Group Proceedings, Water-Resources Investigations Report 01-4011, p. 164-166

Data obtained from quantitative dye-tracing tests can also be used to calculate a variety of parameters related to the geometry and hydraulic properties of any type of conduit network system. A good demonstration of the application of these techniques was presented by Fountain (1993) in a study of subglacial conduit networks. Examples of physical properties that can be calculated include conduit diameter, surface area, and hydraulic depth (assuming open-channel flow conditions). Estimates of fluid dynamic parameters that can be determined include the Peclet number, Reynolds number, Froude number, and hydraulic head loss. A summary of these methods and a software program that greatly facilitates the calculations involved in dye-hydrograph analysis was recently published by Field (1999).

Traditional aquifer tests can be used to estimate rates of ground-water movement and hydraulic properties such as transmissivity and storativity if special consideration is given to the dual-flow nature of karst aquifers while interpreting the aquifer test data. Recognition must be given to the fact that the framework of the karst aquifer is composed of integrated networks of fractures and solutional conduits of different sizes and interconnection. The aquifer test data represents a measurement of the composite hydraulic response of families of fractures and solutional conduits having different hydraulic characteristics (Streltsova, 1988). The larger solutional openings act collectively as the initial source of water being pumped during an aquifer test. Typically, these larger solutional openings are hydraulically connected to smaller, more diffuse sets of fractures in the aquifer. As the pumping continues, the fluid pressure in the larger solutional openings is reduced, resulting in hydraulic gradients which allow water in the diffuse fractures to provide recharge to the larger solutional openings. Thus, in describing the hydraulic properties of the karst aquifer, four physical parameters must be described: T, the transmissivity of the solutional openings; S, the storativity of the solutional openings; $S_f$, the storativity of the network of diffuse fractures; and $\beta$, the rate of fluid exchange between the network of fractures and the solutional openings (Greene, Shapiro, and Carter, 1999).

As a tool for simulation of flow and transport in karst aquifers, numerical models are frequently used. At present, great difficulties exist in accurately simulating karst flow systems at the local or subregional scale because of the difficulty in developing numerical models that realistically represent boundary conditions for conduit networks. Nevertheless, numerical models are among the best quantitative tools for gaining a better understanding of the functioning of individual karst hydrology components and for predicting how the system works as a whole. The two most common types of approaches can be classified as either a black box (or lumped parameter) approach or a distributed parameter approach. The black box approach uses techniques such as recession analysis and transfer/kernel functions to simulate karst aquifers. Several examples are shown where recession analysis is used to estimate regional hydraulic parameters and the volume of available ground-water resources. In addition, we demonstrate the use of kernel functions to interpret and simulate karst responses to precipitation.

The limitations of the black box (or lumped parameter) model approach become apparent when a known heterogeneity that has a physical basis needs to be modeled. Tracer-test results (velocity, breakthrough times) often show that slow-flow or fast-flow dominates different parts of the aquifer, for example, as ground water moves from the epikarst to the conduit network. Each of these aquifer components represents a particular heterogeneity that cannot be ignored in the modeling process. Thus, a distributed parameter model approach is used to incorporate known heterogeneities determined from field data. The three major distributed parameter approaches to describing flow include; 1) equivalent porous media, 2) discrete fracture, and 3) double porosity or double continuum approach. Examples are shown and each type of model is discussed.

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# A Multracer Approach for Assessing the Susceptibility of Ground-Water Contamination in the Woodville Karst Plain, Northern Florida

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## INTRODUCTION

The unconfined Upper Floridan aquifer (UFA) is highly susceptible to contamination in the Woodville Karst Plain (WKP). Sinkholes, solution conduits, and other karst features provide direct pathways for the flow of surface water into the UFA, the source of potable water in southern Leon and Wakulla Counties (fig. 1). Water also can infiltrate rapidly through a relatively thin layer of highly permeable sands that mantle the underlying limestone that makes up the UFA.

Stream waters typically contain high concentrations of dissolved organic carbon (DOC) with considerable amounts of colloidal organic compounds, such as tannins and lignins. When surface water recharges the aquifer near sinkholes, tannins and lignins are transported into the ground water system. Tannins and lignins in ground water used for public drinking water can react with chlorine during the disinfection process resulting in the production of trihalomethanes and other disinfection byproducts (Rostad and others, 2000). Some of these byproducts are suspected to be mutagenic (Kronberg and Christman, 1989) and are difficult to remove from the water. Also, elevated levels of bacteria from surface water enter the aquifer predominantly during wet periods. Generally, these problems have not affected the water quality in large parts of the UFA mainly due to the high dilution factor (Katz and others, 1997) and sorption of reactive colloidal organic (DBP-precursor) compounds on the limestone matrix (Rostad and others, 2000).

Nonpoint sources of contamination, however, have the potential to degrade ground-water quality over large parts of the WKP. Fertilizers, septic-tank effluent, and atmospheric deposition can contribute large amounts of nitrate to ground water, as documented by previous studies in northern Florida (Fu and Winchester, 1994; Katz and others, 1999).

This study uses a multitracer approach to examine the susceptibility of the UFA to contamination, with particular emphasis on nitrate and its sources in ground and surface water in the WKP. Naturally occurring isotopes and other chemical tracers are used to assess sources of water and solutes, and age (residence time) of water in the aquifer. Information provided by naturally occurring tracers on the age and sources of contamination of ground water has been shown to be very useful in assessing the susceptibility of karst aquifers to contamination from anthropogenic sources (Plummer and others, 1998; Cook and Bohlke, 1999).

## DESCRIPTION OF STUDY AREA

The Woodville Karst Plain is characterized as a flat or gently rolling surface of highly porous quartz sand overlying Oligocene and Miocene age limestones (St. Marks Formation) that make up the UFA (Rupert, 1988). Limestone is generally within 8 meters of the surface in most of the karst plain. The top of the limestone has undergone extensive dissolution by chemically aggressive waters, and as a result, the karst plain contains numerous wet and dry sinkholes, natural bridges, and disappearing streams (Rupert and Spencer, 1988). Sinkholes in the area typically have formed as a result of dissolution, subsidence, and collapse of the limestone bedrock.

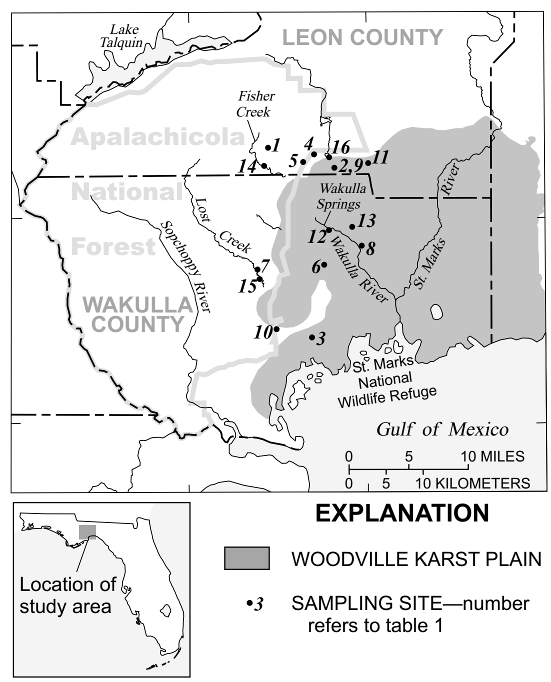


Figure 1. Study area showing location of sampled wells and streams in the Woodville Karst Plain.

Disappearing streams, such as Lost and Fisher Creeks, originate in the Apalachicola National Forest (Coastal Lowlands Physiographic Province), a flat, sandy part of western Leon and Wakulla Counties that

In Eve L. Kuniansky, editor, 2001, U.S. Geological Survey Karst Interest Group Proceedings, Water-Resources Investigations Report 01-4011, p. 167-176

is underlain by thick clastic (clays and sands) deposits. The lowlands area, which reaches elevations of nearly 46 meters in northwestern Wakulla County, is characterized by shallow "bays" (densely wooded, swamplike areas) and numerous creeks with poorly defined channels that drain the bays (Rupert and Spencer, 1988). Fisher and Lost Creeks receive water from smaller creeks flowing out of bays and swamps, and both creeks flow in a southeasterly direction through narrow. meandering valleys and into sinkholes that are connected to the UFA.

Lost and Fisher Creeks are referred to as blackwater streams due to the high concentrations of tannins and lignins dissolved in the water. Tannins and lignin in river water originate from a spectrum of compounds that possess carboxyl and phenolic groups on their aromatic structures (Tissot and Welte, 1984). Tannins are typically concentrated in the bark and leaves of higher plants, such as cypress trees and bottomland hardwoods, which are present in the wetlands where Lost and Fisher Creeks originate. Lignin, a high-molecular weight polyphenol consisting of units constructed from phenylpropane, is located between the cellulose micelles of supporting tissues of plants (Tissot and Welte, 1984).

Land use (1995) in the WKP is predominantly forested (42%). Other land-use types in order of decreasing percentages of the total are as follows: water and wetland (35), agriculture and rangeland (11), transportation (9), and urban (3). Land-cover classification was based on remote-sensing techniques (Vogelmann and others, 1998). Most of the population in Wakulla County is concentrated in the WKP because most of the western half of Wakulla County and southwestern Leon County lies in the Apalachicola National Forest and is not open to development.

The climate of the study area is humid subtropical, with a mean annual temperature of 19.6 $^{o}$C and a mean annual rainfall of 167 centimeters, for 1961-90, measured at the National Oceanic and Atmospheric Administration weather station at Tallahassee Regional Airport located in the northern part of the study area (Owenby and Ezell, 1992). The highest amount of monthly rainfall typically occurs during July and August, whereas the lowest typically occurs in April, October, and November.

Rainfall is the source of recharge to the UFA in the study area. Based on data from a recent study in which a calibrated ground-water flow model was developed for the UFA in Leon and Wakulla Counties and the surrounding area, recharge to the aquifer ranges from about 20 to 46 centimeters per year in the northern and southern parts of the study area, respectively (Davis, 1996). More recharge occurs in the southern part because there is very little direct surface runoff compared to the northern part. Recharge to the UFA probably occurs throughout the year, based on the close agreement between the mean annual air temperature and ground-water temperatures measured during sampling.

## METHODS

Several criteria were used to select ground- and surface-water sites (table 1). Wells were selected mainly from the western part of the WKP due to their proximity to sinking streams. Wells with a range of depths were selected to assess the vertical distribution and movement of contaminants in ground water. Water from Wakulla Springs was selected to represent ground-water discharge that is integrated both vertically and laterally from large parts of the UFA. McBride Slough is a discharge seep from the UFA. Lost and Fisher Creeks and Munson Slough are the largest streams that originate in wetland or highland areas outside of the WKP and flow directly into the UFA through sinkholes.

Table 1. Location of ground- and surface-water sites sampled in Woodville Karst Plain, well-construction information, and surface-water discharge

[Map No. is shown in fig. 1. Latitude and longitude are in degrees, minutes, seconds; well and casing depths measured in meters below land surface]

| Map No. | Site name | Latitude | Longitude | Well depth | Casing depth | Casing diameter, centimeters |
|---|---|---|---|---|---|---|
| ***Ground water sites*** | | | | | | |
| 1 | TPW-1 | 302008.47 | 842317.96 | 103.6 | 93.6 | 15 |
| 2 | DSW-1 | 301840.88 | 841738.27 | 24.4 | 17.1 | 10 |
| 3 | SCW-1 | 300618.00 | 841938.00 | 62.5 | 39.9 | 15 |
| 4 | FPW-1 | 301939.00 | 841924.00 | 61.0 | 34.1 | 15 |
| 5 | SPW-1 | 301905.00 | 842019.00 | 87.5 | 87.2 | 15 |
| 6 | SHADE-1 | 301135.00 | 841834.00 | 98.1 | 54.9 | 15 |
| 7 | AWC-1 | 301114.08 | 842412.23 | 39.3 | 22.9 | 15 |
| 8 | GDW-1 | 301258.45 | 841524.41 | 14.6 | 6.4 | 10 |
| 9 | DDW-1 | 301839.49 | 841738.07 | 47.9 | 34.4 | 10 |
| 10 | WPR-1 | 300654.83 | 842236.93 | 36.6 | 10.7 | 15 |
| 11 | WOOD-1 | 301859.59 | 841450.20 | 60.7 | 35.7 | 36 |
| **Map No.** | **Site Name** | **Latitude** | **Longitude** | **Discharge measurement date** | | **Discharge, meters cubed per second** |
| ***Springs or surface water sites*** | | | | | | |
| 12 | Wakulla Springs | 301407.00 | 841810.00 | 10/16/97 | | 8.779 |
| 13 | McBride Slough | 301421.00 | 841612.00 | 9/11/97 | | 0.484 |
| 14 | Fisher Creek | 301848.00 | 842336.00 | 9/4/97 | | 0.131 |
| 15 | Lost Creek | 301033.00 | 842401.00 | 9/4/97 | | 0.680 |
| 16 | Munson Slough below 8_mile Pond | 301926.00 | 841808.00 | 9/11/97 | | 0.002 |

Water samples from streams, springs, and wells were collected in September and October 1997 using standard techniques (Koterba and others, 1995). Water samples from 10 wells were collected after a minimum of three well-bore volumes of water had been purged and readings of specific conductance, pH, dissolved oxygen (DO), and temperature had stabilized. A closed flow-through chamber was used to measure these properties to prevent contact of the ground water with the

In Eve L. Kuniansky, editor, 2001, U.S. Geological Survey Karst Interest Group Proceedings, Water-Resources Investigations Report 01-4011, p. 167-176

atmosphere. At wells TPW-1, DSW-1, SCW-1, FPW-1, SPW-1, SHADE-1, and WOOD-1, water samples were collected from the existing pump and water-supply system upgradient of the pressure tank. Samples of water from Wakulla Springs were collected by lowering a positive-displacement, dual-piston pump about 15 meters into the spring vent. Water was pumped at about 0.06 liter per second through 0.63-centimeter copper (refrigeration grade) tubing. Samples of stream water were collected using a peristaltic pump, and sampling methodology varied somewhat depending on accessibility to the stream and site characteristics. Specific conductance, pH, DO, and temperature were measured at the time of sampling using either a mutiprobe unit that was lowered directly into the spring vent or streamwater column or a closed flow-through chamber.

Samples of ground water were collected for major element chemistry, DOC, environmental isotopes, and tritium/helium-3 ($^3H/^3He$). Samples of surface water were collected for the previously listed constituents with the exception of $^3H/^3He$. Samples for major ions were preserved in the field and analyzed in the U.S. Geological Survey laboratory in Ocala, Fla., using standardized procedures (Fishman, 1993). Concentrations of tannin and lignin in river water and ground water were measured in the field using a colorimetric procedure (tyrosine method) with a portable spectrophotometer (Hach Company, 1989). This method detects all hydroxylated aromatic compounds, including tannin, lignin, phenol and cresol (Kloster, 1974). Results are expressed as milligrams per liter tannic acid, with a method detection limit of about 0.1 mg/L and a 1 standard deviation ($\sigma$) precision of 0.1 mg/L.

The stable isotopes of oxygen ($^{18}O/^{16}O$), hydrogen ($^2H/H$), carbon ($^{13}C/^{12}C$), and nitrogen ($^{15}N/^{14}N$) were used to determine the origin of water and solutes, and to identify biogeochemical processes that control the chemical composition of ground water. Standard delta ($\delta$) notation (Gonfiantini, 1981) was used for the stable isotopes, $^{18}O/^{16}O$, $^2H/H$, $^{13}C/^{12}C$, and $^{15}N/^{14}N$. Oxygen- and hydrogen-isotope results are reported in per mil relative to Vienna Standard Mean Ocean Water and are normalized on scales such that the oxygen and hydrogen isotopic values of Standard Light Antarctic Precipitation are -55.5 and -428 per mil, respectively (Coplen, 1994). The $2\sigma$ precision of $\delta^{18}O$ and $\delta^2H$ results is 0.2 and 1.5 per mil, respectively. The analysis of $\delta^{13}C_{DIC}$ of dissolved inorganic carbon (DIC) in ground and surface water involved the precipitation of the DIC using $NH_4OH$-$SrCl_2$, followed by filtering, drying, and acidifying the $SrCO_3$ precipitate to produce $CO_2$, which was analyzed by mass spectrometric methods (Gleason and others, 1969). The $2\sigma$ precision for the analytical procedure is 0.2 per mil (Coplen, 1994). Values of $\delta^{13}C$ of DIC are reported relative to Vienna Pee Dee Belemnite (Coplen, 1994).

Tritium ($^3H$) and its daughter product of radioactive decay, tritiogenic $^3He$, were used to estimate the age of ground water by comparing measured $^3H$ concentrations in ground water with the long-term $^3H$ input function of rainfall measured at the International Atomic Energy Agency precipitation monitoring station in Ocala, Fla., (Michel, 1989). The term “age” used in this paper is actually an apparent age because it could be affected by mixing or other physical processes. This age represents the time elapsed since recharge and isolation of the recharge water from the modern atmosphere. Combined measurements of $^3H$ and $^3He$, define a relatively stable tracer of the initial $^3H$ input to ground water, which can be used to calculate the $^3H/^3He$ age from a single water sample (Schlosser and others, 1988, 1989). The $^3H/^3He$ age is based on a He-isotope mass balance used to calculate the amount of tritiogenic and nontritiogenic $^3He$ in the sample. Nontritiogenic $^3He$, which generally is negligible in a shallow aquifer with predominantly young water, is corrected for by measuring the concentrations of $^4He$ and neon in the water sample and assuming solubility equilibrium with air at the water temperature measured during sampling (Schlosser and others, 1988, 1989). $^3H/^3He$ ages generally are not affected by contamination, sorption, and microbial degradation processes that can alter the concentrations of other transient tracers, such as chlorofluorocarbons (Plummer and others, 1998). The distribution of $^3H$ and $^3He$ can, however, be affected by hydrodynamic dispersion and mixing of different age waters (Solomon and Sudicky, 1991, Reilly and others, 1994).

Ground-water samples for $^3H/^3He$ analyses were collected in pinch-off copper tubes (10 millimeter diameter, 80 centimeter length, about 40-milliliter volume) for the determination of $^3H/^3He$, $^4He$, and Ne (neon) while applying back pressure to the discharge from the sample tube to prevent formation of gas bubbles during sample collection. These samples were analyzed at the Noble Gas Laboratory of Lamont-Doherty Earth Observatory, Palisades, N.Y., using quantitative gas extraction followed by mass spectrometric techniques (Schlosser and others, 1989). Surface-water samples were collected and analyzed for $^3H$ according to methods presented by Michel (1989). Tritium activity is reported in tritium units (TU) (1 TU is equal to 1 $^3H$ atom in $10^{18}$ hydrogen atoms and 7.1 disintegrations per minute per gram of water) with a $1\sigma$ precision of <10% for waters containing >2 TU.

## CHEMICAL AND ISOTOPIC COMPOSITION OF GROUND AND SURFACE WATER

Water from the UFA in the WKP is a calcium-bicarbonate type with generally low dissolved-solids (DS) concentrations ranging from 130 to 274 mg/L (table 2). Water from WOOD-1 had a higher DS concentration of 432 mg/L; this public-supply well draws ground water from beneath residential land that could contribute elevated concentrations of solutes. DO concentrations were highly variable in shallow and deep parts of the aquifer; water from three of six deep wells (depths >60 meters) contained low DO concentrations, typically >0.5 mg/L. Saturation indices of water with respect to calcite (calcite SI) and dolomite (dolomite SI) (table 2) from the shallow part of the flow system and Wakulla Springs typically were <0, indicating that the waters generally are slightly undersaturated with these two minerals, which make up the limestone matrix of the UFA. Water from deeper parts of the aquifer tend to have higher saturation indices with respect to calcite and dolomite, possibly indicating the longer residence time of water in the ground-water flow system.

DOC concentrations in surface water ranged from 11 to 31 mg/L (table 2); however, DOC concentrations in ground water typically were much lower than in surface water. Water samples from six wells, Wakulla Springs, and McBride Slough were equal to or <0.6 mg/L (analytical method reporting limit is 0.1 mg/L) (table 2). Water samples from DSW-1, SCW-1, SHADE-1, AWC-1, and DDW-1, however, contained DOC concentrations above 1 mg/L, indicating that surface water containing elevated DOC concentrations probably is locally recharging the aquifer system and mixing with ground water. The fraction of colloidal organic carbon of the DOC was high in Fisher Creek (23%) and Lost Creek (30%), but was considerably lower in ground-water samples from DSW-1 (1.6%) and TPW-1 (3.6%) (Rostad and others, 2000). The large reduction in DOC and its colloidal fraction from surface water to ground water likely indicates significant biogeochemical attenuation of organic compounds by biodegradation, sorption, precipitation, and/or dilution processes (Rostad and others, 2000).

Delta $^{13}C_{DIC}$ values of ground water, which ranged from -16.8 to -12.8 per mil (table 2), are higher (enriched) relative to water from Fisher and Lost Creeks (-24.7 per mil). The $\delta^{13}C_{DIC}$ values of water from the UFA likely represent a mixture of DIC from two sources: $CO_2$ respired by microorganisms from the oxidation of organic carbon in the unsaturated zone (Deines, 1980), and dissolution of calcite in the aquifer matrix. Calcite dissolution in the aquifer is indicated by the: (1) substantial increase in $\delta^{13}C_{DIC}$ in water from the UFA compared to water from Fisher and Lost Creeks and the surficial aquifer (Katz and others, 1997), (2) large increase in the saturation index of ground water with respect to calcite compared to that of water from Lost and Fisher Creeks (table 2), and (3) large increase in pH and calcium and bicarbonate concentrations (table 2).

Differences in values of $\delta^2H$ and $\delta^{18}O$ in ground water and stream water provide information about the source(s) of water and possible mixing of ground water with surface water. Stable H and O isotopic values for most ground-water samples plot along the global meteoric water line (Craig, 1961), as do samples of rainfall from northwest of the study area (Katz and others, 1997), indicating that they are not affected by evaporation (fig. 2). $\delta^2H$ and $\delta^{18}O$ values for water from DDW-1 were slightly enriched relative to other ground-water samples and indicate that mixing with surface water affected by evaporation is likely. Values of $\delta^2H$ and $\delta^{18}O$ were similar for water from Fisher and Lost Creeks and most ground-water samples, indicating that shallow ground-water discharge is a major contributor to streamflow. In contrast, $\delta^2H$ and $\delta^{18}O$ values of water from Munson Slough (table 2) were highly enriched relative to most ground-water samples and were displaced to the right of the global meteoric water line, indicating that Munson Slough receives water that has been affected by evaporation. Similarly enriched values of $\delta^2H$ and $\delta^{18}O$ were found for other ground- and surface-water sites in the northern part of the WKP, which were attributed to evaporation or mixing with surface water with an enriched isotopic signature (Katz and others, 1997).

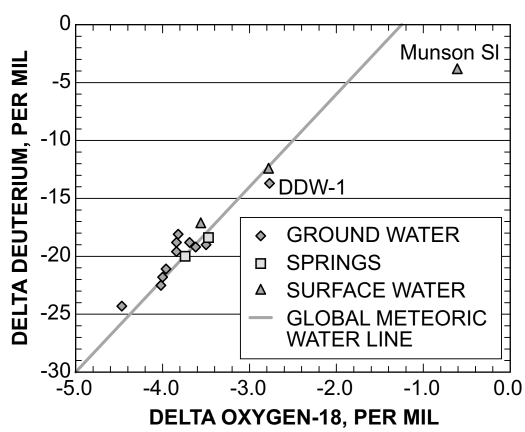


Figure 2. Deuterium and oxygen-18 content of ground- and surface-water samples, relative to the global meteoric water line, Woodville Karst Plain.

Nitrate-N concentrations in ground water were highly variable among sites. There was no discernible relation between nitrate concentration and well depth or top of open interval (depth to bottom of casing). For instance, water from deep wells, FPW-1, SPW-1, and WOOD-1, had nitrate-N concentrations that ranged from 0.22 to 0.71 mg/L, whereas water from other deep wells and wells open to shallow zones in the UFA had nitrate-N concentrations <0.02 mg/L (table 2). Background nitrate-N concentrations in ground water from northern Florida were <0.05 mg/L (Katz, 1992; Maddox and others, 1992). Water from Wakulla Springs and McBride Slough had elevated nitrate-N concentrations of 0.89 and 0.47 mg/L, respectively.

Nitrate concentrations in ground water were related to concentrations of DO and DOC. There was a statistically significant ($p<0.05$) correlation (Spearman's Rho nonparametric statistic) between nitrate-N and DO concentrations, and a significant ($p<0.001$) inverse correlation between nitrate and DOC concentrations.

Nitrate-N concentrations were at or below 0.02 mg/L in water samples from Lost Creek, Fisher Creek, and Munson Slough (table 2). Ammonium-N concentrations were higher in surface-water samples compared to ground-water samples and ranged from 0.03 to 0.28 mg/L. Two ground-water samples had elevated ammonium-N concentrations, TPW-1 and DDW-1. Both sites likely have direct connections to surface water, based on elevated DOC concentrations. Low concentrations of nitrate-N in surface water indicate that sinking streams are unlikely to contribute to the elevated concentrations of nitrate in ground water in the WKP.

$\delta^{15}N\text{-}NO_3$ values in ground water were variable, ranging from 1.7 to 13.8 per mil (table 2). Some of these variations can be attributed to local differences in land use and nitrate sources. For example, well WPR-1 is located in a county park adjacent to a turf-grass field, which receives artificial fertilizer. Water from this well had a $\delta^{15}N\text{-}NO_3$ value of 1.7 per mil, which was similar to values observed elsewhere where nitrate in ground water is recharged beneath cultivated land receiving artificial fertilizers (Heaton, 1986; Kendall and Aravena, 1999). In contrast, water from well GDW-1 had a $\delta^{15}N\text{-}NO_3$ value of 13.8 per mil, which is indicative of human or animal waste sources that commonly yield ground-water nitrate with $\delta^{15}N$ values about 8 per mil or more (Heaton, 1986; Kendall and Aravena, 1999). GDW-1 is a shallow residential well that is located near a septic-tank system. Water from other wells and Wakulla Springs had $\delta^{15}N\text{-}NO_3$ values of 6.8 to 8.9 per mil (table 2), which indicate that nitrate likely originates from a mixture of inorganic and organic sources of nitrogen. Delta $^{15}N\text{-}NO_3$ values could not be determined in ground-water samples from six sites because nitrate-N concentrations were near or below analytical detection limit.

Denitrification can decrease the concentrations and increase the $\delta^{15}N$ values of nitrate in ground water (Korom, 1992). Denitrification, however, is unlikely in shallow ground water from the WKP due to DO concentrations >1 mg/L in most ground-water samples. Conversely, water from deeper wells had lower DO concentrations (generally <0.5 mg/L) and denitrification could not be ruled out given the low DO concentrations and elevated concentrations of DOC (table 2).

## AGES AND RESIDENCE TIMES OF WATER FROM THE UPPER FLORIDAN AQUIFER

Tritium/$^3He$ measurements, error estimates, and $^3H/^3He$ ages for water samples from the UFA are summarized in table 3. Ground-water ages ranged from 5.2 years (DSW-1) to 25.5 years (AWC-1). For Wakulla Springs, the apparent age (38.7 years) more realistically represents the average residence time of ground water that discharges from the UFA because springs integrate water from converging flow paths of varying age. The relatively small $1\sigma$ variations in age represent analytical uncertainty (table 3).

Other factors could result in additional uncertainty in age estimates. For instance, $^3H/^3He$ ages were calculated using water temperatures measured at the time of sampling (table 2) to calculate concentrations of He and Ne in water in solubility equilibrium with air. Variations in recharge temperature (21 $^oC$) of about $\pm 2\ ^oC$ (mean annual air temperature is 19.6 $^oC$) would not significantly affect the calculated ground-water age (Plummer and others, 1998). The calculated age, however, is more sensitive to the fraction of terrigenic He (from excess air and decay of U-series radionuclides). Some ground-water samples contained elevated values of terrigenic $^4He$ (table 3); for instance, the percent of terrigenic $^4He$ of the total $^4He$ was 83.6% for water from Wakulla Springs, 26% from SCW-1, and 47% from SHADE-1. With large fractions of terrigenic helium, the $^3He/^4He$ ratio of the terrigenic He, $R_{terr}$, must be accurately known. Nontritiogenic $^3He$ resulting from terrigenic sources were adjusted using an $R_{terr}$ of $2 \times 10^{-8}$ (Schlosser and others, 1988). If this ratio were allowed to vary over 2 orders of magnitude (somewhat unlikely, but useful for illustrating estimates of age uncertainty), ground-water ages would be younger by <5 years for SHADE-1, SCW-1, SPW-1, and FP-1 using $R_{terr} = 2 \times 10^{-7}$ and there would be no appreciable age difference for $R_{terr} = 2 \times 10^{-9}$. For ground water discharging from Wakulla Springs, the minimum age would be 21 years ($R_{terr} = 2 \times 10^{-7}$), compared to 38.7 years using ($R_{terr} = 2 \times 10^{-8}$ and $2 \times 10^{-9}$). The large amount of terrigenic He at Wakulla Springs may result from the use of He-mixtures of gas used by scuba divers for cave exploration.

Since the $^3H/^3He$ dating method is independent of the mixing fractions of young surface water with ground water (Plummer and others, 1998) or younger ground water that mixes in the well bore during pumping, the $^3H/^3He$ age represents the age of the youngest fraction in the mixture. Consequently, this apparent age (or average residence time in the case of spring waters) provides a realistic assessment of the susceptibility of the UFA to contamination by approximating the travel time for contaminants to reach a particular zone in the aquifer. If water in the UFA were vertically stratified by age, relatively young water (0-5 years) from the top of the open interval is assumed to mix with older water (>10 years) from greater depths in the borehole during sampling of the well. However, this assumption is complicated by ground-water flow paths of varying ages.

## Relation Between Age of Ground Water and Other Variables

There was no discernable relation between the $^3H/^3He$ age of ground water and the depth of wells or depth to the top of the open interval. Although there was a positive correlation (Spearman's Rho nonparametric statistic) between ground-water ages and depth of well (r=0.10) and depth to top of casing (r=0.13), these correlations were not statistically significant ($p<0.05$). Likewise, there were correlations between ground-water age and DS (r=0.25), calcite SI (r=0.15), and dolomite SI (r=0.23), but none were statistically significant. Inverse correlations were found between ground-water age and DOC (r= –0.30; p=0.39) and DO (–0.15, p=0.66).

Important implications can be drawn about ground-water age and water chemistry even though most correlations were not statistically significant, which could be due in part to the relatively small sample size (n=11). For instance, younger ground water (<10 years) tends to have higher DOC concentrations, lower DS, and lower saturation indices with respect to calcite and dolomite, and probably represents a mixture of ground and surface water. Older waters (>20 years) tend to have higher DS, lower DOC concentrations, and higher mineral saturation indices, all characteristics of ground water with longer residence times in the aquifer. Also, the inverse trend between ground-water age and DO concentrations indicates that oxygen has been consumed in water with longer aquifer residence times. Similar inverse trends (but statistically significant) were found between the age of spring waters and DO concentrations in the Suwannee River Basin of northern Florida (Katz and others, 1999).

## Tritium in Surface Water and Ground Water

Tritium concentrations in water from Lost Creek, Fisher Creek, and Munson Slough, 5.4, 4.4, and 4.9 TU, respectively, were very similar to those measured in most ground-water samples (table 2). This finding supports the contention that ground water is a major contributor to base flow for Fisher and Lost Creeks based on the similarity between the stable isotopic composition of ground and surface water. Water from Munson Slough below 8-Mile Pond has undergone extensive evaporation based on its stable isotopic composition and has similar tritium concentrations to the other two sampled surface waters. Water samples were collected during base-flow conditions and could represent a mixture of ground and surface water draining from wetlands or ponds.

The close similarity among $^3H$ concentrations for ground water, surface water, and rainfall (Katz and others, 1997) indicates that rainfall recharges the shallow ground-water system and moves through the UFA very rapidly, <10 years for parts of the shallow flow system and <30 years for deeper parts of the aquifer. $^3H$ concentrations in ground and surface water in the WKP reflect the passing of the $^3H$ transient through the hydrologic system.

Prior to the advent of the atmospheric testing of fusion weapons in 1953, $^3H$ concentrations were on the order of 2 TU or less in this region (Thatcher, 1962). Atmospheric weapons testing increased $^3H$ concentrations in this area to a maximum of several hundred TU during the mid-1960s, followed by a sharp decline in concentrations after the nuclear testing moratorium (fig. 3). Estimates of $^3H$ concentrations in precipitation during the post-nuclear testing period were derived for the study area using data from the long-term station in Ocala, Fla., and the Ottawa correlation (Michel, 1989). $^3H$ concentrations in rainfall have declined rapidly and have been <10 TU for over a decade. $^3H$ concentrations in pre-bomb (1953) recharge would have decayed to 0.2 to 0.8 TU by 1997, assuming estimates of 2 TU (Thatcher, 1962) to 8 TU (Plummer and others, 1993) in pre-bomb rainfall and a $^3H$ half-life of 12.43 years. Therefore, based on elevated tritium concentrations (table 2), waters sampled from the WKP is of relatively recent origin, and almost certainly from the period of the falling limb of the tritium transient (post-1963). To further resolve and constrain the average residence time of ground water that discharges to stream waters, additional $^3H$ data would need to be collected over a range of hydrologic conditions.

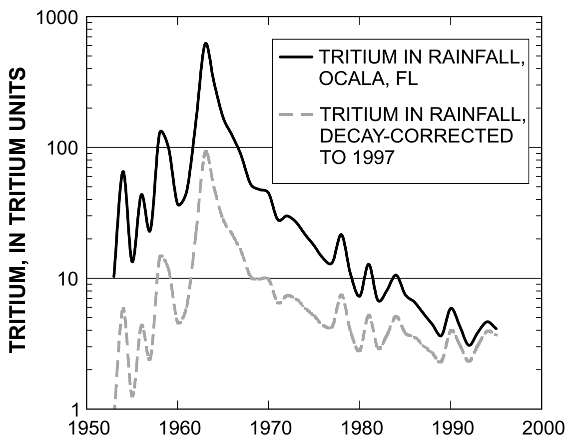


Figure 3. Concentrations of tritium in rainfall collected at Ocala, Fla. (Decay corrected to 1997.)

Table 2. Chemical characteristics, concentrations of selected chemical constituents, calcite, and dolomite saturation index for ground and surface water in Woodville Karst Plain

[ºC, degrees Celsius; DOC, dissolved organic carbon, SI, saturation index; TU, tritium units; ND, not determined. Concentrations of elements and chemical species are in milligrams per liter unless otherwise noted; specific conductance is in micosiemens per centimeter]

| Site name | Sample date | Temper-ature, °C | pH | Specific conduc-tance | Dissolved oxygen | Calcium | Magnesium | Bicarbonate | Nitrate-nitrogen | Ammonia- | DOC | Calcite SI | Dolomite SI | Dissolved solids | Delta C-13, per mil | Delta N-15, per mil | Delta O-18, per mil | Delta Deuterium, $^2H$, per mil | Tritium, TU |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| TPW-1 | 9/3/97 | 22.0 | 7.61 | 265 | 0.30 | 47.4 | 1.8 | 124 | <0.02 | 0.45 | 0.6 | -0.08 | -1.27 | 222 | -13.0 | ND | -3.8 | -18.8 | 1.8 |
| DSW-1 | 9/3/97 | 21.1 | 7.15 | 251 | 4.90 | 45.6 | 1.9 | 149 | <0.02 | <0.01 | 2.2 | -0.47 | -2.03 | 219 | -16.8 | ND | -3.5 | -19.0 | 4.0 |
| SCW-1 | 9/10/97 | 21.9 | 7.86 | 355 | 0.20 | 48.7 | 3.5 | 151 | <0.02 | 0.04 | 1.0 | 0.26 | -0.32 | 267 | -12.8 | ND | -4.0 | -22.5 | 4.0 |
| FPW-1 | 9/10/97 | 22.7 | 8.11 | 147 | 4.90 | 24.0 | 2.3 | 85 | 0.48 | 0.01 | 0.4 | 0.02 | -0.67 | 130 | -12.9 | 6.8 | -4.0 | -21.8 | 4.4 |
| SPW-1 | 9/10/97 | 21.3 | 7.65 | 202 | 1.50 | 27.4 | 6.2 | 112 | 0.22 | 0.01 | ND | -0.30 | -0.95 | 162 | -13.1 | 8.9 | -3.6 | -19.2 | 5.2 |
| SHADE-1 | 9/10/97 | 21.7 | 7.93 | 300 | 0.20 | 47.2 | 7.1 | 173 | <0.02 | 0.02 | 1.7 | 0.37 | 0.22 | 257 | -14.6 | ND | -3.7 | -18.8 | 4.6 |
| AWC-1 | 10/6/97 | 23.3 | 7.36 | 274 | 0.40 | 52.9 | 0.7 | 167 | <0.02 | 0.04 | 1.2 | -0.12 | -1.79 | 239 | -14.3 | ND | -3.8 | -19.6 | 7.6 |
| GDW-1 | 9/24/97 | 21.6 | 7.53 | 306 | 0.50 | 52.1 | 5.4 | 178 | 0.18 | 0.01 | 0.4 | 0.03 | -0.62 | 263 | -15.0 | 13.8 | -3.8 | -18.1 | 4.5 |
| DDW-1 | 9/24/97 | 22.5 | 7.49 | 155 | 0.30 | 23.8 | 3.2 | 88 | <0.02 | 0.09 | 6.1 | -0.59 | -1.74 | 134 | -14.3 | ND | -2.8 | -13.7 | 4.8 |
| WPR-1 | 10/15/97 | 22.3 | 7.63 | 194 | 7.20 | 34.4 | 0.7 | 85 | 3.10 | <0.01 | <0.1 | -0.32 | -2.04 | 151 | -14.9 | 1.7 | -4.5 | -24.3 | ND |
| WOOD-1 | 10/8/97 | 22.8 | 7.31 | 497 | 2.20 | 62.2 | 24.6 | 310 | 0.71 | 0.01 | <0.1 | 0.10 | 0.12 | 432 | -14.9 | 7.8 | -4.0 | -21.1 | 4.2 |
| Wakulla Springs | 10/8/97 | 21.1 | 7.20 | 312 | 2.20 | 42.1 | 9.8 | 168 | 0.89 | 0.01 | <0.1 | -0.42 | -1.18 | 259 | -13.7 | 6.8 | -3.5 | -18.4 | 4.3 |
| McBride Slough | 9/11/97 | 20.0 | 7.54 | 203 | 3.37 | 50.5 | 7.1 | 180 | 0.47 | 0.02 | 0.6 | 0.01 | -0.56 | 274 | -14.3 | 7.1 | -3.7 | -20.0 | 4.9 |
| Fisher Creek | 9/4/97 | 23.9 | 3.91 | 52 | 6.90 | 0.3 | 0.3 | ND | <0.02 | 0.08 | 21.0 | -10.40 | -20.50 | 42 | -24.7 | ND | -3.6 | -17.1 | 4.4 |
| Lost Creek | 9/4/97 | 24.8 | 4.66 | 38 | 7.00 | 3.6 | 0.3 | ND | <0.02 | 0.03 | 31.0 | -7.76 | -16.20 | 54 | -24.7 | ND | -2.8 | -12.4 | 5.4 |
| Munson Slough below 8-Mile Pond | 9/11/97 | 23/6 | 6.66 | 71 | 2.60 | 12.5 | 2.5 | 51 | <0.02 | 0.28 | 11.0 | -7.89 | -4.15 | 94 | -17.8 | ND | -0.6 | -3.8 | 4.9 |

Table 3. Summary of $^3H/^3He$ results for ground-water samples, Woodville Karst Plain

[TU, tritium units, $cm^3$ STP/g, cubic centimeters at standard temperature and pressure per gram of water; σ, standard deviation]

| Site name | Sampling date | $^3H \pm 1\sigma$, TU | | $^3He$ trit ± 1σ, TU | | $\delta^3He \pm 1\sigma$, percent | | $^4He$ $10^{-8}$ | Ne, $cm^3$STP/g | $^4He$-terrigenic, percent of $^4He$ | Apparent $^3H/^3He$ age ± 1σ, years | |
|---|---|---|---|---|---|---|---|---|---|---|---|---|
| TPW-1 | 09/10/97 | 1.78 | ±0.27 | 2.15 | ±0.11 | 4.84 | ±0.31 | 6.38 | 25.85 | -4.2 | 5.8 | ±1.55 |
| DSW-1 | 09/10/97 | 3.96 | ±0.26 | 2.16 | ±0.09 | 6.04 | ±0.31 | 5.14 | 21.24 | -3.0 | 5.2 | ±0.49 |
| | | 4.10 | ±0.08 | | | | | | | | | |
| SCW-1 | 09/10/97 | 3.95 | ±0.08 | -2.29 | ±0.15 | -6.7 | ±0.36 | 7.30 | 21.63 | 25.7 | 19.8 | ±2.29 |
| | | 4.03 | ±0.16 | | | | | | | | | |
| FPW-1 | 09/10/97 | 4.15 | ±0.20 | 7.25 | ±0.09 | 23.1 | ±0.31 | 5.31 | 21.83 | -4.0 | 16.2 | ±0.56 |
| | | 4.62 | ±0.09 | | | | | | | | | |
| SPW-1 | 09/10/97 | 5.14 | ±0.14 | 13.50 | ±0.09 | 39.48 | ±0.27 | 5.97 | 22.70 | 23.1 | 24.4 | ±0.36 |
| | | 5.22 | ±0.10 | | | | | | | | | |
| SHADE-1 | 09/10/97 | 4.62 | ±0.16 | -18.78 | ±0.11 | -28.82 | | 11.98 | 24.80 | 47.1 | 23.1 | ±0.82 |
| | | 4.62 | ±0.16 | | | -25.52 | | 10.88 | 22.65 | 47.5 | 22.4 | ±0.86 |
| AWC-1 | 10/06/97 | 7.53 | ±0.15 | 25.60 | ±0.11 | 93.31 | ±0.23 | 4.84 | 20.47 | -5.9 | 25.5 | ±0.33 |
| | | 7.63 | ±0.18 | | | | | | | | | |
| GDW-1 | 09/24/97 | 4.42 | ±0.09 | 5.92 | ±0.11 | 11.85 | ±0.24 | 8.33 | 31.07 | 2.2 | 16.8 | ±0.40 |
| | | 4.47 | ±0.15 | | | | | | | | | |
| DDW-1 | 09/24/97 | 4.64 | ±0.09 | 2.90 | ±0.11 | 8.8 | ±0.24 | 5.05 | 19.93 | 1.9 | 9.5 | ±0.26 |
| | | 4.92 | ±0.15 | | | | | | | | | |
| WOOD-1 | 10/08/97 | 4.18 | ±0.08 | 4.17 | ±0.11 | 11.97 | | 5.62 | 22.93 | -3.7 | 9.8 | ±0.17 |
| | | 4.18 | ±0.08 | | | | | | | | | |
| Wakulla Springs | 10/08/97 | 4.08 | ±0.08 | | | -67.46 | ±1.00 | 37.09 | 23.95 | 83.6 | 38.7 | ±1.07 |
| | | 4.40 | ±0.15 | | | | | | | | | |

## MIXING OF SURFACE WATER AND GROUND WATER

A two-component mixing model was used to estimate the fraction of surface water in the ground- and surface-water mixture in water from wells with elevated DOC concentrations. DOC concentrations were effective tracers of mixtures of river water with water from the UFA (Katz and others, 1998; Plummer and others, 1998). The fraction of surface water ($F_{sw}$) in the mixture is defined as: $F_{sw}= (Y_m-Y_{gw})/(Y_{sw}-Y_{gw})$, where $Y_m$, $Y_{gw}$, and $Y_{sw}$ denote the DOC concentrations in the mixture, ground water, and surface water, respectively.

DOC concentrations of the two end members were 20 mg/L for surface water (average of DOC concentrations for Fisher and Lost Creeks, and Munson Slough) and 0.10 mg/L for ground water. The mixing calculations assumed that DOC concentrations were not modified by reactions after mixing had occurred. For ground-water samples with DOC concentrations >0.1 mg/L, $F_{sw}$ values ranged from 0.015 at GDW-1 and FPW-1 to 0.30 at DDW-1. The higher $F_{sw}$ value at DDW-1 indicates the likelihood of a direct connection to a nearby sinkhole containing tannic water that contributes to the relatively young ground-water age at this site. The very young (5.8 years) water at TPW-1 (104 meters deep) is surprising; however, the $F_{sw}$ value of 0.025 at TPW-1 indicates that there likely is a direct connection to a surface-water feature at this site.

## SUSCEPTIBILITY OF THE AQUIFER TO CONTAMINATION

The use of naturally occurring isotopic and other chemical tracers in water from wells provides information about the susceptibility of the UFA to contamination on a local or site-specific scale that cannot be determined from other methods, such as mapping of surface karst features (Maddox, 1998). Even if solution features (sinkholes) were identified on a map, the susceptibility of a zone of interest in the aquifer near that feature could only be established by chemical data from a sampled well. Tracer data for Wakulla Springs, a first-magnitude spring, provide an assessment of regional impacts on water quality. Even though sinkholes provide direct pathways for water to enter the UFA from disappearing streams, such as Fisher and Lost Creeks, nitrate concentrations in stream water are at or below detection limits. Where direct connections exist between sinkholes and zones of the UFA, elevated concentrations of DOC from the tannic-rich surface waters result in elevated DOC concentrations in the ground water. Reactions between these naturally occurring organic compounds and chlorine during disinfection of the water for public use can result in the formation of harmful products such as trihalomethanes (Rostad and others, 2000).

Nitrogen-isotope data for ground-water samples indicate that nitrate in water that recharges the aquifer originates from nonpoint sources, such as leachate from fertilizers and septic tanks, and not from sinking streams. Nitrate-N concentrations in Wakulla Springs increased from 0.25 mg/L in 1972 (Rosenau and others, 1977) to 0.86 mg/L in 1997. This large increase in nitrate over 25 years, coupled with a delta $^{15}N-NO_3$ of 6.8 per mil, indicates that ground water draining the large regional area likely is contaminated with nitrate from several nonpoint sources.

Direct connections between ground water and surface water are indicated by elevated DOC concentrations. The high susceptibility of ground water to contamination and the presence of a dynamic flow system in the WKP are further indicated by young ages ($^3H/^3He$) for ground water of <10 years from both shallow and deep parts of the aquifer. $^3H/^3He$ ages define the age of the young fraction(s) in the mixture, which can result from mixing of surface water and/or recently recharged ground water near the water table. Ongoing studies in the WKP are assessing the spatial and vertical distribution of nitrate in ground water. Also, chemical tracer data for ground water along flow paths are being used to assess residence time of ground water and biogeochemical processes during various hydrologic conditions.

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# The Use of Artificial Tracers to Determine Ground-Water Flow and the Fate of Nutrients in the Karst System Underlying the Florida Keys.

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## Abstract

To determine the fate and movement of sewage derived contaminants and their possible interaction with surface waters in the Florida (USA) Keys, several types of experiments were conducted using $SF_6$ as an artificial tracer. The first type of experiment examined fluid flow from septic tanks placed in Miami Oolite on Big Pine Key, where there is a shallow freshwater lens overlying saline ground waters. Here ground water transport rates were constrained to be between 0.11 to 1.87 m/hr, traveling in an easterly direction (Dillon et al., 1999). The second type of experiment took place on Key Largo where there is no freshwater lens and the matrix of the aquifer is solely the more porous Key Largo limestone (KLL). Here we injected the tracer into a shallow well, which was screened from 0.6 to 10 m. This allowed us to evaluate groundwater movement in the shallow upper portion of the aquifer, the area to which inputs by septic tanks occur. Groundwater transport rates in the Upper Keys were as great as 3.7 m/hr and were controlled by the Atlantic tide (Dillon et al., 1999). $SF_6$ laden groundwater plumes moved back and forth due to tidal pumping. $SF_6$ reached nearby surface waters within 8 hours. Our results indicate that wastewater injected into the shallow subsurface can travel rapidly and may reach marine surface waters within a few hours.

Three dual tracer experiments were conducted on Long Key, Florida USA to examine the fate of waste water following sewage disposal in 10 to 30 m deep injection wells. This waste disposal practice introduces extraordinary amounts of nutrients into the ground waters of the Florida Keys. In three experiments, artificial ground water tracers, sulfur hexafluoride ($SF_6$) and radioiodine ($^{131}I$) were used to determine transport rates and directions of soluble non reactive substances injected into the saline ground waters underlying the Keys. Simultaneously, reactive tracers (bulk unlabeled phosphate, $PO_4$, and nitrate, $NO_3$, and radio-labeled phosphate ($^{32}PO_4$) were also added to determine the fate of nutrients in the subsurface. Two types of transport were observed: (1) rapid flow (0.20 - 2.20 m/hr) presumably due to the many conduits present in the limestone and (2) slow diffusive flow (< 0.003 - 0.14 m/hr) associated with the limestone's primary porosity (Dillon et al., 2000). Vertical flow was comparable to horizontal flow due to either the density driven buoyancy of the waste water plume or to preferential flow paths which allow upward advection or combination of both. These experiments showed that conservative artificial tracers injected into the subsurface reach surface waters in a matter of days and can remain in the immediate vicinity of the injection well for several months.

At this low discharge site (2600 L/day) the reactive tracers' behaviors in the subsurface indicate that $PO_4$ and $NO_3$ are both partially removed from solution in the subsurface. Phosphate showed an initial rapid uptake followed by a slower removal, caused by adsorption-desorption reactions with the KLL (Corbett et al., 2000). Based on our observations, we estimate that approximately 95% of the PO4 injected into the subsurface could be removed in 20-50 hours. There was also evidence for some removal of nitrate from solution, most likely due to denitrification. Approximately 65% of the nitrate was removed over several days, suggesting denitrification rates between 2700 and 7000 $\mu$moles $m^{-3}$ $hr^{-1}$. Collectively, our results from this site suggest that much of the nutrients injected into the subsurface are removed from solution and may not have a significant impact on surface waters. However, these experiments were conducted at a relatively small facility, while some facilities in the Keys inject as much as 750,000 L per day. Saturation of available adsorption sites and organic substrate availability may limit the efficiency of waste water nutrient removal under such conditions.

In Eve L. Kuniansky, editor, 2001, U.S. Geological Survey Karst Interest Group Proceedings, Water-Resources Investigations Report 01-4011, p. 177-178

To evaluate the fate of waste waster at a high volume injection well facility, another dual tracer study using $^{32}PO_4$ and $SF_6$ was conducted. During this study, rapid conduit flow as high as 7.9 m/day was observed. At this site, waste water rises rapidly after injection to 18-27 m due to the buoyancy of the low salinity waste water plume. It buoys upward until it meets an impermeable mud layer at about 5 m that overlies the KLL. The majority of the plume is then advected to the east due to the local hydraulic gradient that exists across this site. Initially, phosphate was rapidly adsorbed as observed at the Long Key site. After approximately 36 hours, however we began to see radio labeled $PO_4$ returning to solution, indicating that phosphate is being desorbed from the KLL and slowly returning to solution. The KLL underlying this site seems to be acting as a phosphate buffered system. As a result, the $PO_4$ concentrations 15 m from the injection well seem to be maintained at approximately 25 μM. This is supported by column experiments (Elliot, 1999), which also show that adding low phosphate water to KLL that is saturated with PO4 will cause desorption of $PO_4$ until an equilibrium value of approximately 25 μM is reached.

Denitrification assays (Acetylene-block technique) suggest rates of denitrification as high as 3000 μmoles $m^{-3}$ $hr^{-1}$, comparable to estimates from the Long Key study. Seagrass and macroalgae samples collected from surface waters around the site indicate that isotopically heavy nitrogen, which is indicative of sewage derived nitrogen (McClelland et al., 1997), is being incorporated into the biomass of these primary producers. Macroalgae collected in a canal to the east of the disposal well showed a del $^{15}N$ value of +13.55 per mil. Stable nitrogen isotopic samples of subsurface $NO_3$ as well as $N_2$/Ar samples have been collected and should provide a clearer picture of the fate of $NO_3$ in the subsurface at this site.

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# In-Cave Dye Tracing and Drainage Basin Divides in the Mammoth Cave Karst Aquifer, Kentucky


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## Abstract

Karst ground-water basin divides are generally depicted as two-dimensional lines on maps, but they are better considered as three-dimensional surfaces within the subsurface. Dye traces are necessary to map out these surfaces and to locate conduits inaccessible to cave surveyors, and are indispensable for understanding the geometry of the complex networks of flow paths through the aquifer.

A key reason why the Mammoth Cave System is the world's longest known cave is that its passages extend over several major ground-water basins. The divides between these basins define the drainage system geometry and precise location of them is critical for understanding and protecting the cave and its remarkable aquatic ecosystem. In 1999 we initiated a long-term program of dye tracing within the Mammoth Cave System to more precisely locate the divides and to understand their increasingly apparent complexities. In this paper we report on results of some of the more interesting dye injections of the program. Although the Mammoth Cave Karst Aquifer is perhaps the best understood conduit flow network in the world (with over 700 traces), we have found that much more work is needed to provide the level of understanding necessary for protection and conservation. The first reason is a matter of scale and resolution: with the current distribution of traces that define the ground-water basins, many regional basin divides are only approximately defined. In areas where this condition exists in combination with potential threats to ground-water quality (primarily urban and transportation areas) additional tracing is needed to know the flow paths of individual recharge points. A second reason for additional traces is to increase our understanding of the plumbing of active conduits through the karst aquifer. While this type of dye tracing is logistically demanding, requiring visits to in-cave dye recovery locations, it is adding a new level of detail to our understanding of the nature of the karst aquifer.


## Introduction

Since the first dye-trace was performed at Mammoth Cave in 1925 – Anderson, working for Louisville Gas and Electric, conducted several traces, including Three Springs to Echo River Spring, to demonstrate that ground water can travel through adjacent ridges which would pose engineering problems for a proposed high-dam on the Green River near Pike Spring (Brown, 1966) – dye-tracing has been used as the primary investigative tool for understanding, and thus, protecting the karst aquifer. A major effort was undertaken in the mid 1970’s through the early 1980’s (Quinlan and Ray 1981, 1989), which resulted in the first major survey of the ground-water Basins in the Mammoth Cave region. Among other important conclusions, their work showed that approximately 60% of the aquifer’s recharge area extends beyond the boundary of Mammoth Cave National Park, and was thus impacted by agricultural land use. During the early 1990’s Meiman and Ryan (1992) concentrated dye tracing efforts to the north side of the Green River where 16 ground-water basins were defined, again with many basins recharged by private lands. There have been many “specialized” traces over the past ten years, including a series of quantitative traces by Ryan (1992), flow-velocity traces by Murphy (1992), flood pulse tagging Hall (1994) and Ryan and Meiman (1996), and traces along Interstate 65 by Meiman and Capps (1995). It cannot be overstated that without these works we would know very little about the overall hydrogeology of the Mammoth Cave karst aquifer.

Within the past year and a half there has been a renewed interest in dye tracing at Mammoth Cave, with more than 20 dye injections since late May of 1999. While at first glance it might appear that the ground-water basin definition gained by these earlier studies


In Eve L. Kuniansky, editor, 2001, U.S. Geological Survey Karst Interest Group Proceedings, Water-Resources Investigations Report 01-4011, p. 179-185

has provided all the information needed to fully delineate karst ground-water basins in and around Mammoth Cave National Park, we have found it necessary to further these efforts with additional tracing.

There are two main reasons, in this arguably most intensively investigated karst aquifer in the world, that additional tracing is needed. The foremost reason is a matter of scale and resolution; as the areal distribution of traces, which define the ground-water basins, is quite lacking in detail proximal to many basin divides – of special concern as 60% of the national park's karst watershed is located on private lands (Figure 1). In some cases there are several kilometers between injection sites with basin divides drawn somewhere in between, and in large part based upon ground-water potentiometric data. In areas where this condition exists in combination of potential threats to ground-water quality (primarily urban and transportation areas) additional tracing is needed to know the flow paths of individual recharge points. This situation is best represented in the Cave City area where the boundary between the Turnhole Spring ground-water basin (which drains northwestward into the national park) and the Gorin Mill ground-water basin (which drains to the east through Hidden River Cave). This three-kilometer segment of boundary, through the continuously developing urban sections of Cave City, is defined by only three dye traces.

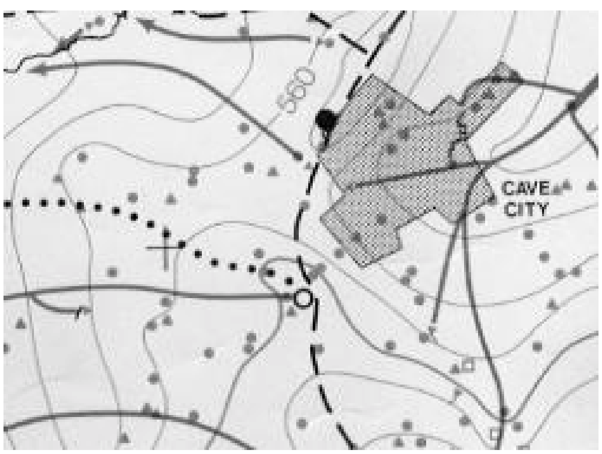

Figure 1. A portion of the 1:138,000 scale map "Groundwater Basins in the Mammoth Cave Region, Kentucky", Quinlan and Ray (1989). Ground-water basin and sub-basin divides are shown as dashed and dotted lines, respectively. Approximately 10 km$^2$ are shown in this view.

A second reason for additional tracing is to increase our understanding of the plumbing of active conduits through the karst aquifer. For the most part past tracing efforts have focused on surface injection points and spring recovery points, defining regional boundaries in a basic two-dimensional projection of the aquifer. We know, for example, that if a dye is introduced into a sinkpoint on the surface, and emerges at a spring, that the dye course has traversed active flow cave conduits. It would be very helpful, in respect to examining cave evolution and development, if a more detailed picture of the flow routes were described. As one could imagine this type of dye tracing is logistically demanding (requires visits to cave dye recovery locations), but it can be very rewarding. This in-cave tracing is the main focus of this paper.

Upon reading this paper you may notice the ubiquitous use of place-names. While the reader may not be familiar with these locations it would not be possible to write this paper without them. We will attempt to briefly describe these areas when necessary.

## Turnhole Spring Ground-water Basin

The 244 km$^2$ ground-water basin, by far the largest contributing flow into the park, has been the focus of study and water quality initiatives over the past 25 years. It is divided into the Mill Hole, Proctor, Cave City and Patoka Creek sub-basins. Each sub-basin is drained by a main trunk conduit: the Mill Hole stream, Hawkins River (Red River in Whigpistle Cave) Logsdon River, and Hawkins River, respectively. The vast majority of the sub-basins are under-represented by accessible cave streams. We know that there is active cave passages within these areas, but, for the most part, cannot gain access – usually due to passage collapse or sumps.

## Denial River

The portion of the Cave City sub-basin adjacent to the boundary of the Pike Spring basin within Roppel Cave presents an atypical situation, as many active flow passages are present. Near the upstream termination of Logsdon River at the S-188 Sump lies a complex tangle of small stream passages, one of which is called Denial River. This upstream portion of Logsdon River is also geologically interesting as Logsdon is apparently perched atop the Corydon Chert for much of its eight-kilometer traverse downstream to Pete Strange Falls, and then flows to Turnhole Spring on the Green River. There have been numerous air-filled stream passages found at elevations above and *below* Logsdon River in this area. Denial River was traced in order to begin unraveling this picture. Previous traces show that lower streams in this area, the flow from the Western Kentucky Parkway passage for example, drains northwards to Pike Spring. On October 19, 1999 approximately 0.5 kg of powder eoscien dye was injected into Denial River, which receives flow from a series of vertical shafts. Dye was recovered within two weeks at Pete Strange Falls and not at Pike Spring, showing that Denial River is a tributary of Logsdon River and thus within the Turnhole Spring Basin. This was the first in a series of traces from this area of the cave and many more traces will be necessary to fully understand its complex plumbing.

## Turnhole-Roaring/Echo Overflow

During the flurry of dye tracing in the late 1970's and early 1980's, Quinlan and Ray (1981 and 1989) discovered the existence of a high-level overflow route linking the main trunk of Turnhole Spring's Proctor sub-basin to the Echo River Spring ground-water basin. This overflow junction is inaccessible due to sumps in all river passages leading to it. Meiman and Ryan (1993) injected dye into Logsdon River during the recession limbs of various floods and demonstrated that this overflow route – which allows water from the Turnhole Basin to spill over into the adjacent Echo River Basin– is active when the stage of Logsdon River near its confluence with Hawkins River exceeds three meters over base flow stage (Figure 2). This was important information as the Echo River ground-water basin, which is almost exclusively within park boundaries during low flow times and harbors the endangered Kentucky Cave Shrimp, receives flow from nearly 100 $km^2$ of lands beyond the park boundary when Logsdon River stage is greater than three meters. Stage excesses of three meters is quite common. A reconstruction of float level records between 1984 and 1988 shows that this horizon was exceeded 38 times; periods when the aquatic ecosystem of Roaring and Echo Rivers were recharged largely by non-park lands.


Figure 2. Simplified diagram of the hydrogeology of the Turnhole Spring basin with the inter-basinal overflow to Echo River Spring.

Some discoveries are unintentional. On April 24, 2000, in support of a pesticide flood pulse-sampling effort, approximately two kg of powder fluorescein dye were injected into the Big Clifty/Girkin contact swallet at Cinnamon Falls – the Big Clifty is a sandstone unit overlying the 100 meter-thick carbonate sequence, with the Girkin being the uppermost limestone formation . Under almost any other circumstances investigators would not use such a copious quantity of dye. However, this trace, injected on the leading edge of a forecasted large rainfall event, was designed to place a sufficient amount of dye to "tag" a flood pulse and to detect in grab samples some five kilometers downstream in Logsdon River. Rainfall was intense during the injection, however within minutes after the dye was underground the rain event stopped. It did not rain again for weeks. This left 2 kg of fluorescein slowly coursing through base-flow cave streams. The flood pulse sampling operation was aborted. Two weeks later, while collecting dye receptors along the Green River, the Echo River Spring site exhibited a very strong presence of fluorescein. The highest stage in Logsdon River since the injection was approximately 0.5 meters above base level. A subsequent dye receptor recovery within Roaring River (an upstream tributary to Echo River Spring within the cave) also showed a strong positive. Dye (and thus water) is leaking into Roaring/Echo Rivers from Logsdon River during base flow. We believe that this flow route was not observed by previous investigators because, until now, no one had made the mistake of using far too much dye during base flow conditions, as “normal” quantities of dye would dilute into Roaring/Echo Rivers in concentrations below detection limits.

What does this mean? One possible explanation, relative to speleogenesis, is that we are witnessing the earliest stages of ground-water basin piracy. For many years Hawkins and Logsdon Rivers were the upstream portion of Roaring/Echo Rivers. At some point – perhaps due to a retreating nick-point in the Green River passing the Turnhole Bend cause by a increased incision rate of the Green – a more efficient route was established pirating flow from Hawkins River to Turnhole Bend Spring. Over time, as this piracy became the dominant flow route, the original route into Roaring/Echo was only used during high-flow periods as an overflow. It may be possible that this low-level “leak” marks the beginning of Roaring/Echo taking its flow back. Another explanation may be that the present overflow route, although a large, higher conduit, is simply wrought with many small “leaks” at lower elevations.

It is also possible for dye introduced into Logsdon River to find a secondary route before it even reaches the confluence of Hawkins River and the downstream sump and overflow area (one possibility is that the dye may enter Roaring/Echo via is upstream tributary Mystic River, however that idea was discounted as the above-mentioned dye receptor, which is far upstream of the confluence of Mystic, was positive). It remains possible, however, that the dye is taking an unknown route to the upstream segment of Roaring River. Additional simultaneous traces at Cinnamon Falls and at the Hawkins Sump under similar flow conditions are planned.

Aside from a hydrogeologic curiosity, what relevance does this have to cave management and protection? The discovery of the overflow route brought about major changes to conservation efforts of park managers, as it was proven that the sewer-less community of Park City flowed directly into endangered species habitat of the main active flow trunk of Mammoth Cave. This discovery was extremely important to the realization of a regional sewer system. Until now, it was thought at least Roaring/Echo River would be spared contamination events during low flow, but apparently not.

## Crystal Cave

Because Cooper and Collins Springs (both Haney Limestone springs perched upon the Big Clifty Sandstone) are well within the delineated bounds of the Pike Spring ground-water basin they, have apparently never been traced. With the assistance of Art and Peg Palmer, we placed dye receptors at six locations within the Crystal Cave section of the Mammoth System and introduced dye into the Big Clifty/Girkin contact swallets below Cooper and Collins Springs. Both dyes appeared at Pike Spring as suspected. Although the dye from Cooper Spring was not recovered within any of the monitored springs within Crystal Cave, the trace from Collins Spring (located 900 m north of Cooper) was detected in a strong positive at the C-3 Waterfall Campsite within the cave. We have not given up on finding and in-cave recovery point for Cooper Spring. This fall we re-traced the Cooper swallet after placing receptors in three locations in Sides Cave (nearby but not directly connected to the Mammoth System by exploration), which terminates near Crystal Cave and will retrieve them when weather permits.

## Outward Bound

In support of a Western Kentucky University Hoffman Environmental Research Institute study of land use delineation for the U.S. Fish and Wildlife Service a trace was initiated from the eastern-most known stream in the Roppel Cave section of Mammoth, therefore, the eastern most stream in the Mammoth System. It was possible that this stream recharged the proximal Suds Spring along the Green River. If so, this would mean that the Mammoth Cave System spans seven major ground-water basins and sub-basins – perhaps adding additional relevance and protection to the Suds Spring basin. The Outward Bound trace flowed to Pike Spring, and thus did not end up adding to the known recharge area of the cave system. Nonetheless, this added new hydrogeologic information.

## Three Springs

As mentioned in the introduction, the first dye trace in the Mammoth Cave area was performed at Three Springs. Quinlan and Ray (1981, 1989) traced this perched Haney Spring to Styx Spring. Meiman and Ryan (1993) found both Echo River Spring and River Styx Spring to receive flow from Three Springs. During the late fall of 1999 dye was again introduced into the Big Clifty/Girkin contact swallet of Three Springs, except this time, in addition to placing receptors at Echo and Styx Springs, we also monitored eight locations along River Styx within the cave. All receptors were positive – from the First Arch to the Dead Sea in the Cave, as well as Echo and Styx Springs. Although the Styx/Echo conduits were not experiencing a flow reversal event (water from the Green River entering Styx Spring, running through the cave and exiting at Echo River Spring), the system was not quite at base-flow levels. The flow conduit from Three Springs may enter the Echo/Styx Rivers upstream from the flow bifurcation of Echo River (see the next trace description to further examine the relationship between Echo and Styx). Additional tracing is needed, with receptors placed in Roaring and Mystic Rivers to pin-down its location.

An additional curiosity was found during autumn 1999 when Park Ecologist Rick Olson removed a visitor-constructed rock dam at Styx Spring. The removal of this dam, which increased the pool elevation of Styx Spring by several centimeters, did not register with the stage monitoring probe in the Dead Sea (which can resolve stage changes of 0.01 mm). This may indicate that there exists a free-surface segment of River Styx between its sump, as the change in stage was not propagated to the Dead Sea from Styx Spring.

## Service Station

All parking lots (and associated contaminants) within Mammoth Cave National Park drain into the cave system. The lot surrounding the service station near the main campground is no exception. During the fall of 1999 we injected dye into the Big Clifty/Girkin contact swallet behind the service station after placing receptors throughout the River Styx area of the cave. It should be noted that five endangered Mammoth Cave Shrimp were observed while placing receptors in the Hades section of River Styx. The highest elevation that dye was recovered was Shaller's Brook (a shaft drain off of Gratz Avenue). From there, after dropping through Lee's Cistern, the dye appeared at Hades, River Styx, the Dead Sea, and Styx and Echo River Springs. The park will soon begin construction of oil, grit, and metal removal filters on all major parking lots within the park.

The fact that dye which entered River Styx via the Hades stream – which is at the downstream end of Styx near the Dead Sea and well downstream of the Echo-Styx split – was also detected at Echo River is perplexing. Recall that the trace from Three Springs also appeared in Styx and Echo Springs, although that injection site was well distal and upstream of the Echo-Styx bifurcation. The service station trace was also conducted under "normal" flow conditions, that is to say, low-flow with no flow-reversal. Is it possible that a second conduit exists between River Styx and Echo River downstream of the River Styx sump? If not, was there a flow bifurcation within the vertical shafts leading down into the base-level, shunting flow to both Echo and Styx Springs? A rather simple trace by dumping dye into the Styx sump was preformed in November of 2000 and dye was only detected at Styx Spring, leading us to believe that the bifurcation is somewhere in the vertical shafts above the base-level streams upstream from Echo and Styx springs.

## Floating Mill Hollow

Upon close inspection of the map "Groundwater Basins of the Mammoth Cave Region, Kentucky" (Quinlan and Ray, 1989) one might find a few apparently minor springs along the Green River for which their recharge area is unknown. There are other small springs that do not appear on this or any other map. One such spring was found near the mouth of Floating Mill Hollow in 1999. Most of the time this spring does not appear to flow, because like other small springs, base-flow is discharged into the riverbed through alluvial filled under-flow spring. Basically, the surface springs orifice, where the dye receptors were placed, acts as an overflow discharge.

In the late spring of 1999 we placed receptors in all known springs from the upstream park boundary to Echo River Spring in preparation for a trace from Candlelight River, a small shaft drain stream along the famous Flint Ridge-Mammoth Cave connection route of 1972. According to the map of Quinlan and Ray (1989), Candlelight River should drain to Pike Spring. A few "old-timers" also had opinions on the spring that receives water from Candlelight – either Pike or Echo River Spring. If nothing else, dye-tracing came be a humbling enterprise: the dye appeared at neither Pike nor Echo, rather the newly discovered Floating Mill Hollow Spring (Figure 3). A new watershed was defined.

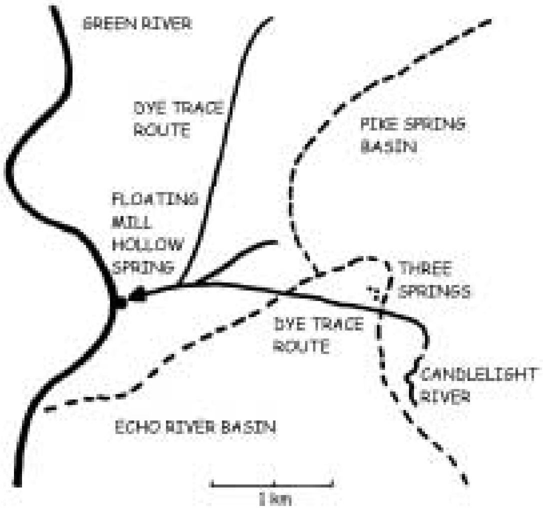


Figure 3. Dye traces to Floating Mill Hollow Spring. Dashed black lines, Quinlan and Ray (1989) ground-water basin boundaries, Quinlan and Ray (1989), solid lines, hypothesized flow routes from recent traces.

The fact that Candlelight River, which was originally placed within the Pike Spring basin, drains to Floating Mill Hollow Spring is of particular interest as in order to reach Floating Mill Hollow Spring it must either pass over, under, or around the Three Springs portion of the Echo River basin. We can hypothesize three possible scenarios. It seems quite possible that the flow route from Candlelight River actually passes over the conduit draining Three Springs. As far as we know the shaft complexes beneath such large Haney springs such as Three Springs have not been discovered – if a cave explorer is interested in finding the largest vertical shafts in the Mammoth Cave System we have some ideas on where to look. The fact that they exist should not be an issue, as even non-descript sink-points can produce large shafts (Bottomless Pit is a good example). Given that a shaft exists beneath Three Springs, it is very likely that it is very large and deep, as it receives perennial flow from Three Springs in addition to runoff from rainfall events. It is possible that such a shaft cuts deep to base-level, ignoring lithologic changes. Conversely, a small shaft drain as Candlelight River is perched upon local resistant lithologies. This perching might cause such a small stream to run horizontally for several hundred meters before it falls towards base-level (Figure 4).

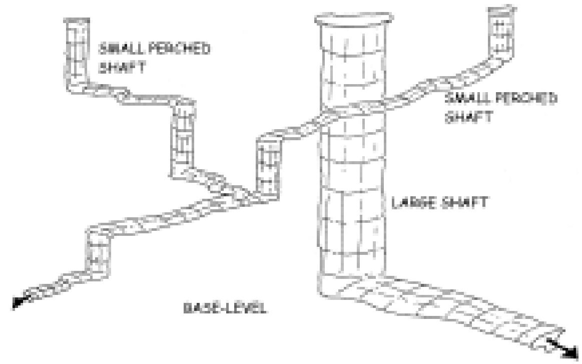


Figure 4. A simplified sketch of the complex three-dimensional nature of over-lapping karst ground-water basin divides – a possible explanation for the Candlelight River/Three Springs conundrum.

The discovery of a new ground-water basin led to two additional traces. Receptors were placed at Bögli Shafts and the river springs for a trace originating from the Big Clifty/Girkin contact swallet above the shafts. This trace was also recovered at Floating Mill Hollow Spring, however the receptors at Bögli Shafts were negative. It is very possible that individual shafts have such specific inputs that it is difficult to always make a connection.

A third trace was recovered at Floating Mill Hollow Spring from at trace initiated from the Big Clifty/Girkin contact swallet of Holton Hollow (the sinkpoint approximately 2.25 km north-northeast of the spring). Additional traces are planned for Rigdon, Taylor Coats, and an unnamed hollow, all which lie between Holton and Floating Mill Hollows.

## Conclusions

The past year and a half has been used as a trial to gauge the success of in-cave dye tracing. As predicted, such work is strenuous but provides important detail to interpret the hydrogeology of the cave and aquifer. We have learned a great deal already. Over the course of the next few years we plan many additional traces along these same lines. In addition to the future traces described above, we also plan on focusing on refining ground-water basin boundaries on the Pennyroyal Plateau, especially within developed portions of the watershed.

Another aspect of karst hydrogeology is becoming readily apparent from detailed in-cave dye tracing, especially proximal to basin boundaries: the complex three-dimensional nature of basin divides. This actually comes as no surprise as mature cave systems, such as the Mammoth Cave System, display intricate three-dimensional relationships between cave conduits, and therefore, ground-water basin boundaries are governed by the detailed geometry of flow routes.

## Acknowledgements

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# Delineation of Recharge Areas for Karst Springs in Logan Canyon, Bear River Range, Northern Utah

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## Abstract

Fluorescent-dye tracing was used to determine recharge areas, general directions of ground-water flow, and residence times for water from four karst springs in the Logan Canyon area of the Bear River Range in northern Utah. Dewitt, Wood Camp Hollow, Logan Cave, and Ricks Springs discharge from Paleozoic-age carbonate rocks into the Logan River, which is base level for ground water that discharges from this alpine region.

Recharge to the carbonate aquifer occurs through point sources, as seepage losses through fluvioglacial deposits, and as diffuse infiltration. On the basis of dye tracing to date (1999), 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 ground-water 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 ground-water basins. Ground-water movement in a large part of the area is influenced by the Logan Peak syncline.

Discharge of springs ranges from less than 1 to about 75 cubic feet per second. Spring discharge responds primarily to snowmelt runoff, with peak flow from late spring to early summer and base flow during the winter months. Specific conductance of spring water from May 1994 to July 1997 ranged from 250 to 420 microsiemens per centimeter at 25 degrees Celsius and water temperature ranged from 5.5 to 8.0 degrees Celsius. For all springs, specific conductance and temperature were inversely related to discharge. Observed differences between measured values result primarily from mixing of recharge derived from snowmelt with ground water.

## INTRODUCTION

The Bear River Range in northern Utah and southern Idaho is part of the Middle Rocky Mountains Physiographic Province (Stokes, 1988). The range is part of a thrust sheet (allochthon) that was emplaced eastward by a deeply buried thrust fault during Cretaceous time (Dover, 1987). In northern Utah, the range is bisected by several east to west rivers, including the Logan River (fig. 1). In the Logan Canyon area, east of Logan, Utah, altitude ranges from about 5,000 feet (ft) along the river to almost 10,000 ft on higher peaks. Mean annual precipitation at Tony Grove Lake (8,000 ft) for 1979-99 was 49.8 inches (Natural Resources Conservation Service SNOTEL data, http://utdmp.utsnow.nrcs.usda.gov). Most precipitation occurs as snow from October to March.

Karst features in the Logan Canyon area are indicative of a hydrologic system that is developed within more than 3,000 ft of Paleozoic limestone and dolomite. Karst features in this alpine region 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 (Wilson, 1976). 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. To better characterize the hydrologic system in this alpine karst, an investigation was begun to (1) determine variations in discharge of selected large springs; (2) correlate temperature and specific conductance of spring water with changes in discharge; (3) determine recharge areas for the springs and general directions of ground-water flow; (4) delineate ground-water basin divides; (5) determine ground-water travel times; and (6) evaluate the effects of geology on ground-water movement.

In Eve L. Kuniansky, editor, 2001, U.S. Geological Survey Karst Interest Group Proceedings, Water-Resources Investigations Report 01-4011, pp. 186-193.

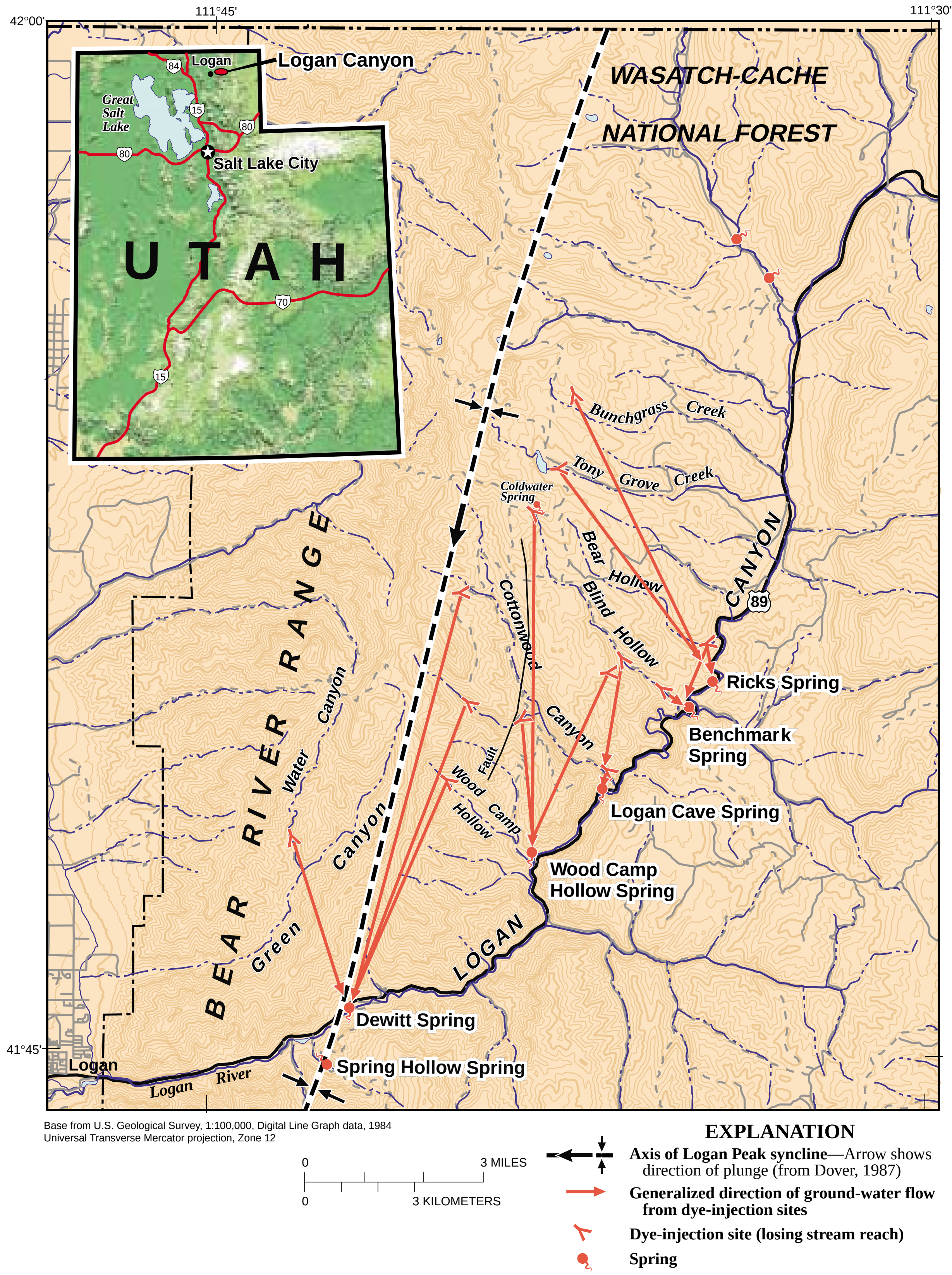


**Figure 1.** Generalized ground-water flow paths to selected springs, on the basis of results of dye tracing, Logan Canyon, Utah.

In Eve L. Kuniansky, editor, 2001, U.S. Geological Survey Karst Interest Group Proceedings, Water-Resources Investigations Report 01-4011, pp. 186-193.

## METHODS

Dye-tracing methods commonly are used to determine ground-water flow paths, relations between surface-water and ground-water basins, and ground-water travel times in karst aquifers (Mull and others, 1988). Dye tracers have been successfully used in karst and other high-permeability terrains where other types of tracers have limited use. Sodium fluorescein dye was selected as the principal tracer for this investigation. Characteristics of the dye include detectability at low concentrations and over long distances, relatively low sorption tendencies, good solubility in cold water, low cost, and an affinity for activated charcoal. Dye tracing was done during low to moderate spring flows and dye amounts were based on discharge and distance from the springs. In place of automatic water samplers, passive (cumulative) dye detection with activated charcoal was used to recover the dye for analysis. Dye packets were exchanged every 1 to 4 weeks; consequently, calculated ground-water travel times are considered maximums and in most cases, are probably substantially less. Procedures for recovery of dye from activated charcoal are outlined in Mull and others (1988).

Discharge from most springs in the study area is not gaged. Spring discharge was measured periodically at different flow rates with a pygmy current meter, and intervening flow rates were estimated. Small spring flows were measured with a modified Parshall flume. Discharge at peak flows is estimated to be within plus or minus 20 percent of actual values. Discharge data for Dewitt Spring were obtained from the City of Logan Water Department (Dennis Corbridge, written commun., 1998). Specific conductance was measured with a Beckman conductivity meter that was periodically calibrated to known standards. Water temperature was measured to the nearest one-half degree with a mercury thermometer.

## HYDROGEOLOGY

The Bear River Range consists in large part, of a thick sequence of carbonate (limestone and dolomite) rocks that range in age from Cambrian to Mississippian (Dover, 1987). The principal geologic units in this area and approximate thicknesses are the Garden City Formation (1,400 to 2,000 ft), Swan Peak Quartzite (200 to 400 ft), and Fish Haven Dolomite (350 ft) of Ordovician age; the Laketown Dolomite (1,500 to 2,000 ft) of Silurian age; the Water Canyon Formation (425 to 600 ft), Hyrum Dolomite (850 ft), and Beirdneau Formation (1,000 ft) of Devonian age; and the Lodgepole Limestone (750 ft) of Mississippian age. Karst is more developed in the Garden City Formation and Laketown Dolomite than in the other carbonate units. All of the units, however, are capable of transmitting water along dissolution-enhanced fractures, faults, and bedding planes. The Swan Peak Quartzite is probably a barrier to downward movement of water from the Fish Haven Dolomite to the Garden City Formation in some areas and likely influences the direction of ground-water movement. All of the formations make up the upper part of a large regional structure, the Logan Peak syncline (Williams, 1948) (fig. 1). The syncline plunges to the southwest at about 15 degrees and rocks on the west limb dip at a considerably steeper angle than those on the east limb. This structural feature and associated fractures influence the movement of ground water in much of the region.

### Aquifer Recharge

Recharge to the carbonate aquifer takes place through point sources (sinkholes and pits), as seepage losses through fluvioglacial deposits that fill valley drainages, and as infiltration along ridges and valley slopes. Sinkholes (dolines) and pits are typically developed in high-altitude meadows where snow accumulates and may persist throughout much of the year. Water entering point sources moves vertically downward along solution-enlarged fractures to principal conduits that channel water to the springs. Pits range in depth from less than 100 to as much as 300 ft, but many of these have been occluded by fluvioglacial materials consisting primarily of quartzite boulders. Fluvioglacial deposits also form a veneer over carbonate bedrock in valley drainages. These deposits are very permeable and streams typically sink into the streambed along distances of several hundred yards rather than in distinct point sources such as swallow holes. These losing reaches are probably related to fracture zones within the underlying bedrock. Most streams in these alpine drainages are fed by snowmelt runoff and, therefore, tend to be seasonal. During periods of peak runoff, however, streamflow that is not lost to the underlying bedrock continues down surface-water courses to the Logan River. Infiltration of snowmelt along ridges and valley slopes provides an additional component of recharge to the aquifer and probably moves along diffuse pathways through the fractured-rock matrix. Diffuse flow can be a significant component of long-term storage in the aquifer and maintenance of base flow of springs.

### Discharge from Springs

Discharge from the carbonate aquifer is primarily from large springs along the Logan River. The Logan River is the principal base level stream for ground-water discharge in this part of the Bear River Range. 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 (fig. 1). These include Dewitt, Wood Camp Hollow, Logan Cave, and Ricks Springs. Only one large (second magnitude) spring is known to discharge along the south side of the river (fig. 1). Collective discharge of the springs provides a substantial component of streamflow in the Logan River. Wilson (1976) estimated that the combined flow of Wood Camp Hollow, Logan Cave, and Ricks Springs could be as much as 20 percent of the discharge of 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 (fig. 2).

Dewitt Spring discharges from the Water Canyon Formation along the flood plain of the Logan River (fig. 1). Water discharges from an unknown number of outlets that are capped and is diverted into a collection system. Water from the spring serves as a public supply for the city of Logan, about 7 miles (mi) to the west. Discharge of the spring is metered and generally ranges from about 10 to 35 $ft^3/s$ (Dennis Corbridge, City of Logan, written commun., 1998) (fig. 2).

Wood Camp Hollow Spring discharges from two adjacent outlets in the Laketown Dolomite. Discharge of the spring ranges from about 3 to at least 40 $ft^3/s$. Discharge at peak flow is difficult to determine because backwater from the Logan River partially impounds free flow from the spring. During base flow, discharge from both outlets appears to be similar. As discharge increases, however, substantially greater amounts of water discharge from the upstream outlet, and flow from the adjacent (downstream) outlet is estimated to be less than 10 $ft^3/s$.

Logan Cave Spring discharges from the base of the Garden City Formation. Water from the spring normally discharges from talus below the entrance to Logan Cave, but during spring runoff, also discharges directly from the cave entrance. Discharge of the spring normally ranges between 1 and 10 $ft^3/s$; Wilson (1976) estimated flows of as much as 25 $ft^3/s$ in June 1975. Peak flow of Logan Cave Spring generally occurs earlier than that of the other springs. Most of Logan Cave is developed along a master joint that trends due north, with dissolution along secondary northwest-trending joints in some areas.

Ricks Spring discharges from four outlets along several hundred yards of a normal fault in the Garden City Formation. Discharge of the spring ranges from less than 1 to a reported 75 $ft^3/s$ (Mundorff, 1971). During snowmelt runoff, most water discharges from a large alcove developed on the fault, and estimated combined flow from the three additional outlets is less than 5 $ft^3/s$. During winter, flow generally ceases from the alcove, and total base flow from the smaller springs, which are at a slightly lower altitude along the Logan River, can be less than 1 $ft^3/s$.

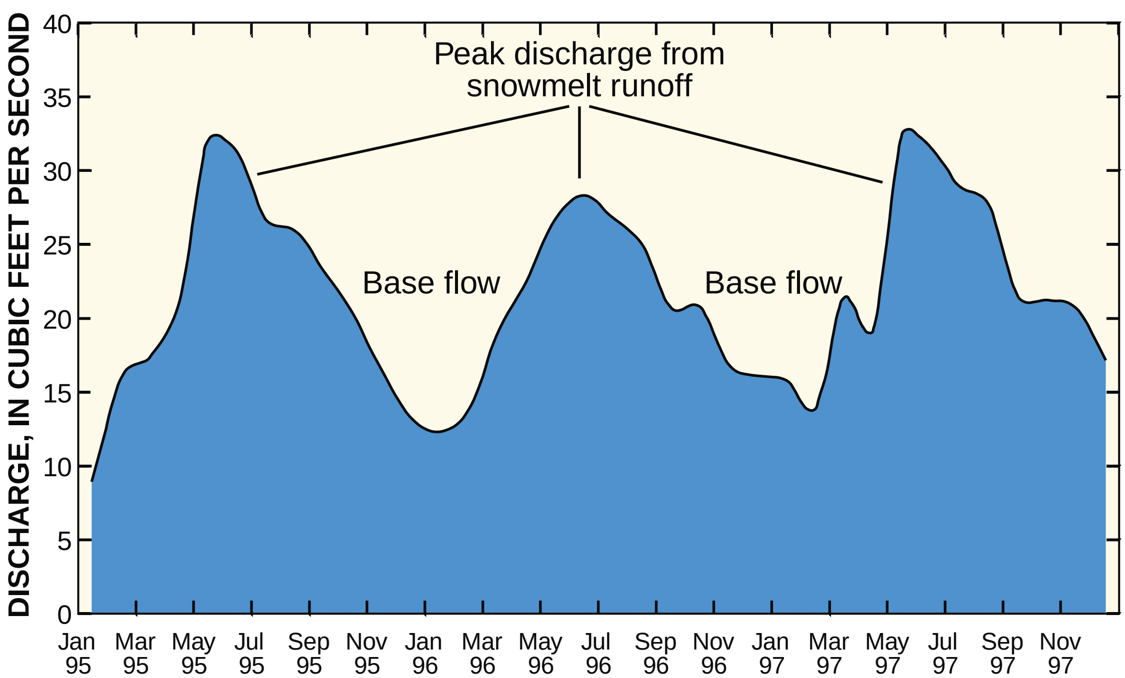


**Figure 2.** Hydrograph showing typical seasonal response of an alpine karst spring to snowmelt runoff, Dewitt Spring, Logan Canyon, Utah, January 1995 to January 1998 (Data from City of Logan Water Department, written commun., 1998).

## RESULTS OF DYE TRACING

Results of dye tracing indicate that water from losing streams moves downward through fluvioglacial deposits into underlying carbonate rocks to discharge areas along the Logan River. In some areas, water moves vertically downward through more than one carbonate unit before discharging. Dye tracing also indicates that surface-water drainage basins do not necessarily coincide with ground-water basins. Water from losing streams typically moves beneath ridges from one surface-water basin to discharge areas in adjacent surface-water basins. Furthermore, streams in the upstream and downstream parts of some drainages lose water to different ground-water basins. Streams in the Cottonwood Canyon surface-water basin lose water to three separate ground-water basins (fig. 1). Interbasinal ground-water movement between springs during periods of high flow has been documented in alpine karst areas of eastern Utah (Maxwell and others, 1971). Although this phenomenon has not been observed in this area on the basis of dye tracing at low to moderate flows, ground-water basin divides can be dynamic boundaries, particularly where these divides cross stream drainages.

Four dye traces to Dewitt Spring indicate a recharge area northwest to northeast of the spring that largely coincides with the areal extent of the Logan Peak syncline. Ground-water movement is probably downdip along the west and east limbs of the syncline toward the axis (fig. 1), 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 ground-water travel time of 22 days for losing streams in drainages 7.2 mi from, and 2,900 ft higher than, Dewitt Spring (table 1). The substantial base flow of this spring relative to peak flow may indicate a large storage component that is recharged primarily from infiltration. Ground-water travel times from this diffuse component of flow are likely to be considerably longer than those from losing streams. On the basis of dye tracing, geology, and discharge, the ground-water basin for Dewitt Spring is estimated to be about 15 square miles ($mi^2$).

Three dye traces to Wood Camp Hollow Spring indicate a recharge area generally north of the spring and as much as 3,200 ft higher than the spring. Results of dye tracing during moderate flow also indicate a maximum ground-water travel time of 28 days to the spring from as far as 5.6 mi (table 1). Ground-water movement within the Wood Camp Hollow Basin is probably along north-trending fractures and possibly faults (fig. 1). Southwest-dipping strata along the east limb of the Logan Peak syncline, however, may influence ground-water movement from the northeast. In-cave tracing from this area indicates largely vertical movement of water for almost 1,000 ft. Water then probably moves downdip to the southwest, possibly along the top of the Swan Peak Quartzite, to where it eventually merges with flow in the principal conduit. On the basis of dye tracing, discharge, and relations to adjacent ground-water basins, the recharge area for Wood Camp Hollow Spring may be as large as 10 $mi^2$.

Two dye traces to Logan Cave Spring indicate a recharge area generally north of the spring and at a lower altitude than that of the recharge areas for the other springs (fig. 1). Substantial contributions to discharge of this spring originate from the lower part of Cottonwood Canyon and the Blind Hollow drainage (fig.1). Movement of ground water to this spring is probably largely within the Garden City Formation. Dye tracing, discharge, and relations to adjacent ground-water basins indicate that the recharge area for Logan Cave Spring is probably less than 5 $mi^2$.

Three dye traces to Ricks Spring indicate a recharge area that extends more than 5 mi to the northwest and about 2,600 ft higher than the spring (fig. 1). Water that originates from the lower part of Bear Hollow also appears to move laterally along the fault on which the spring is located, to discharge at Benchmark Spring, about half a mile to the southwest (fig. 1). In addition, investigations by the U.S. Forest Service during the 1970s (District Ranger, Wasatch-Cache National Forest, oral commun., 1999) indicate that part of the flow to Ricks Spring originates directly from the Logan River. Losses from the Logan River are probably along an extension of the fault where it intersects the river upstream from the spring. Maximum travel time of ground water to the spring from 5.3 mi during low flow was 28 days (table 1). Ground-water movement in the Ricks Spring Basin appears to be generally along the east limb of the Logan Peak syncline and probably within the Garden City Formation. Northwest-trending high-angle faults are mapped northeast of the Ricks Spring ground-water basin, on the east side of the Logan River; however, younger deposits mantle the carbonate rocks in the basin and obscure structural relations. On the basis of dye tracing, the recharge area for Ricks Spring is estimated to be at least 7.5 $mi^2$.

**Table 1.** Summary of dye traces to selected karst springs in Logan Canyon, Utah

[—, no data; <, less than]

Maximum travel time: Travel time calculated from initial dye recovery on activated charcoal; actual travel time probably substantially less.
Linear distance: Straight-line distance between dye-input and dye-recovery sites; actual distance probably substantially greater.
Vertical distance: Difference between altitudes of dye-input and dye-recovery sites.
Spring discharge: Estimated/measured discharge of spring at time of dye injection.

| Dye-input site | Altitude (feet) | Date and time of dye injection | | Amount of fluorescein dye (pounds) | Dye-recovery site | Altitude (feet) | Date and time of dye recovery | | Maximum travel time (days) | Linear distance (miles) | Vertical distance (feet) | Spring discharge (cubic feet per second) |
|---|---|---|---|---|---|---|---|---|---|---|---|---|
| **Dewitt Spring Basin** | | | | | | | | | | | | |
| South Fork, Cottonwood Canyon | 7,160 | 09/15/1995 | 1700 | 2.4 | Dewitt Spring | 5,040 | 10/09/1995 | 1745 | 24.0 | 5.6 | 2,120 | 25 |
| Upper Wood Camp Hollow | 7,120 | 09/13/1996 | 1800 | 2.0 | Dewitt Spring | 5,040 | 10/14/1996 | 1415 | 30.8 | 4.3 | 2,080 | 23 |
| Upper Cottonwood Canyon | 7,920 | 07/05/1998 | 1800 | 4.4 | Dewitt Spring | 5,040 | 07/27/1998 | 1740 | 22.0 | 7.2 | 2,880 | 28 |
| Water Canyon | 6,320 | 11/11/1999 | 1500 | 2.0 | Dewitt Spring | 5,040 | 11/19/1999 | 1500 | 8.0 | 3.0 | 1,280 | 20 |
| **Wood Camp Hollow Basin** | | | | | | | | | | | | |
| South Fork, Cottonwood Canyon | 6,460 | 09/21/1991 | 1815 | 1.6 | Wood Camp Hollow Spring | 5,360 | 10/12/1991 | 1230 | 20.8 | 2.3 | 1,100 | 5 |
| Coldwater Spring | 8,520 | 09/11/1993 | 1100 | 3.4 | Wood Camp Hollow Spring | 5,360 | 10/09/1993 | 1640 | 28.2 | 5.6 | 3,160 | 12 |
| Nielsens Cave | [1]7,000 | 09/10/1994 | — | 1.0 | Wood Camp Hollow Spring | 5,360 | 09/30/1994 | 1505 | 20.0 | 3.3 | 1,640 | 11 |
| **Logan Cave Spring Basin** | | | | | | | | | | | | |
| Blind Hollow | 6,800 | 11/11/1993 | 1530 | 1.3 | Logan Cave Spring | 5,520 | 12/04/1993 | 1545 | 23.0 | 2.3 | 1,280 | 1 |
| Cottonwood Canyon | 5,640 | 10/14/1996 | 1800 | 1.0 | Logan Cave Spring | 5,520 | 11/11/1996 | 1345 | 27.8 | .30 | 120 | 1 |
| **Ricks Spring Basin** | | | | | | | | | | | | |
| Bear Hollow | 6,120 | 10/12/1991 | 1645 | .72 | Ricks Spring | 5,880 | 10/26/1991 | 1550 | 14.0 | .68 | 240 | 4.5 |
| Bear Hollow | 6,120 | 11/11/1992 | 1700 | 1.1 | Ricks Spring | 5,880 | 11/28/1992 | 1450 | 16.9 | .68 | 240 | 2 |
| | | | | | Benchmark Spring | 5,800 | 11/28/1992 | 1420 | 16.9 | 1.0 | 320 | <1 |
| Blind Hollow | 6,200 | 04/26/1992 | 1930 | .83 | Benchmark Spring | 5,800 | 05/16/1992 | 1915 | 20.0 | .80 | 400 | 1 |
| Tony Grove Creek | 7,980 | 06/07/1992 | 1130 | 3.4 | Ricks Spring | 5,880 | 07/02/1992 | 1940 | 25.3 | 4.3 | 2,100 | 10 |
| Bunchgrass Creek | 8,520 | 08/13/1994 | 1700 | 3.6 | Ricks Spring | 5,880 | 09/10/1994 | 1650 | 28.0 | 5.3 | 2,640 | 7.5 |

[1] Approximate altitude of dye-injection site inside cave.

## PHYSICAL PROPERTIES OF SPRING WATER

Water temperature and specific conductance were measured at Wood Camp Hollow, Logan Cave, and Ricks Springs from May 1994 to July 1997, along with estimates of discharge. For all springs, specific conductance and temperature of water were inversely related to discharge. During periods of runoff and peak flow (late spring to early summer), specific conductance of water from the springs ranged from 250 to 290 microsiemens per centimeter (µS/cm) at 25 degrees Celsius (°C) and temperature ranged from 5.5 to 6.0°C (fig. 3). Conversely, during periods of base flow (late fall to late winter), specific conductance of water ranged from 340 to 420 µS/cm and temperature ranged from 6.5 to 8.0°C (fig. 3). Observed differences between measured values result from mixing of recharge derived from snowmelt with ground water. Snowmelt (low dissolved-solids concentration) from point sources and losing streams moves rapidly through the aquifer along dissolution-enhanced flow paths (joints, faults, and bedding planes), mixing with ground water in storage (higher dissolved-solids concentration) that has had a longer residence time. McGreevy and Bjorklund (1970) reported specific-conductance values of water from Ricks Spring that averaged 341 µS/cm for periods of base flow between 1958 and 1962. This average value is virtually identical to values obtained during this investigation and indicates that the dissolved-solids concentration in water from this spring has probably remained essentially the same for at least 40 years.

Hydrogen-ion activity (pH) was not measured during this investigation. However, Wilson (1976) reported monthly pH values of 7.4 to 7.7 for Wood Camp Hollow, Logan Cave, and Ricks Springs. McGreevy and Bjorklund (1970) also reported pH values of 7.4 to 8.2 for Ricks Spring and values of 8.0 and 8.1 for Dewitt Spring. Reported values of pH and corresponding estimated discharge do not indicate a definitive relation between these parameters.

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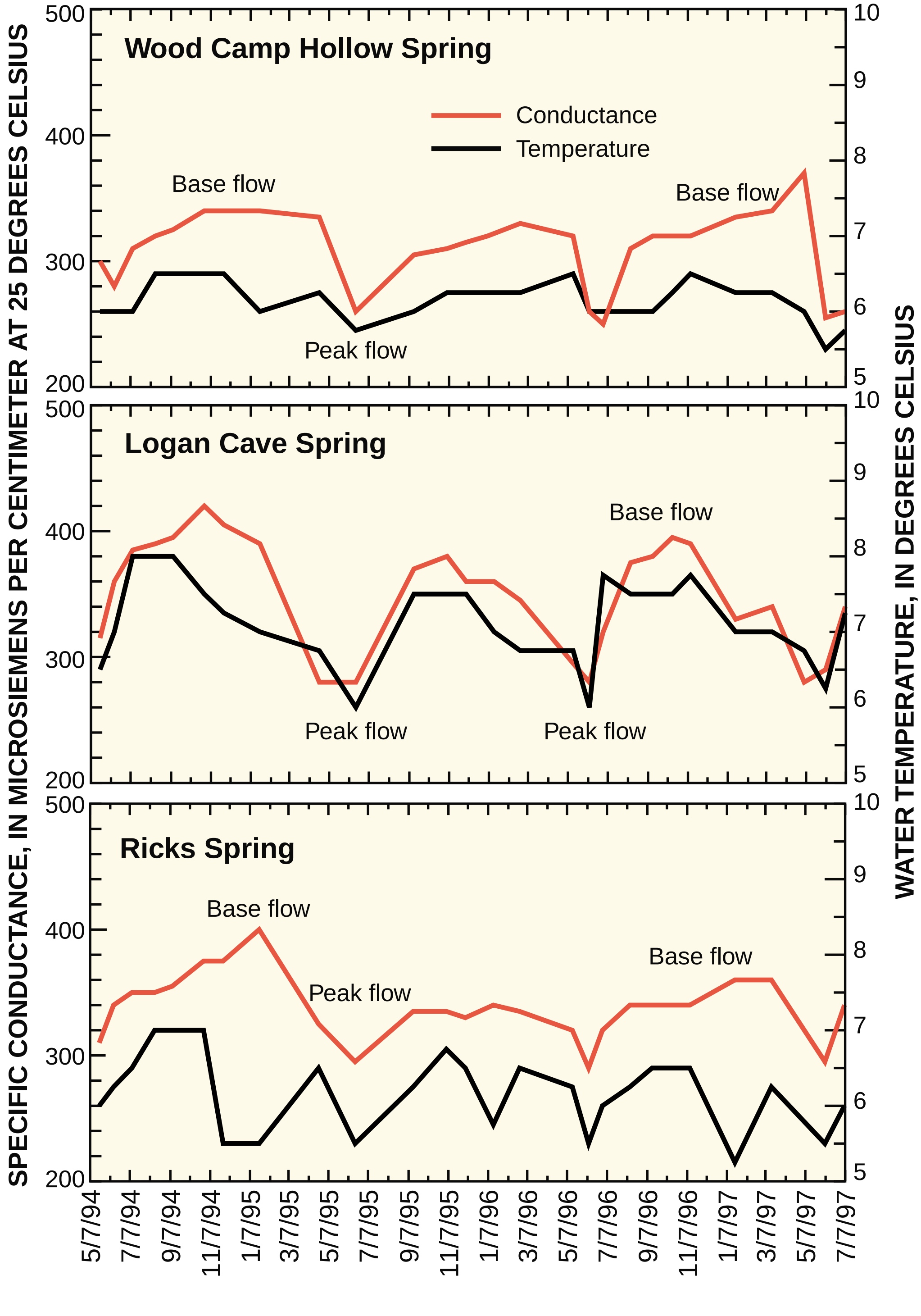


**Figure 3.** Relation of specific conductance and water temperature to peak and base flows of selected springs, Logan Canyon, Utah, May 1994 to July 1997.

# Submarine Ground-Water Discharge in Upper Indian River Lagoon, Florida

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## Abstract

The discharge of submarine groundwater has recently been shown to be an important process in many environmentally fragile coastal ecosystems. However, groundwater discharge into coastal bottom water is still an often-overlooked component of many hydrologic and oceanic models. The exchange of interstitial water across the sediment/water interface may introduce anthropogenic pollutants, may be an important part of coastal nutrient cycles, and may cause excess nutrient loading, thereby potentially degrading the coastal water quality. Here we report on Year-1 results from a co-operative (USGS-UF-LSU) project that is investigating the role of submarine groundwater discharge into Indian River Lagoon, Florida.

## INTRODUCTION

The Indian River Lagoon system extends over 250 km along the east-central coast of Florida and consists of three inter-connected lagoonal basins: Mosquito, Banana River, and Indian River lagoons. Exchange of lagoon water with the Atlantic Ocean is limited to four tidal inlets (Sebastian, Ft. Pierce, St. Lucie and Jupiter) that occur in the southern part of Indian River Lagoon. Precipitation, the exchange of water through these inlets, wind, tidal forcing, evaporation, surface runoff, and potential submarine groundwater discharge control the salinity of lagoon water. In this system, the intensity and duration of wind have the most pronounced affect on lagoon water levels. The overall objective of this project was to determine the rate and potential ecological significance of submarine groundwater discharge (Table 1) to Indian River Lagoon.

The study area during the first year of the project included the northern most 10 km of the Indian River Lagoon (~48 km$^2$). Of the 28 sampling stations, 22 were arranged in shore-perpendicular transects; the remaining six stations were distributed within the lagoon center (Fig. 1). At each station, lagoon and interstitial water samples were collected, and groundwater seepage rates were measured using conventional seepage meters. Interstitial water samples were obtained from four stations using custom-built multi-samplers. Six groundwater samples were collected from wells surrounding the lagoon. Two additional samples were collected from tributaries to the lagoon including Turnbull Creek and Haulover Canal. Sampling of the seepage stations, groundwater wells, and tributaries occurred in May 1999, to coincide with the end of the normal dry season, and in August 1999, during the normal rainy season. A third trip in December 1999 was used only to sample interstitial water.


Table 1. Factors that affect submarine groundwater discharge.

### Hydrogeology

The hydrogeology along the northeastern coast of Florida can be broadly divided into two aquifer systems: the Surficial and the Floridan aquifer system. Sand, silt and clays of the Intermediate confining unit, which constitutes most of the Hawthorn Formation, separates these two aquifer systems (Leve, 1970; Spechler, 1994). The Surficial

In Eve L. Kuniansky, editor, 2001, U.S. Geological Survey Karst Interest Group Proceedings, Water-Resources Investigations Report 01-4011, p. 194-197


Figure 1. Site location map for upper Indian River Lagoon, Florida.

aquifer system consists of Miocene to Holocene interbedded sand, shell, silt, clay and dolomitic limestone strata. The Surficial aquifer system is mostly unconfined, although the hydrogeology can be very heterogeneous. Four clastic, very regional surficial aquifers border the Indian River Lagoon including Terrace, Atlantic Coastal Ridge, Ten-mile Ridge, and Inter-ridge. Terrace aquifer occurs on the barrier islands separating Indian River Lagoon from the Atlantic Ocean. The Atlantic Coastal Ridge occurs on the western bank of Indian River Lagoon, in the northern reaches of the lagoon. This aquifer is composed of the Pleistocene Anastasia Formation, and provides most of the water supply for towns on the western edge of the northern Indian River Lagoon (Mims and Titusville). The Floridan aquifer system can be further divided into two water-bearing aquifers (Upper and Lower Floridan), separated by less permeable semi-confining units. The Upper Floridan aquifer in the study area corresponds to the Ocala Limestone and in some parts, the Avon Park Formation. The Ocala Limestone is characterized by high permeabilities that can be enhanced along bedding planes, fractures, and conduits.

Significant variations in ground-water levels occur seasonally (Fig. 2). Superimposed on such seasonal variations is a long-term decrease in the potentiometric surface that is largely attributed to increased groundwater withdrawals (Fig. 3). Nonetheless, recent potentiometric surface maps of the Upper Floridan aquifer indicate elevations that are above sea level for the entire length of Indian River Lagoon. Such potentiometric surface elevations increase from north to south, where the Hawthorn Formation increases in thickness. The elevated potentiometric surface of the Upper Floridan, combined with the general lack of a

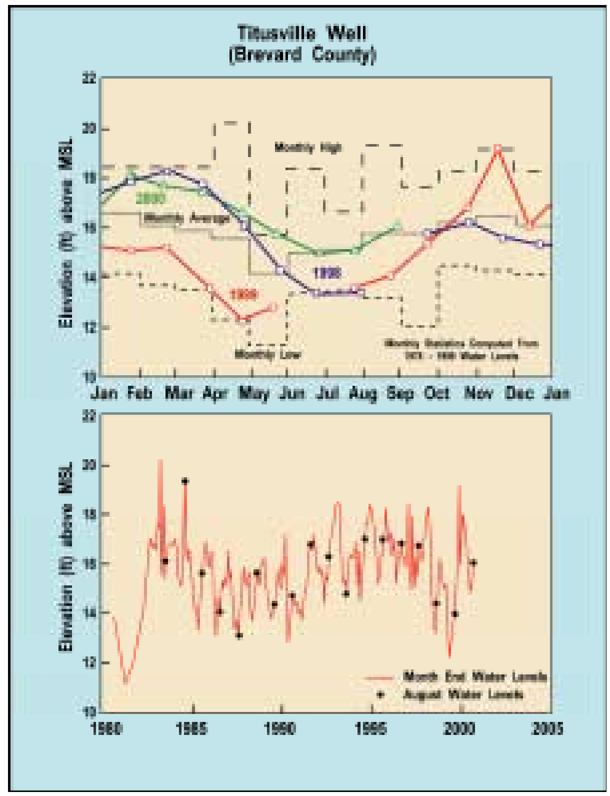


Figure 2. Hydrograph of a Titusville well (adapted from St. John's River Water Management District, 2000).

confining unit in the vicinity of the study area makes much of Indian River Lagoon a potential zone of submarine groundwater discharge.

## Geochemistry

To derive estimates of ground-water seepage into Indian River lagoon, the following suite of tracers, chemical constituents and sampling devices were measured or utilized: nutrients, $Cl^-$, conductivity, pH, temperature, dissolved oxygen, $^{87}Sr/^{86}Sr$, $\delta^{18}O$, $^{223,224,226}Ra$, $^{222}Rn$, seep meters, multi-samplers, and benthic flux chambers (Martin et al., 2000). Seepage rates were spatially and temporally heterogeneous,

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yet similar to rates previously measured in Indian River Lagoon using identical techniques. The seepage rates ranged from 3 - 100 ml $m^{-2}$ $min^{-1}$ during May (dry season) to 22 - 144 ml $m^{-2}$ $min^{-1}$ during August (rainy season). The average value for all meters increased from 40 to 63 ml $m^{-2}$ $min^{-1}$ from the dry to the rainy season, implying that there may be a connection between rainfall and increased seepage rates. The heterogeneous nature of these rates is likely caused by fluctuations in sediment permeabilities and other geologic characteristics.

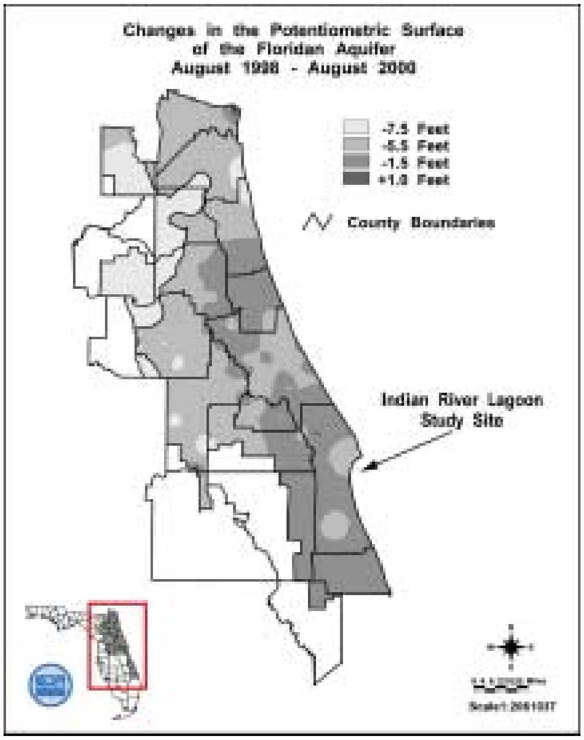


Figure 3. Potentiometric change map (adapted from St. John's River Water Management District, 2000).

## Geochemistry

Radon-222 and Ra isotopes have previously provided regionally integrated estimates of seepage flux in varied coastal environments (Cable et al., 1996; Moore, 1999; Swarzenski et al., in press). Benthic fluxes of Ra to the Indian River Lagoon are calculated using three independent methods that rely on the activities of short-lived Ra isotopes: 1) lagoon budget, 2) benthic flux chambers and 3) pore-water modeling (Fig. 4). The first two methods yield direct measurements of flux across the sediment/water interface, whereas the third technique generates an indirect flux estimate on the basis of pore-water Ra profiles. Calculations of the benthic flux of Ra range up to almost 500 dpm $m^{-2}$ $day^{-1}$. Using $^{226}Ra$ pore-water activities, a maximum upward subsurface water flow of about 5 - 17 cm $day^{-1}$ is required to sustain these fluxes. These values are similar to the values measured directly with the seepage meters.

By using $^{222}Rn$ and $^{226}Ra$ as mass balance tracers of seepage flux to the northern Indian River Lagoon it is possible to obtain measurements of seepage that are independent of the short-lived Ra isotopes. Assumptions required for this mass balance approach are that negligible effects were observed from surface water exchange to the lagoon, tides, and diffusion from the sediments. Analogous to the short-lived Ra isotopes, seepage fluxes measured on the basis of excess $^{226}Ra$ activities are similar in magnitude to those estimated using seepage meters. Each submarine groundwater discharge technique has individual strengths and weaknesses.

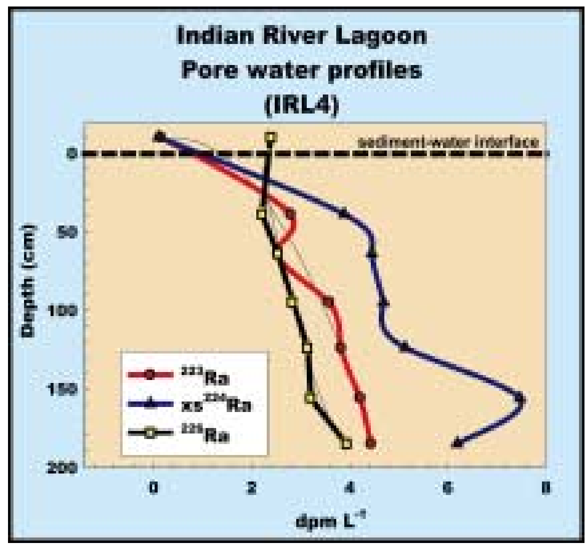


Figure 4. Interstitial radium activities at Station IRL 4, upper Indian River Lagoon.

Seepage meters provide a direct measurement of localized flow. They can also easily provide 'clean' seep water samples. However, seep meters may be susceptible to possible artifacts caused by interaction of tides and waves, although such limitations have not been thoroughly tested.

The radioisotopes are less difficult to sample in the field than using seepage meters, but their measurement requires sophisticated laboratory equipment that is not widely available. One important characteristic of the radioisotope techniques is that they provide an integrated value of

In Eve L. Kuniansky, editor, 2001, U.S. Geological Survey Karst Interest Group Proceedings, Water-Resources Investigations Report 01-4011, p. 194-197

seepage rates across the entire lagoon. They are thus complementary to the seepage meter technique.

Chloride concentrations indicate that only a minor component (1 - 5%) of seep water originates from meteoric groundwater. This implies that 95 - 99% of the interstitial water has to be recycled lagoon seawater. The isotopic concentration of strontium ($^{87}Sr/^{86}Sr$) was nearly identical in the seep water and lagoon water, yet was measurably lower than that in modern seawater. The $^{87}Sr/^{86}Sr$ ratios were also systematically lower during the rainy season, reflecting the greater influx of seep water into lagoon water and short groundwater residence times. Nutrient concentrations were 3 to 5 times elevated in the seep water over the lagoon water, and suggest that sediment/water interface exchange processes, such as submarine groundwater discharge, are critical components of coastal nutrient budgets (Johannes, 1980; Krest et al., 2000).

## SUMMARY

The hydrogeologic framework of northeastern Florida indicates that the Hawthorn confining unit is thinnest or absent in the vicinity of upper Indian River Lagoon. This feature might imply a vigorous hydrologic exchange between the Floridan and the Surficial aquifer systems that may extend into Indian River Lagoon in the form of submarine groundwater discharge.

Conservative solutes such as Sr isotopes and $Cl^-$ concentrations, however, suggest that only a very small fraction (1 - 5%) of interstitial water is composed of meteoric groundwater. Therefore, almost all of the interstitial water must consist of recycled seawater. A combination of seep meters and naturally occurring isotopes (Ra and Rn) produced seepage rates that are generally in close agreement with one another. The observed 5-fold variation in seepage rates can be attributed to fundamental differences in each technique.

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# Karst Features and Hydrogeology in West-central Florida—A Field Perspective

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## Abstract

Karst features in west-central Florida play a dominant role in the hydrogeologic framework of the region. Urban development in karst regions present unique problems for land- and water-resource managers and can potentially impact both land and water resources if not managed adequately. Understanding how karst features control ground-water flow and respond to varying hydrologic conditions is critical for effective long-term planning and resource management. This field guide highlights the karst features of west-central Florida and the geologic units of the region that define the hydrogeologic framework and influence ground-water flow and transport. The four field trip stops include: (1) an active limestone quarry where the representative carbonate units can be seen in cross section; (2) a dry cave system that intersects land surface showing where dissolution activity has created cavities that range from subterranean conduits to large caverns; (3) Chassahowitzka Springs, a coastal spring complex, where flooded karst features form spring vents, fissures, and highly-eroded limestone at or near land surface; and (4) Health Springs, a coastal spring located 2,500 feet down gradient from a golf course and waste-water treatment plant with a spray-effluent facility. At Health Springs, the existence of a preferential ground-water flow path was documented by timing the movement of artificially dyed ground water between a well in the spray field and the spring. Ground water from the upgradient area has impacted the spring by increasing nitrate concentrations above background levels. These stops provide an opportunity to examine various karst features typical of Florida, especially in the context of their susceptibility to impact from land- and water-resource development activities.

## INTRODUCTION

Thick carbonate deposits underlying most of Florida comprise the Floridan aquifer system. The Floridan aquifer system is the principal aquifer in Florida and is among the most productive in the world. The high productivity of this aquifer is due to the development of secondary porosity caused by dissolution or karst processes (fig. 1). Karst processes characteristically develop zones of enhanced porosity within carbonate rocks creating a highly heterogeneous aquifer system with rapid rates of ground-water movement and recharge. Subsidence events caused by the collapse of materials into overlying caverns and caves can result in structural damage at land surface. More importantly karst-related features can create direct pathways for introducing surface contaminants into the ground-water system where remediation is difficult.

Most of Florida is prone to karst-related water-resource problems and examples of karst-related environmental impacts are well-documented. As karst features evolve, they also respond to changing hydrologic conditions. Evidence of hydrologic changes are generally observed at karst features. In west-central Florida examples of effects from hydrologic changes are well-documented. Ground water has been degraded by surface contaminants

(William A. Wisner, 1972)

Figure 1—Mining exposed this typical karst limestone surface which exhibits the characteristically enlarged porosity created by dissolution (from Tihansky, 1999).

(Stewart, 1982). Springs have ceased to flow or have responded to upstream sinkhole formation by temporarily increasing discharge (Peek, 1951; Trommer, 1992). Lakes have been drained by sinkholes located in the lake sub-bottom due to lowering of ground-water levels (Stewart, 1982;

In Eve L. Kuniansky, editor, 2001, U.S. Geological Survey Karst Interest Group Proceedings, Water-Resources Investigations Report 01-4011, p. 198-211

Sinclair and others, 1985). Sinkholes have formed in response to drilling wells, clearing land, and rerouting surface-water drainage (Tihansky, 1999). Water-quality changes at a number of springs document significant increases in nutrient concentrations reflecting the influence of urban and agricultural land use in the ground-water basin (Jones and Upchurch 1993, 1996; Jones and others, 1994, 1997).

Activities impacting land and water resources place a measurable strain on the water resources of the region and affect the unique ecosystems in west-central Florida. Rerouting and reducing surface drainage and altering natural recharge patterns affects water resources by subjecting new areas to increased drainage and eliminating recharge from others. Such changes alter the equilibrium that exists between subsurface cavities and overburden materials. Development of ground-water resources for municipal, industrial, and agricultural water supplies places additional stresses on the hydrologic system and can alter the balance of the natural hydrologic cycle. Increased ground-water and surface-water withdrawals lead to regional ground-water declines that can induce sinkholes to form, contribute to dewatering of wetlands and lakes, reduce spring flow and stream discharge. When fertilizers and other agricultural chemicals are applied to land surface in a karst terrane, they are often transported rapidly into the aquifer materials where they degrade the ground-water quality. All of these examples demonstrate how activities associated with urban growth can increase the susceptibility of karst aquifers to contamination from surface-water drainage.

## THE MANTLED KARST OF WEST-CENTRAL FLORIDA

The exposed land mass that constitutes the Florida peninsula is only part of a larger, mostly submerged carbonate platform that is partially capped with a mantling sequence of relatively insoluble sand and clay deposits (figs. 2 and 3) (Tihansky, 1999). In mantled karst regions, the carbonate units are not exposed at land surface, but their presence may be indicated by sinkholes and the hummocky topography that results as covering deposits settle into the irregular surface and voids within the highly soluble carbonate rocks beneath them. As a result of the depositional history and infilling processes, the sand and clay deposits vary in composition and thickness throughout Florida (fig. 3).

In west-central Florida, the thickness of the mantling deposits (overburden) overlying the carbonates influences the circulation and chemical

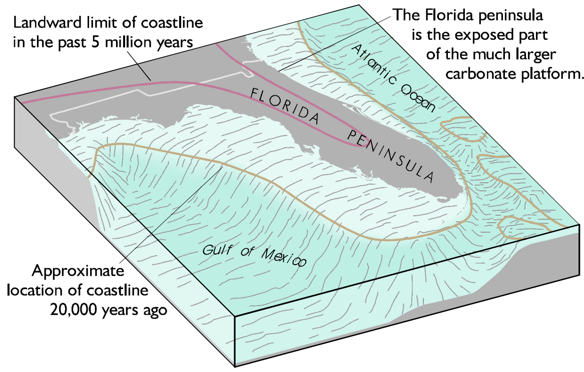


Figure 2—Changes in sea level have alternately submerged and exposed the carbonate platform (from Tihansky, 1999).

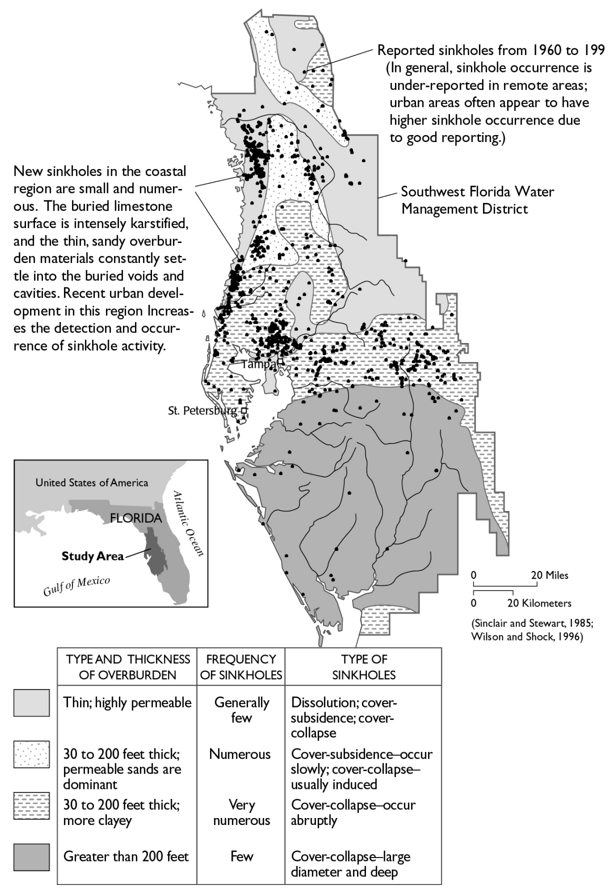


Figure 3—Type and thickness of materials mantling the carbonate units vary significantly in west-central Florida. Sinkhole occurrence throughout the region provides evidence of enhanced porosity at depth (from Tihansky, 1999).

quality of recharge waters to the Upper Floridan aquifer. Additionally, these deposits create distinct geomorphic regions that result in various types of karst features and influence ground-water flow.

The carbonate platform can be more than 3,000 feet (ft) thick and overlies a metamorphic basement (Miller, 1986). The occurrence of the top of carbonate rocks ranges from land surface to depths greater than 500 ft below land surface. The overburden deposits thicken toward the south and central parts of the platform (fig. 4). Throughout recent geologic time, the carbonate rocks of the Upper Floridan aquifer have been extensively and repeatedly subjected to chemical dissolution and depositional processes in response to sea level fluctuations (Randazzo, 1997). These chemical processes are most active near or at the water table (saturated/unsaturated interface) and near or at saltwater/freshwater interfaces (the seawater mixing zones). The spatial locations of interfaces are not temporally constant; and therefore, multiple horizons of concentrated karst features can occur within the carbonate strata. The wide fluctuations in sea-level stands over the Floridan platform were accompanied by periods of intense karst development (James and Choquette, 1988; Watts,1980). As sea- and ground-water levels rise and fall, the karst features continue to evolve. During high sea-level stands many of the karst features become submerged. Reversing head gradients convert sinkholes into flowing springs. Many of the numerous lakes and ponds of west-central Florida occupy depressions formed by overburden materials settling into cavities in the underlying limestone.

Karst features in Florida include sinkholes, springs, caves, disappearing streams, internally-drained basins, subsurface rather than surface drainage networks and highly transmissive but heterogeneous aquifers. Most of the documented karst features in west-central Florida are within 300 ft of land surface although cave divers have explored deeper passages in submerged caves. Also, exploratory well drilling has indicated the presence of enlarged fractures and cavities and associated flows at depths greater than 300 ft. General, in areas where the overburden ranges in thickness from 30 to 200 ft and the clay content is significant, subsidence activity is common and sinkholes are numerous (Sinclair and others, 1985) (fig. 3). Where permeable sands are predominant in the overburden sediments, cover-subsidence sinkholes develop gradually as the sands fall into underlying cavities. Where overburden contains more clay, cover-collapse type sinkholes are predominant. The more cohesive, less permeable clay-rich deposits deform, postponing failure until the underlying cavity grows too large and the overburden collapses into underlying voids.

South of Tampa Bay, the overburden materials thicken in excess of 200 ft and consist of cohesive siliciclastic sediments interbedded with carbonate sediments. Little surface expression of karst occurs, although the buried limestone units can have significant secondary porosity. Although sinkhole features are not common, when observed, they are usually large-diameter, deep, cover-collapse type sinkholes, indicating the presence of buried karst limestone.

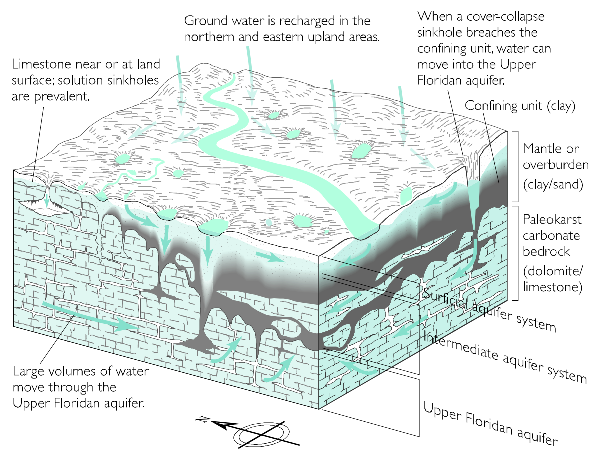


Figure 4—The regional geology influences the hydrogeologic setting and varies significantly in the west-central Florida region (from Tihansky, 1999).

## Karstification and Hydrogeology

Karstification or post-depositional alteration of geologic units plays a critical role in the hydrogeology of west-central Florida. Throughout the carbonate units, specific flow zones have developed along fractures, fissures, and bedding planes. The enlarged openings in the carbonate rocks continue to concentrate ground-water flow potentially leading to further dissolution; creating sinks and springs. The well-developed interconnected secondary porosity creates the highly-transmissive zones in the carbonate aquifers.

Because karst features evolve in response to specific hydrologic conditions that have changed significantly over geologic time, many karst features are in areas where they could not form readily today. Florida's coastal springs originally formed as sinkholes in a recharge area where acidic waters were undersaturated with respect to calcite. These springs now occur in a modern discharge area where carbonate dissolution does not occur extensively.

Today, the coastal springs discharge millions of gallons per day of ground water from the Upper Floridan aquifer (Mann and Cherry, 1970).

| SYSTEM | SERIES | STRATIGRAPHIC UNIT | | HYDROGEOLOGIC UNIT | |
|---|---|---|---|---|---|
| QUATERNARY | HOLOCENE PLEISTO-CENE | UNDIFFERENTIATED SAND AND CLAY DEPOSITS | | SURFICIAL AQUIFER SYSTEM | |
| TERTIARY | PLIOCENE | HAWTHORN GROUP | PEACE RIVER FORMATION | INTERMEDIATE AQUIFER SYSTEM OR INTERMEDIATE CONFINING UNIT | |
| | MIOCENE | | ARCADIA FORMATION | | |
| | | | TAMPA MEMBER | FLORIDAN AQUIFER SYSTEM | UPPER FLORIDAN AQUIFER |
| | OLIGOCENE | SUWANNEE LIMESTONE | | | |
| | EOCENE | OCALA LIMESTONE | | | |
| | | AVON PARK FORMATION | | | |
| | | | | | MIDDLE CONFINING UNIT |
| | PALEOCENE | OLDSMAR AND CEDAR KEYS FORMATIONS | | | LOWER FLORIDAN AQUIFER |

Figure 5—Hydrogeologic framework (modified from Knochenmus and Robinson, 1996).

## HYDROGEOLOGIC FRAMEWORK

The depositional environment and resultant hydrogeologic framework underlying west-central Florida, are discussed in Miller (1986) and Randazzo and Jones (1997). Hydrogeologic and stratigraphic units of west-central Florida are shown in figure 5. The massive carbonate sequence in west-central Florida is comprised of the Oldsmar and Cedar Keys Formations, the Avon Park Formation, the Ocala Limestone and the Suwannee Limestone. Overlying the carbonate sequence is the Hawthorn Group, comprised of interbedded carbonates and siliciclastics. Above the Hawthorn Group is an unconsolidated and undifferentiated unit comprised of quartz sand, clay, phosphate, organics (peat) and shell deposits that blanket nearly all of Florida in varying thickness and composition. Varying amounts of clay, generally at the base of the undifferentiated surficial deposits, provide confinement of the underlying rocks. However, throughout west-central Florida, karst features, such as sinkholes, breach clay deposits that would otherwise have provided confinement of the underlying carbonate rocks (Trommer, 1987).

Regionally, the carbonate units dip to the south and west and the overlying Hawthorn Group thickens and becomes a significant Intermediate aquifer system (fig. 6). Near the Gulf of Mexico, north of Tampa Bay, the carbonate rocks are at or near land surface and karst features can be observed. South of Tampa Bay, the clay units thicken, confining the carbonate units. Increased confinement reduces recharge of chemically aggressive water and the carbonate rocks are less susceptible to dissolution activity. Subsidence activity is less frequent and karst features are sparse at land surface (figs. 3 and 4).

### Hydrostratigraphy

One or more aquifer systems separated by confining units, occur in west-central Florida. Correlation between geologic units and aquifer systems typically coincides with lithostratigraphic boundaries. The Surficial aquifer system is predominantly sand, the Intermediate aquifer system is interbedded siliciclastics and carbonates, and the Floridan aquifer system is massive carbonates. The relationship between lithostratigraphic and hydrostratigraphic units is described by the Southeastern Geological Society (1986), Green and others (1995) (fig. 5).

The uppermost aquifer, where present, is the

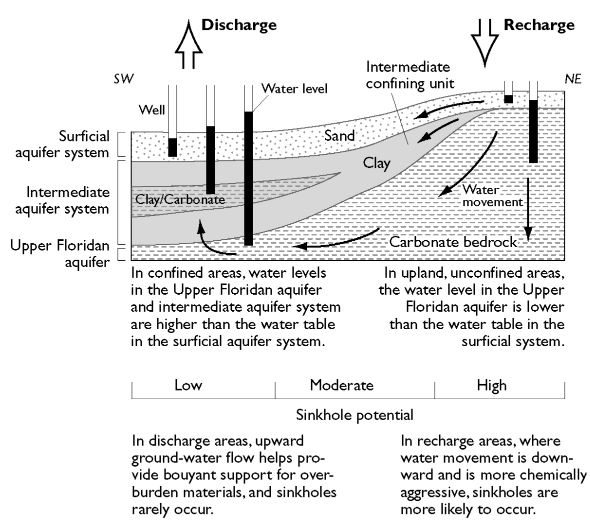


Figure 6—The potential for karst features to develop in west-central Florida is controlled by regional variations in geology and hydrology (from Tihansky, 1999).

Surficial aquifer system. The Intermediate aquifer

In Eve L. Kuniansky, editor, 2001, U.S. Geological Survey Karst Interest Group Proceedings, Water-Resources Investigations Report 01-4011, p. 198-211

system occurs south of Tampa Bay. When only the low permeability siliciclastic units of the Hawthorn Group are present, they form a confining unit that separates the Surficial and the Floridan aquifer systems. The Floridan aquifer system occurs throughout the state and is artesian where it is confined. The degree of confinement depends on the thickness and composition of the overburden materials.

### Surficial Aquifer System

The Surficial aquifer system is composed primarily of Pliocene-Holocene unconsolidated siliciclastics (quartz sand, clay, organics and shell), and generally correlates with undifferentiated sand and clay deposits that blanket west-central Florida. Surficial deposits may be missing, having been eroded away and exposing limestone at the surface. In contrast, these same deposits may exceed 100 ft in thickness where they infill karst features or are remnants of ancient dune deposits. The limited thickness of the Surficial aquifer system in most areas of west-central Florida makes the aquifer a limited water resource. North of Tampa Bay, a regionally persistent surficial aquifer system does not occur and the underlying aquifer is unconfined. In areas to the south and east of west-central Florida, the surficial aquifer system is sufficiently thick to be utilized as a water supply.

### Intermediate Aquifer System

The Intermediate aquifer system includes all permeable or water-bearing units and confining beds occurring between the overlying Surficial aquifer system and the underlying Floridan aquifer system (Duerr and Enos, 1991). Within west-central Florida, the intermediate aquifer system coincides with the Hawthorn Group and ranges from 0 to more than 700 ft thick. In the northern part of west-central Florida, the Hawthorn Group has been eroded leaving locally occurring patches of sediments (Scott, 1988). Generally, north of Tampa Bay, the Intermediate aquifer system (Hawthorn Group) is not regionally extensive. When the remaining sediments are predominantly clays, the unit is designated as the intermediate confining unit. In the northern region the remnant sediments are siliciclastics comprising the Peace River Formation and range in thickness from less than 10 ft to more than 140 ft.

### Floridan Aquifer System

The Floridan aquifer system underlies all of Florida and parts of Georgia, Alabama and South Carolina. The Floridan aquifer system is divided into the Upper and Lower Floridan aquifers, generally separated by the middle confining unit (Miller, 1986, 1997). In west-central Florida, the thickness of the Upper Floridan aquifer ranges from 600 ft to more than 1,400 ft (Wolansky and Garbade, 1981). The top of the Upper Floridan aquifer is typically considered the occurrence of vertically persistent carbonates. The bottom of the Upper Floridan aquifer occurs within the lower Avon Park Formation, where vertically and laterally persistent evaporite minerals (gypsum and anhydrite) are present in the carbonate rocks (Ryder, 1985). These evaporites, which occur either as beds, vug fillings, or as intergranular minerals within the carbonate matrix, cause a major decrease in permeability. Referred to as the middle confining unit of the Floridan aquifer system, the evaporites are of regional extent and underlie the Upper Floridan aquifer throughout west-central Florida (Hickey, 1990).

In the west-central Florida, the Upper Floridan aquifer typically consists of the Avon Park Formation, Ocala Limestone and Suwannee Limestone. Additionally, in areas where the Tampa Member (Arcadia Formation) of the Hawthorn Group is hydraulically connected to the underlying carbonates (Suwannee Limestone), the Tampa Member is included in the Upper Floridan aquifer.

## FIELD TRIP STOP DESCRIPTIONS

Four field trip stops provide access to various examples of karst features of west-central Florida. The map locations of the stops are included in the appendix at the end of this paper. From north to south, these stops include: 1-The Southdown Limestone Quarry, 2-Dry Caves of Citrus County, 3-Chassahowitzka Springs complex, and 4-Health Springs.

### 1-Southdown Limestone Quarry

The Southdown Quarry is an operating limestone quarry that mines limestone for aggregates and manufacture of concrete. Fine and coarse aggregate is manufactured from limestone deposits with softer deposits favored for limerock base material. Limestone varies from very hard and dense to soft and porous and may contain significant quantities of fossils. Karst features are often related to the introduction of allogenic materials into the pure carbonate units via solution pipes, sinkholes and fissures. The presence of silica sand and chert along joints and bedding planes often requires that mining activities process the materials either by selectively mining, blending materials, or removing fines.

Carbonate units exposed in this quarry include the Avon Park Formation, the Ocala Limestone and the Suwannee Limestone. The Middle Eocene Avon Park Formation (Miller, 1986) occurs in the subsurface throughout west-central Florida. This unit varies from light gray to brown dolostone to cream to light orange limestone with minor clay interbeds and dispersed organic laminations. Accessory minerals include chert, pyrite, and gypsum, with gypsum becoming more abundant with depth. Although the uppermost part of the Avon Park Formation varies between limestone and dolostone, dolostone predominates deeper within the unit, especially toward the south. Porosity in this formation is generally intergranular in the limestone section. Fracture porosity occurs in the more densely recrystallized dolostone, and intercrystalline porosity is characteristic of sucrosic textures. Pinpoint vugs and fossil molds are present to a lesser extent. The most diagnostic fossils include the foraminifers *Dictyoconus americanus* and *Coskinolina floridana*. The echinoid *Neolaganum (Peronella) dalli* is also common (J. Arthur, Florida geological Survey, 2000, written commun.).

The base of the Avon Park Formation occurs at depths ranging from 1,100 ft to 1,850 ft below sea level varying in thickness from 1,000 ft in the north to 1,500 ft south (Miller, 1986). In the field trip area, the top of the formation is encountered from within 20 ft of land surface to a depth of 425 below sea level in the southwest. In many cases, high gamma-ray activity at the top of the Avon Park Formation is due to thin (<2 inches) layers of organic material.

The Ocala Limestone unconformably overlies the Avon Park Formation. The Upper Eocene Ocala Limestone consists of white to light-gray to light-orange limestone with a diverse fossil assemblage. More specifically, the lithology of this formation ranges from a variably chalky wackestone or packestone in the upper parts to a biogenic packstone to grainstone in the central and lower parts of the unit. Accessory constituents include organics, clay, dolomite and chert. Porosity is variable within this unit and is generally moldic and intergranular with occasional macrofossil molds. This formation contains characteristic fossils such as the foraminifers *Lepidocyclina* spp., Nummulites (Operculinoides) and echinoids such as *Eupatagus antillarum*. Other fossils observed in the unit include mollusks, bryozoans and corals.

The Ocala Limestone is typically bound by unconformities. Depths to the top of the formation range from land surface to 285 ft below sea level. The Ocala Limestone extends throughout west-central Florida except for some regions to the north and east towards the central part of the state. The Ocala Limestone obtains a maximum thickness of 230 ft. These maximum depths and thicknesses occur in regions on the flanks of the Ocala Platform, which trends south-southeast.

Gamma-ray logs for the Ocala Limestone consistently exhibit low gamma-ray activity. In some southern areas, the Ocala Limestone gamma-ray signature is “quiet” when compared to the underlying Avon Park Formation and the overlying Suwannee Limestone. In cases where the Ocala Limestone is dolomitized, the gamma-ray logs may exhibit a slightly higher and more sporadic signal. Many peaks in the gamma-ray logs correlate with the presence of organics.

The Lower Oligocene Suwannee Limestone ranges from a light-gray to yellowish-gray packstone to grainstone. These carbonates are variably moldic with trace amounts of sand and clay within the upper parts. Trace amounts of chert and organics occur throughout the unit. Fossils in the unit include mollusks, echinoids (primarily *Rhyncholampus gouldii*), abundant miliolids and other benthic foraminifers including *Dictyoconus cookei*. This formation unconformably overlies the Ocala Limestone and is unconformably overlain either by Hawthorn Group units or UDSC sediments. In several areas towards the south, the Suwannee Limestone is less than 20 ft below land surface. It is limited in extent towards the north and is reported to occur as exposed remnant boulders where it thins (Campbell, 1989). Depth to the top of the Suwannee Limestone ranges from 80 ft below to 132 ft above sea level, where present. The unit thickens to the south and west, ranging up to 255 ft thick. The Suwannee Limestone is characterized by gamma-ray activity that has an overall higher count rate than the underlying Ocala Limestone. Additionally, there exists much more variability in its signature relative to the Ocala Limestone. This variability in the gamma-ray signature correlates with dolomite, clays and organics within the formation.

## 2-Dry Caves of Citrus County

The history of the development of caves visited during this field trip (Brinkmann and Reeder, 1994) is included in the field trip package. These caves were formed by structural and chemical processes. Structural uplift created northwest-southeast trending joints and dissolution was enhanced by geochemical reactions associated with ground-water mixing. The mixing of waters with variable salinity and acidity enhanced carbonate dissolution. The caves visited on this trip are fossil remnant segments from a much larger, interconnected cave system that formed in the Ocala and Suwannee Limestones. Subsequent lowering of base level by erosion and collapse has destroyed and infilled many of the original caves.

Peace Sign cave is predominantly a large sinuous conduit with the land surface opening occurring at the base of a tree. Vandal cave is a good example of how collapse and subsidence processes destroy the original cave features. Bats and other typical cave fauna are usually present in the smaller, less-frequently visited caves in this complex.

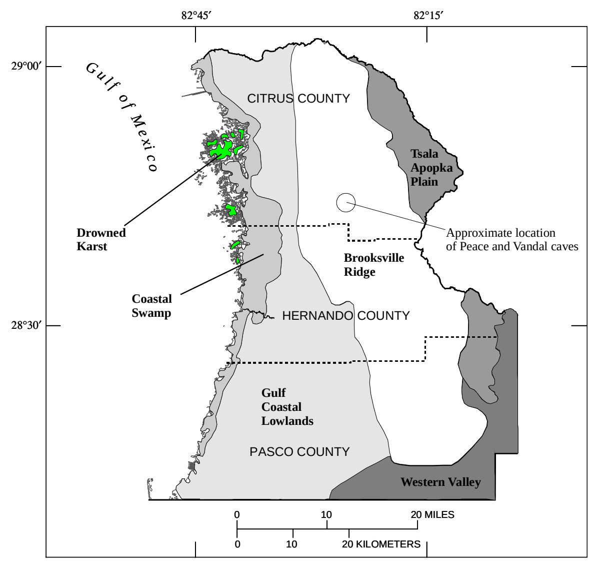


Figure 7—Physiographic regions of west-central Florida (modified from White, 1970)

## 3-Chassahowitzka Springs Complex

The springs that contribute flow to the Chassahowitzka River occur in the physiographic region designated as the Coastal Swamp (White, 1970) (fig. 7). This region is an area of upward flow from the Upper Floridan aquifer and active sinkhole formation is minimal (0-2 karst features per square mile). To the east, in the sand hills of the Gulf Coastal Lowlands, recharge conditions, exist so the karst feature density is higher (10-25 solution features per square mile) and the well drained soils support a unique scrub habitat (HydroGeoLogic, 1997 and Wolfe, 1990).

The springs visited on this trip include the Chassahowitzka Main Spring, Bubba Spring (also known as Chassahowitzka Number 1), Crab Creek Head Spring and Baird Creek Spring. These springs contribute the majority of the freshwater flow to the Chassahowitzka Springs complex.

The Chassahowitzka River is a shallow, flat, and sluggish stream that meanders through about 6 miles of lowland swamps and tidal marshes to the Gulf of Mexico. At least 12 springs contribute flow to the Chassahowitzka River and have been described by Wetterhall (1965), Yobbi (1992), and Jones and others (1997) (fig. 8).

Subsurface geology is reflected in the types of

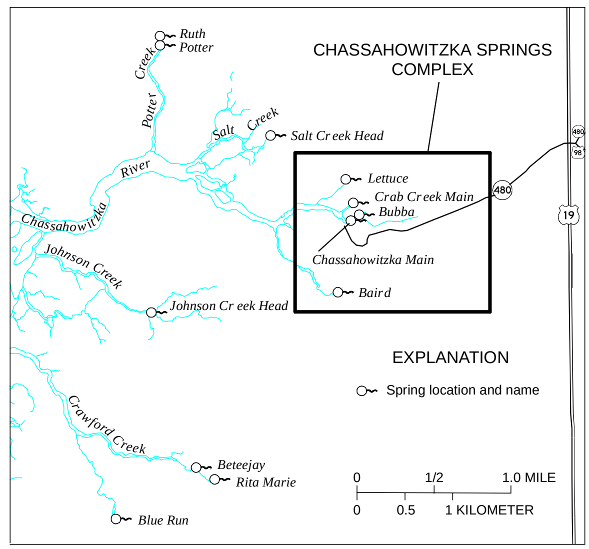


Figure 8—Springs and surface discharge network of the Chassahowitzka Springs complex.

spring vents observed in west-central Florida and can be viewed in the Chassahowitzka Springs complex. The types of spring vents include: (1) linear fracture, (2) circular rock vent (vertical pipe), and, (3) sediment filled vent (sand boil). The types of vents that form usually reflect characteristics of the rock. Hard, brittle zones in limestone units maintain larger openings such as caverns and fractures while softer limestones collapse more readily.

The Chassahowitzka Main Spring is an example of a sediment-filled vent. The spring boils from a sand bottom along a crevice approximately 25-ft long and in 35 ft of water depth. The spring pool is approximately 150 ft in diameter and is a 50 ft wide cone located in the middle of the river channel. Salinity ranges from 700 to more than 4,000 microsiemens per centimeter (us/cm). The spring is affected by tides with flow ranging from near zero to more than 52 million gallons per day (80 cubic ft per second ($ft^3$/sec)).

Bubba Spring is the largest of several springs that contribute to flow in the unnamed tributary above

Chassahowitzka Main Spring. It is an example of a circular rock vent. It is comprised of two vertical pipes connected by a 15-foot horizontal conduit. The flow from Bubba Spring emanates from a small opening in the horizontal passage, midway between the two vertical pipes. Bubba Spring is the shallowest in the Chassahowitzka Springs complex and was fresh enough to be used as a water supply in the past. However, these waters are now contaminated by leakage from septic tanks.

Crab Creek Head Spring and at least three other springs that contribute flow to Crab Creek are additional examples of circular rock vents. Crab Creek Head Spring is located in 13 ft of water. The vertical pipe is intersected by a horizontal fracture about seven ft below the water surface. These springs discharge brackish water (greater than 3,000 us/cm) and spring flow is relatively constant at 3.2 million gallons per day (50 $ft^3$/sec).

Baird Creek Spring is an example of a linear fracture vent. The fissure is more than 20 ft long and is in about 4 ft of water. This spring discharges brackish water (greater than 5,000 us/cm) and the salinity varies significantly during a tidal cycle.

## 4-Health Springs—impacts of land use in a karst region

Health Springs serves as one of many examples where land-use activities in a karst terrane affect ground-water quality. The region contains coastal springs and karst uplands that are characterized by internal drainage and variable confinement between the surficial aquifer system and the Upper Floridan aquifer (fig. 9). Land use upgradient of the spring was used historically for citrus agriculture but at present includes an extensive golf course, a wastewater treatment facility, and residential and commercial properties. Activities at the wastewater treatment plant have included land application of dried sludge, spray irrigation and ponding of treated effluent. The site is presently planned for a county park; however, elevated nutrient and bacteria concentrations in the spring discharge have delayed this action.

In April 1988, several cover-collapse sinkholes developed in an area where effluent from a wastewater treatment plant is sprayed for irrigation. Ponding of the effluent occurred while water levels were at their seasonal low. The maximum seasonal head difference combined with surface loading likely contributed to the formation of several sinkholes which drained the effluent into the ground-water system. Within several days of sinkhole formation, discharge at Health Springs, 2,500 ft downgradient, increased from 2 $ft^3$/sec to 16 $ft^3$/sec (Trommer 1992). A dye-tracer study confirmed the existence of a preferential ground-water flow path linking the upland spray field with the spring (Tihansky and Trommer, 1994). Dye injected into a well located in the sprayfield was detected in the spring water and in a well adjacent to the spring (fig. 10). Ground-water velocity was about 160 ft per day (ft/d) which is 250 times greater than estimates of the regional ground-water velocity (0.65 ft/d) in this area.

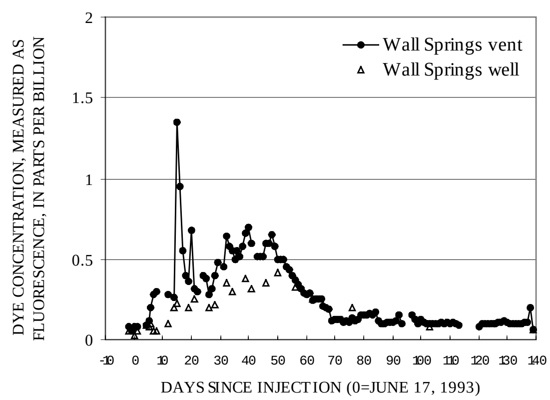


Figure 10—Dye concentrations, measured as fluorescence, at Health Springs and an adjacent well during a tracer test illustrate the rapid movement of ground water.

Background nitrate concentrations for ground water in Florida generally are less than 0.02 milligrams per liter (mg/L). Water samples collected at Health Springs since 1982 have nitrate concentrations ranging from 2 to more than 10 mg/L. These elevated values reflect the impact of land use in the spring's recharge area.

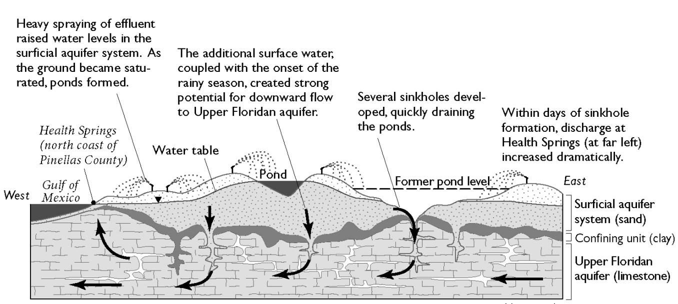


Figure 9—Sinkholes provide a direct hydrologic link between karst uplands and Health Springs along the coast (from Tihansky, 1999).

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## Conclusion

The four field trip stops provide a quick glimpse of the variability in the types of landforms and features characteristic of Florida karst. Examining the rocks in a quarry section reveals the variability of the units in terms of geologic fabric and texture. These initial differences often determine the type and extent of karst development and distribution of zones with enhanced secondary porosity. The heterogenous distribution of overlying mantling deposits further controls the development of karst. The dry caves and the flowing springs observed on this trip demonstrate how karst features function on both the recharge and discharge ends of the hydrologic realm. Each of these regions has unique and significantly different hydrologic controls and potential environmental impacts.

The full range of hydrologic impacts have been documented at Health Springs, where sinkholes and water-quality impacts in the recharge zone have been directly linked to water-quality changes in the discharge region at the spring.

The features observed on this trip include a full range of karst features in west-central Florida. More importantly, close examination of these features provides an opportunity to see how vulnerable water resources are to impacts related to karst landforms. Impacts to resources of the region range from loss of structural support and foundation integrity at land surface to degrading water quality in ground- and surface-water resources. Most of Florida is prone to karst-related impacts on land and water-resources, and the examples seen on this trip are representative of many other locations.

The distribution of highly variable transmissive zones creates a complex hydrogeologic framework and standard analytical and numerical methods often do not adequately quantify the complexities of karst flow systems. Further study of karst systems and increased understanding combined with effective management will minimize detrimental effects to water resources in karst regions.

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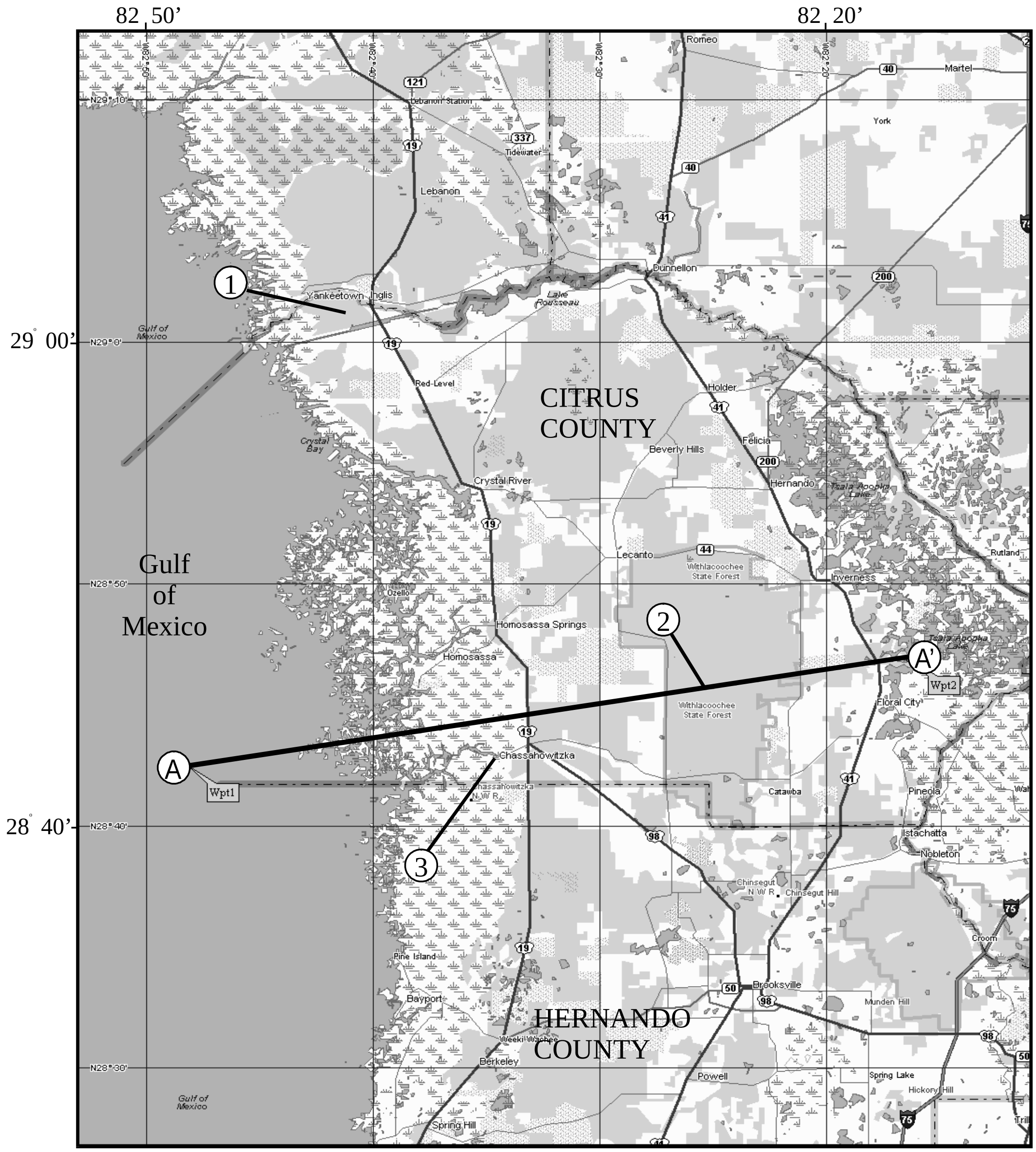
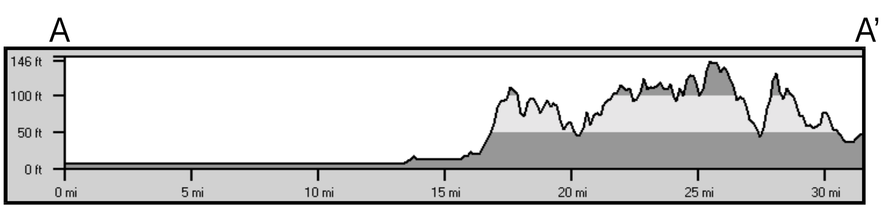
Appendix—Field trip stops 1-4. Stop 1 is Southdown Limestone quarry. Stop 2 is the dry cave complex of Citrus County, Stop 3 is the Chassahowitzka Springs complex. Stop 4 is Health Springs, known locally as Wall Springs, and the surrounding upland area. A representative topographic cross section A-A’ shows the irregular topography and abrupt transition from upland to coastal plan. The section extends from upland dry caves (field trip Stop 2) to the coast near Chassahowitzka Springs (field trip Stop 3). Cross section B-B’ shows the upland coastal strip of northwestern Pinellas County (field trip Stop 4) and the Lake Tarpon Basin separating the coast from the main peninsula of Florida.




3-D TopoQuads 1999 Copyright DeLorme Yarmouth, ME 04096 1 mile Scale: 1:350,000

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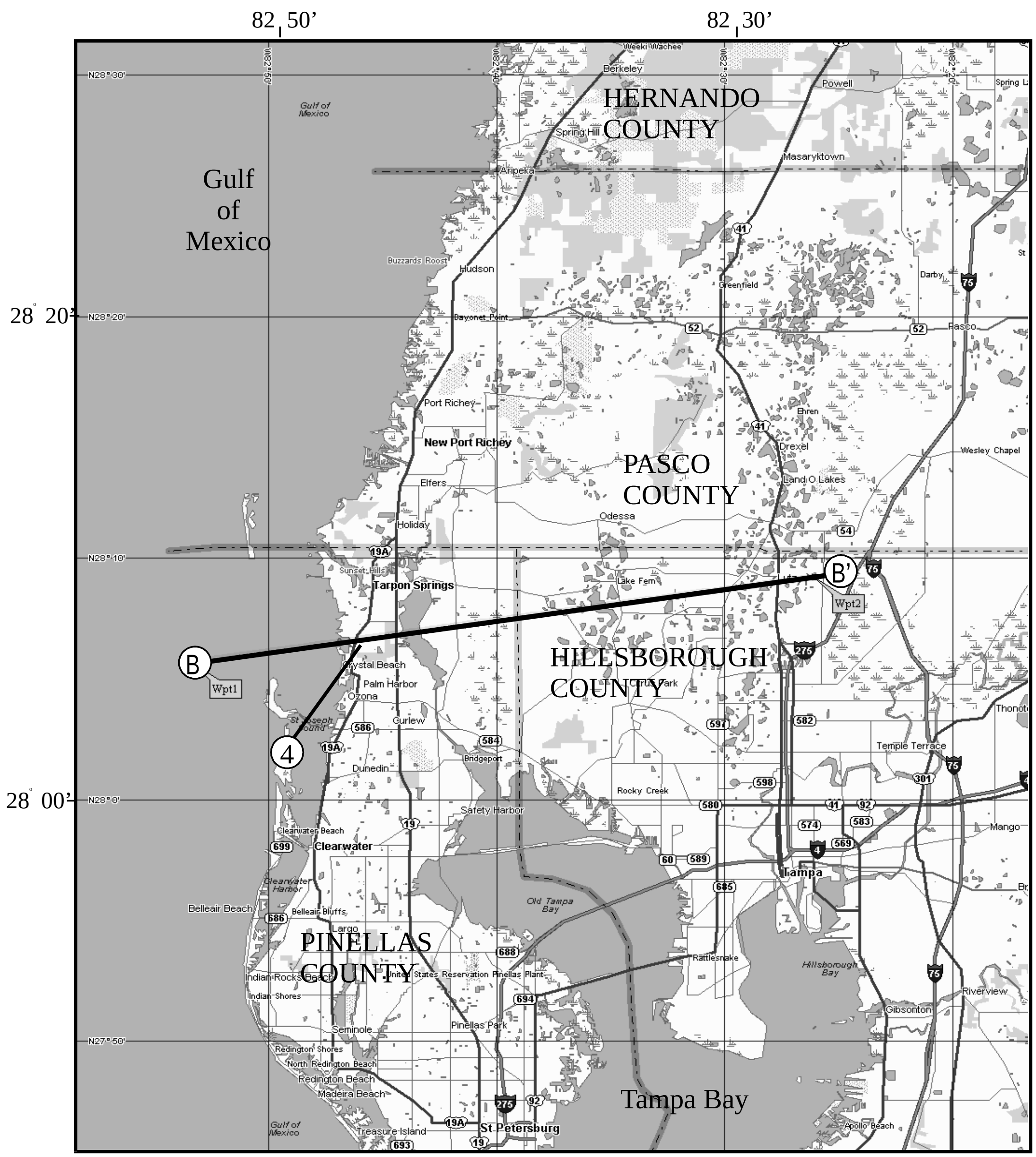


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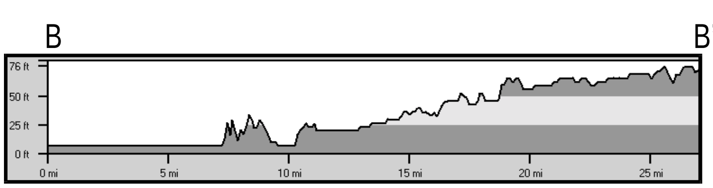


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