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IN THIS ISSUE:

SELECTED PAPERS FROM THE SECOND APPALACHIAN KARST SYMPOSIUM

AND OTHERS...

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Front cover: Surveying in Morrill's/Worley's Cave, Tennessee. Photo by Y. Cho.

EDITORIAL
Frontiers of Appalachian Karst Research

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Following the success of the first Appalachian Karst Symposium held at Radford University (Kastning and Kastning, 1991), the second Appalachian Karst Symposium was held at the Gray Fossil Site, East Tennessee State University on May 7–10, 2008. The symposium was sponsored by the General Shale Brick Natural History Museum and the Office of Research and Sponsored Programs at ETSU and supported by the ESRI software company, the Environmental Protection Agency Region III, the Virginia Department of Conservation and Recreation Natural Heritage Program, the Virginia Cave Board, the Cave Conservancy Foundation, the Cave Conservancy of the Virginias, P.E. LaMoreaux & Associates, Inc., the Karst Waters Institute, the National Speleological Society, and the *Journal of Cave and Karst Studies*.

Seventy-five people attended the symposium to foster communications and to promote the exchange of ideas among all professionals concerned with scientific studies and environmental conservations in the Appalachian karst region. The symposium included three keynote presentations, by Gregory Springer and Harry Rowe, Art Palmer, and Will White. A welcome reception and two lectures, by Barry Beck and Harry Moore, were open to the public on May 7. Keynote, oral, and poster presentations were presented during the next two days. Presentations during the symposium were highly interdisciplinary and included research on karst hydrology and geomorphology, cave exploration and conservation, resource management and database development, biological research, paleontology, paleoclimate, archaeology, and engineering and geotechnical methods. Communications between karst professionals regarding the Karst Information Portal and metadata development have been highly active in the past few years (Gao and Zhou, 2008). Almost all symposium attendees participated in a panel discussion on issues related to karst-database development, data sharing, resource protection, and cave conservation. The symposium banquet featured Russell Graham's talk "Mammal Response to Late Quaternary Climate Fluctuations along the Appalachian Gradient – Implications for Future Global Warming." On May 10, nearly forty people attended a one-day field trip to visit caves and karst sites in northeastern Tennessee and southwestern Virginia. In the late afternoon on May 9, a guided tour to the Gray Fossil Site was offered by the Don Sundquist Center of Excellence in Paleontology at ETSU. Sid Jones and Robert Benfield led an optional field trip to visit karst springs in northeastern Tennessee. Following the tradition of the first Appalachian Karst Symposium, a Friends of Karst (FOK) get-together convened on May 9. The FOK is an informal organization of individuals interested in cave and karst studies. Many FOK members attended the symposium. One of the outcomes of the FOK gathering was an action plan for future Appalachian Karst Symposia. The next symposium is planned for summer 2012.

To insure the quality and scientific value of the symposium, all manuscripts and abstracts were reviewed by experts in the field of cave and karst studies. More than thirty abstracts were submitted to the symposium. The symposium proceedings including all abstracts and field trip guide are available at the Appalachian Karst Symposium website (http://www.etsu.edu/cas/geosciences/appkarst/). Five manuscripts resulting from the symposium are included in this issue of the *Journal of Cave and Karst Studies*. One paper was published in an earlier issue of the *Journal* (Springer, et al., 2009). This paper presented a record of Holocene hydroclimatology for a humid, temperate watershed in the Appalachian Mountains of eastern North America. Two of the authors gave a keynote address about this paper during the symposium.

This issue includes five papers. White reviews the erosional processes of Appalachian fluviokarst and discusses how the Applachian karst has evolved since the late Miocene. This paper was presented as the first keynote address at the symposium. Palmer discusses the use of cave data in quantitative validation of hypotheses for maze-cave origin, the interpretation of geochemical processes that are rarely seen at the surface, and the development of wells and assessment of potential contaminant transport. This paper was also presented as a keynote address at the symposium. Schwartz and Orndorff present a thorough hydrogeologic investigation of a Mississippian scarp-slope karst system in the central Appalachians. Orndorff and Hutchins describe a unique sampling technique to monitor the distribution and abundance of an aquatic subterranean isopod. Lera gives a detailed review of the Virginia Cave Protection Act and presents many examples of actions to preserve the educational, recreational, scientific, historic, and economic values of Virginia caves and karst.

In summary, the papers in this issue represent some recent studies in a variety of disciplines concerning karst hydrology, geomorphology, biology, and cave conservation in the Appalachian karst region. Traditionally, many influential karst studies were conducted by scientists in the Appalachian karst community. Research on karst is now underway in many institutions and agencies and has much wider scientific and societal implications than previously recognized (Martin and White, 2008). This issue represents only one look at the many active karst research and conservation activities in the Appalachian region. Many interdisciplinary and innovative karst research projects are currently underway in the Appalachian karst community. Monitoring sites and field stations such as the Gray Fossil Site (2009) and the Karst Field School (2009) at East Tennessee State University will become valuable resources for education and research activities in the twenty-first century.

Acknowledgments

Symposium co-chairs Blaine Schubert and Wil Orndorff were involved in all phases of the symposium. I thank all the organizing committee members and sponsors for their supports to ensure a successful symposium. I acknowledge all the efforts of Ira Sasowsky, Malcolm Field, Scott Engel, and numerous reviewers to ensure a high-quality special issue on Appalachian karst.

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THE EVOLUTION OF APPALACHIAN FLUVIOKARST: COMPETITION BETWEEN STREAM EROSION, CAVE DEVELOPMENT, SURFACE DENUDATION, AND TECTONIC UPLIFT

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Abstract: The long and complex depositional and tectonic history of the Appalachians has produced a substrate of folded and faulted sandstones, shales, and carbonate rocks (leaving aside the metamorphic and igneous core). The Appalachian fluviokarst is an evolving landscape developed on the carbonate rocks. The erosion of surface streams competes with dissolutional processes in the carbonate rocks, and both compete with tectonic uplift of the eastern margin of the North American plate. The Appalachians have undergone erosion since the Jurassic and 5 to 15 km of sediment have been removed. Many karst landscapes have come and gone during this time period. The earliest cosmogenic-isotope dates place the oldest Appalachian caves in the early Pliocene. Various interpretations and back-calculations extend the recognizable topography to the mid to late Miocene. Much of the present-day karst landscape was created during the Pleistocene. There have been many measurements and estimates of the rate of denudation of karst surfaces by dissolution of the carbonate bedrock and many estimates of the rate of downcutting of surface streams. Curiously, both of these estimates give similar values (in the range of 30 mm ka^{-1}), in spite of the differences in the erosional processes. These rates are somewhat higher than present-day rates of tectonic uplift, leaving the contemporary landscape the result of a balance between competing processes. Introduction of tectonic forces into the interpretation of karst landscapes requires consideration of the long-term uplift rates. In the Davisian point of view, uplift was episodic, with short periods of rapid uplift followed by long static periods that allowed the development of peneplains. In the Hackian point of view, uplift has occurred at a more or less constant rate, so that present topography is mainly the result of differential erosion rates. Attempts to back-calculate the development of karst landscapes requires a conceptual model somewhere between these rather extreme points of view.

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Introduction

The Appalachian Mountains are a belt of folded and faulted Paleozoic rocks that extend southwestward roughly 3000 km from the Canadian Maritimes to central Alabama, where they are covered with young Coastal Plain sediments. The overall width of the belt ranges from 300 to 500 km. The Appalachian karst is composed of a loosely connected set of karst drainage basins that occur in the exposures of carbonate rock, mainly in the folded Appalachians and around the margins of the Appalachian plateaus. The karst regions of concern in this paper span an extensive area from the Mohawk Valley of New York to central Alabama and from the western foot of the Blue Ridge Mountains of Virginia and Tennessee to the western margin of the Cumberland Plateau in Tennessee and Kentucky. Taken as a whole, the Appalachians are one of the world's great karst areas.

The objective of the present paper is to interpret the evolution of the Appalachian karst by comparing the rates of the various processes responsible for its development. Most of the Appalachian karst is fluviokarst, and as a result, there are competing rate processes that together produce the observed landscape. Weathering of non-carbonate rocks, valley deepening by fluvial processes, and chemical denudation of exposed carbonate rocks combine to sculpt the landscape. Regional uplift serves to keep the erosive processes activated. Caves that have developed in response to local base levels have often been taken as markers for pauses in the downcutting of valleys. However, caves drift upward, riding the regional uplift, so they do not form fixed markers for absolute elevations.

This paper builds on early work on the Appalachian karst. Studies have been made of stream profiles (White and White, 1974; White and White, 1983), drainage-basin properties (White and White, 1979), and rates of carbonate-rock denudation (White, 1984). An earlier discussion of the evolution of the Appalachian karst attempted to relate karst surfaces to the classic erosion surfaces long identified in the Appalachians (White and White, 1991). A more recent and broader discussion of the evolution of karst landscapes draws heavily on the Appalachians for examples (White, 2007). The present paper takes up the evolutionary theme again, this time with a better recognition of the role of tectonic uplift in driving karst processes.

THE APPALACHIAN KARST: THE LONG VIEW

The rocks of the Appalachians record a long history of basin filling, plate collisions, mountain building, and erosion, extending back at least to Grenville time, 1.2 Ga. The very complex geology that has resulted is summarized for the north-central Appalachians by Faill (1997a, 1997b, 1998). The sequence of orogenies and depositional basins provides the three main groups of carbonate rocks that support the Appalachian karst: the Cambro-Ordovician limestones and dolomites, the Silurian/Devonian limestones, and the Mississippian limestones. For detail concerning Appalachian tectonics and geologic history, see Hatcher et al. (1989). Overviews of Appalachian geomorphology are given by Fenneman (1938), Thornbury (1965), and Hack (1989).

The development of the Appalachian karst depends on erosion of overlying clastic rocks and consequent exposure of older carbonate rocks to denudation and cave development. The earliest event of interest to karst development was the last of the major Appalachian tectonic events, the Alleghany Orogeny in Permian time. This major plate collision produced the broad-scale structures that guide the development of contemporary karst features. The succeeding Mesozoic period was one of plate rifting, with extensional faults and infilling of graben structures by rapidly eroded material represented by the Triassic red beds and fanglomerates in Pennsylvania. Only with the opening of the Atlantic Ocean in Cretaceous time could the ancestral versions of the present drainage systems begin to take shape.

Calculations based on mass balance suggest that the Appalachian Mountains at the beginning of the Mesozoic were an Andes-like chain with a maximum relief on the order of 3500 to 4500 meters (Slingerland and Furlong, 1989). According to the time scale of Gradstein et al. (2004), the Mesozoic extended from 251.0 to 65.5 Ma ago. During that 185.5 Ma interval, except for some basin-filling with mainly Triassic sediments, the Appalachians were subject to erosion. How much material has been eroded away, and when, is conjectural, since few records remain. MacLachlan (1999) claimed that approximately 15 km of sediment were removed from southeastern Pennsylvania during the Mesozoic. Judson (1975) proposed 6 km of removal from the Valley and Ridge, but only one km or less from the Allegheny Plateau. Most investigators are of the opinion that 90% or more of the erosion took place during the Mesozoic, so that the Appalachian topography was close to its present form by the beginning of the Cenozoic.

The interpretation of Appalachian landscapes taking on roughly their present form by the end of the Mesozoic poses a significant problem for the interpretation of karst development. The entire 65.5 Ma of the Cenozoic is available for further erosion, carbonate denudation, and cave development. Somewhere, in this span of time, there evolved an erosion surface, generally called the Schooley Peneplain, which is represented by the quartzite ridge-tops of the folded Appalachians and the uppermost elevations of the Appalachian Plateaus. Consistent with the notion that erosion of the high Appalachians was largely complete by the end of the Cretaceous, a late Cretaceous or early Tertiary age is often given to the Schooley surface. Also evolved during the Cenozoic is an intermediate level, the Harrisburg Peneplain, which is widely represented by karst surfaces on limestone valley floors. Various estimates place the age of the Harrisburg Surface as mid-Tertiary. If this traditional view is accepted, the karst features and secondary valleys that cut below the Harrisburg surface have roughly 30 million years available for their development. As will be shown below, there is about a ten-fold discrepancy between the rates of karst processes and the traditional view.

THE APPALACHIAN KARST: THE GEOGRAPHIC VIEW

The Appalachians were subdivided into provinces and sub-provinces by early geomorphologists (Fig. 1). The Appalachian karst is mainly concentrated in the folded Appalachians—the Great Valley and Valley and Ridge Provinces—and on the margins of the Appalachian Plateaus—the Allegheny Plateau on the north and the Cumberland Plateau on the south. The karst of the folded Appalachians is mainly developed in the Cambrian/Ordovician limestones and dolomites and the Silurian/Devonian limestones. The karst of the plateaus in mainly developed in the Mississippian limestones. Because the Mississippian limestones thin to the north, karst development is much more extensive in the Cumberland Plateau than in the northern Allegheny Plateau.

An overview description of the Appalachian karst (White and White, 2009) and many detailed descriptions of individual areas are in preparation (Palmer and Palmer, 2009).

PROCESSES OF LANDSCAPE SCULPTURING

The sculpturing of any sort of landscape is accomplished by processes of mass transfer: solid rock and its surficial weathering products are transported by flowing water, wind, ice and, in the special case of karst landscapes, by the chemical dissolution of the carbonate rocks. Each of the landscape-sculpturing processes proceeds at a certain rate dictated by the process itself and by relevant environmental parameters such as temperature, precipitation, and, for karst processes, by available carbon dioxide.

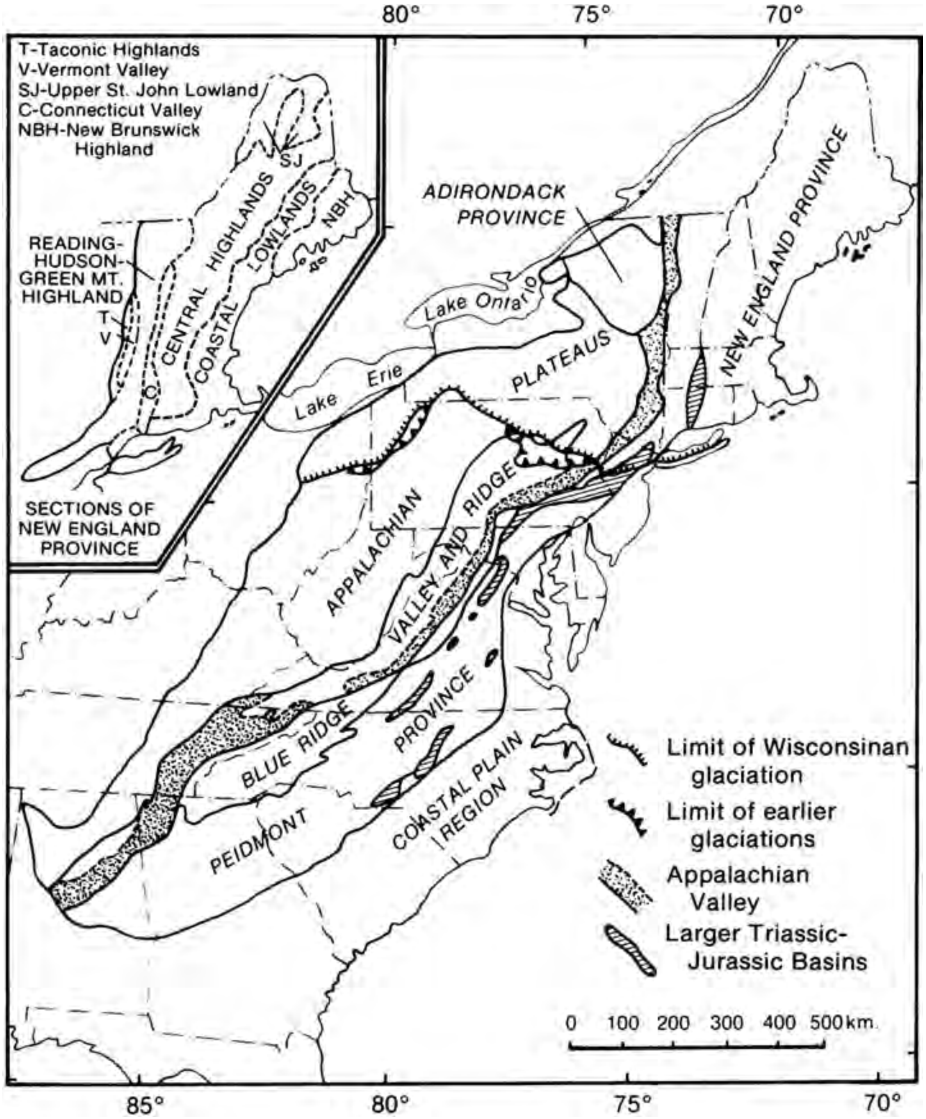

Figure 1. The Appalachian provinces. From Hack (1989).

Karst Denudation

In well-developed karst surfaces such as much of the Great Valley, the lower Greenbrier Valley, and the carbonate-floored valleys of the Valley and Ridge, there is often little surface runoff. Rainfall seeps through the soil, picks up an excess of CO_2 from the upper organic-rich horizons, and then reaches the underlying carbonate rock in the epikarst. The highly undersaturated water attacks the carbonates, often taking Ca^{2+} and HCO_3^{2-} into solution to the saturation limit defined by the soil-CO_2 partial pressure. This carbonate-laden water then migrates downward through the vadose zone along fractures and shafts. The bedrock surface is gradually lowered without dissection, thus retaining the appearance of an erosion surface.

The rate of carbonate dissolution is sufficiently fast that it can be measured directly by micrometer on exposed rock surfaces. Rates are also determined by burying rock tablets in selected locations, then digging them up after specific time periods and determining dissolution rate by weight loss. A more regional estimate can be made by measuring discharge and dissolved carbonate content of water leaving the drainage basin. For descriptions of the methods and for comparisons of measurements, see White (2000). The measured rate of carbonate-rock removal can be recalculated as an average surface-lowering rate, the rate of karst denudation.

Karst denudation has been of interest to karst geomorphologists for a long time, and many measurements have been made (for summaries see Smith and Atkinson, 1976; White, 2000; White, 2007). Rates vary from 5 to 50 mm ka^{-1}, depending on soil characteristics, climate, and precipitation. For soil-covered, temperate karst such as most of the Appalachians, a value of 30 mm ka^{-1} is representative.

Table 1. Rates of River Down-Cutting in the Appalachians.

Name	Rate, mm ka^{-1}	Method	Reference
Cheat River, W.Va.	56–63	Magnetic Reversal	Springer et al. (2004)
East Fork, Obey River, Tenn.	30	Cosmogenic Isotopes	Anthony and Granger (2004)
Juniata River, Newport, Pa.	27	Sediment Load	Sevon (1989)
New River, Pearisburg, Va.	27	Cosmogenic Isotopes	Granger et al. (1997)
South River, Grottoes, Va.	23–41	Magnetic Reversal	Kastning (1995)

Erosion of Resistant Rocks

The resistant rocks that support the high ridges and plateaus of the Appalachians are sandstones, quartzites, and conglomerates, all of which consist mainly of quartz. Quartz rocks are resistant to erosional forces. Quartz has a chemical solubility of about 10 mg L^{-1}, but the kinetics of the dissolution reaction are so slow that runoff from quartzite ridges contains much less silica than the solubility limit.

Quantitative measurements of denudation rates on quartzite are sparse. Sevon's (1989) compilation of erosion rates for the eastern United States gave only values of 2.5, 2, and 5 mm ka^{-1} as erosion rates on Quartzitic rocks. Anthony and Granger (2004) estimated the denudation rate for the quartzite conglomerate on the Cumberland Plateau as 3 to 5 mm ka^{-1}. The available values fall into the same range within a factor of two, and are smaller than carbonate denudation rates by about an order of magnitude.

River Downcutting

There is an important distinction between fluvial landscape sculpting and karstic landscape sculpting. Surface streams downcut their valleys by transport of clastic sediment. Such transport is episodic and occurs mainly during flood flows. Low-gradient streams may have the sediment in their channels balanced, such that input of fresh sediment equals the sediment discharge and there is little net deepening of the channel. Fresh sediment is injected from valley walls by solifluction, by landslides, and by other down-slope movement of weathered material from the underlying bedrock. In well-developed karst areas, drainage is internal through the conduit system. Lowering of the land surface is by dissolution of the carbonate bedrock, with most of the transport in solution along with a certain fraction of clastic load. As a result, karst surfaces tend to have low relief, except for the development of sinkholes. This contrast can be seen in many Appalachian valleys, where those valleys underlain by carbonate rock have a relatively low relief, while those underlain by shales are usually strongly dissected by surface streams.

The rate of down-cutting for streams on bedrock channels can be estimated from measured sediment loads or from the elevation difference between stream channel and dated terraces or caves on the valley walls. The latter should give more accurate values, because sediment load is more dependent on weathering in the entire basin, including all of the tributaries. Rate data for five Appalachian rivers are given in Table 1.

The downcutting rate of surface streams is very similar to the denudation rate for the limestone uplands. If the denudation rates were significantly faster than surface-stream down-cutting, all of the limestone uplands would be planated to local base levels. If down-cutting rates were significantly faster, the limestone uplands would be cut by deep canyons. In most of the Appalachians, neither is the case. Groundwater systems in areas such as the Great Valley and the limestone valleys of the Valley and Ridge are mostly shallow systems. Only when karstic drainage travels beneath sandstone-protected ridges do we find deep flow paths.

Regional Uplift

The east coast of the United States is considered to be a passive margin. The extension and rifting of the Mesozoic have become quiescent. Epeirogenic mechanisms still function, however. There is evidence that at least the Piedmont and Great Valley continue to rise as sediment is eroded from the interior, carried to the coast by rivers, and deposited off the continental shelf. Because of the shift in mass, the crustal plate is bent slightly, with a hinge line near the Fall Line at the eastern edge of the Piedmont. Superimposed on the regional uplift is isostatic rebound from retreating glaciers in the northern part of the region, as well as the effects of rising and falling sea levels during the Pleistocene. Terraces in the Susquehanna River Basin were dated by tracing them downstream to the coastal plain and correlating them with Cenozoic sediments (Pazzaglia and Gardner, 1994). The highest terrace, dated as mid-Miocene, indicates an uplift of 130 m in the Great Valley, giving an uplift rate, if constant, of 9 mm ka^{-1}. Other terraces confirmed uplift rates as high as 10 mm ka^{-1} (Pazzaglia and Gardner, 1993).

What is not well known is the uplift in the Valley and Ridge and in the Appalachian Plateaus. There was less unloading of Paleozoic sediment on the plateaus, but some uplift is expected.

Conceptual Schemes

If the Appalachians (or any other contemporary landscape) have been subject to erosion since the early Mesozoic, any reasonable continuous denudation rate would have planed the land surface down to sea level. There must have been uplift to provide fresh rock for attack by the erosive processes. The key question, and a question that has not been satisfactorily answered, is what is the time-dependence of the uplift. There are two points of view that define the opposite ends of the uplift scale. These may be called the Davisian model and the Hackian model.

In his famous interpretation of the rivers and valleys of Pennsylvania in 1889, William Morris Davis proposed that regional uplift was episodic (Davis, 1889). There were periods of rapid uplift interspersed with long periods of, at most, minor uplift. The landscape was planated during the quiescent periods. These planated surfaces were then dissected by rapidly down-cutting streams during the succeeding episodes of rapid uplift. In the Appalachians, one product was the Schooley Peneplain, the remnants of which are the (roughly) accordant summits of the ridges of the Valley and Ridge. Another product was the Harrisburg Peneplain, which seems coincident with many of the limestone valley floors.

The opposite concept is that the rate of regional uplift is essentially constant. Therefore, denudation is also essentially constant, except that the rate of denudation varies widely with rock type. The landscape, therefore, is simply the product of differential erosion. Sandstones and quartzites, being highly resistant, form the ridge tops, while limestones and shales, being less resistant, form the valleys. This is the concept of dynamic equilibrium. Erosion is balanced against uplift, and the form of the landscape does not dramatically change. The concept goes back at least to G.K. Gilbert, but the name most commonly credited with fleshing out the idea is that of John T. Hack (Hack, 1960). The concepts are illustrated schematically in Figure 2. As end-members, both have their problems. Their application to karst topography introduces some additional problems.

THE EVOLUTION AND DEVELOPMENT OF THE APPALACHIAN KARST

Most attempts to interpret the evolution of Appalachian landscapes have been top-down. Terraces, terrace gravels, filled sinkholes, and related features are given estimated dates and then fitted into the scheme of landscape evolution. The interpretation offered here is bottom-up. We begin with the existing landscape, and then, using the established rates of the various processes, work backward to see how parts of the landscape fit together. There are some horrendous assumptions, the most important being that rates operating today are adequate to evaluate what has happened in the past. Some important features are ignored, such as the wildly fluctuating climate during the Pleistocene and the corresponding dramatic changes in sea level. These are what might be called back-of-the-envelope calculations, but some of the derived conclusions are remarkably consistent. They are also in disagreement with some previous interpretations by an order of magnitude or more.

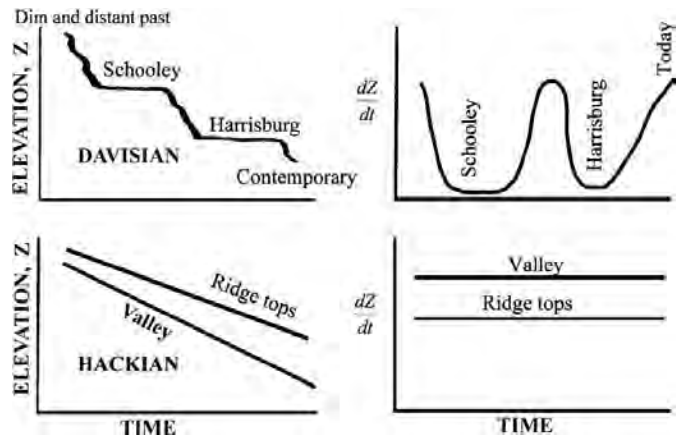

Figure 2. Sketch comparing the classic Davisian concept of landform development with the Hackian concept.

ANCHOR POINTS

Much of the previous interpretation of Appalachian topography, particularly the erosion surfaces, has been based on evidence derived from residual deposits and from river terraces. These are important pieces of the puzzle that must be fitted into their proper places, but the chronology of such features is imprecise. Age-dating of caves has become an important way of interpreting landscape evolution (Atkinson and Rowe, 1992). A much more precise chronology is provided by the recently introduced techniques of cosmogenic isotope dating, especially as applied to clastic sediments in caves.

There has been a dramatic reversal in the role of caves in geomorphic interpretation. The Bretz-Davis view was that caves are deep-seated, random objects, re-excavated by recent streams after the dissection of peneplains, and with no relationship to contemporary topography. Next came the realization that caves, for the most part, are formed as part of contemporary drainage systems and that large, dry passages relate to terrace levels in nearby river valleys. If so, the age of the cave can be estimated from the age of the terrace. With the introduction of cosmogenic isotope dating (Granger and Muzikar, 2001), the caves can be used to provide high-precision dates for the terraces. Infilling of cave passages with clastic sediments is one of the last events in cave development. The burial date of quartz sand and pebbles from these sediments can be determined, a date that is assumed to be the age of the cave and the time at which the cave discharged into a surface stream at base level.

The few cosmogenic isotope dates for Appalachian caves are from the work of Darryl Granger and his colleagues (Granger et al., 1997; Anthony and Granger, 2004, 2006). These dates agree well with the results of back-calculations from denudation and river down-cutting rates and they also serve to anchor those calculations.

THE HARRISBURG SURFACE

Much of an earlier paper (White and White, 1991) was focused on the Harrisburg Peneplain. There does indeed appear to be a well-developed surface that can be traced throughout the Appalachians. Near Harrisburg, Pennsylvania, the type locality, the Harrisburg Surface is the upland level of the Great Valley at 150 meters, now somewhat dissected by surface streams. Along the Juniata River northwest of Harrisburg, the surface is represented by accordant hill summits at 200 meters, mostly on shale, that truncate the local geologic structure. Still farther northwest, the surface appears in the broad interfluve area of the Nittany Valley as a rolling limestone upland at an elevation of 360 meters. The valley uplands of the Shenandoah Valley are at an elevation of 150 meters where the Shenandoah Valley merges into the Potomac Valley, but rise to the southwest, reaching 450 meters at the drainage divide. In Burnsville Cove, west-central Virginia, the Harrisburg Surface is represented by a highly karstic drainage divide and corresponding ridge tops at 760 meters. The surface appears as accordant hill tops in the Swago Creek Basin in the upper Greenbrier Valley (750 m). The Little Levels (730 m) and the Great Savannah (700 m), both sinkhole plains, in the lower Greenbrier Valley, West Virginia, also correspond to the Harrisburg Surface. The Highland Rim of the western Cumberland Plateau is usually considered equivalent to the Harrisburg Surface.

The Harrisburg Surface is clearly not a peneplain in the Davisian sense of the word. It is a surface representing development of wide valleys during a period of stable base level. The surface slopes toward major surface drainages of the Susquehanna, the Potomac, the James, the New, and the Cumberland Rivers.

Using an argument based on residual soils and carbonate denudation in the Nittany Valley of Pennsylvania, Parizek and White (1985) deduced that the dissection of the Harrisburg surface began about 3 Ma ago. A much better anchor point was provided by Anthony and Granger (2004). According to cosmogenic isotope ages of sediments in Big Bone Cave on the Cumberland Plateau, the cave was at grade with the Highland Rim surface at 5.7 Ma. Dissection of the Highland Rim began 3.5 Ma ago, suggesting that the earlier estimate based on denudation rates is not out of line. The secondary valleys, stream networks, and caves below the Harrisburg Surface have developed in the last 3 to 5 million years. Many Appalachian caves, therefore, have ages ranging from mid-Pliocene to relatively recent.

The age of the Harrisburg Surface is a different question. The data cited above show that the dissection of the surface began 3 to 5 Ma ago. The argument has been that the surface could have been in existence as a low-relief, wide valley bottom for much longer. In 5 Ma, chemical denudation would have lowered the Harrisburg Surface by 150 meters. In the Great Valley, the uplift was estimated to be 40 to 50 meters (130 meters in 15 Ma according to Pazzaglia and Gardner, 1994). The net change in elevation of the Harrisburg Surface in the Great Valley is about 100 meters since dissection began.

The classic interpretation of erosion surfaces is that there is a pause in uplift rates. Stream gradients decrease and valleys widen until there is achieved a low-relief valley floor containing a meandering stream of little erosive power. Such a topography could remain stable for long periods of time until uplift was renewed and gradients restored. However, most of the expressions of the Harrisburg surface are karst surfaces. Chemical denudation depends on precipitation, on soil CO_2 (in turn dependent on vegetative cover), and weakly on temperature. Chemical denudation does not depend on gradient as long as the base of the epikarst is above the water table. Although there might be a pause in stream erosion because of decreased gradients, chemical denudation would continue. The low-relief karst surfaces continue to lower, but without dissection.

THE SCHOOLEY SURFACE

An interesting and enigmatic case is that of the mountain/plateau surface that may or may not represent the Schooley Peneplain. While many of the remnants of the Harrisburg surface are karst plains, the remnants of the Schooley surface are resistant quartzites (Valley and Ridge ridges) and conglomerates (Cumberland Plateau). Erosion rates are in the range of 3 to 5 mm ka^{-1}, so that the denudation of the ridge tops is much smaller. During the 3 to 5 million years since the onset of dissection of the Harrisburg Surface, the lowering of the ridge tops would have been no more than 15 to 25 meters.

If the missing carbonate rocks from the carbonate valleys of the Valley and Ridge are back-calculated, the more rapid denudation rate of the carbonates compared with the quartzites of the ridge tops means that the valleys will fill. On the Cumberland plateau, the limestones are relatively thin, so that back-calculating the missing carbonates will intersect the clastic rocks that cap the plateau.

Calculations based on 50 m of residual soil on the Cambrian Gatesburg Dolomite on the crest of the Nittany Anticlinorium in central Pennsylvania concluded that 425 meters of carbonate rock had been removed in order to produce the soil (Parizek and White, 1985). Using the reference denudation rate of 30 mm ka^{-1}, the denudation of the valley center extends back at least 14 million years, to the mid-Miocene. If this column of dissolved limestone is placed in the context of the present Nittany Valley with its bounding Appalachian ridges, the column extends well above the ridge tops (Fig. 3). The valley floor, the Harrisburg surface, is at an elevation of 360 m. The carbonate surface at the beginning of recorded denudation would have been at 785 m. The quartzite crests of the present ridges are at 690 m. Scaling of the elevations would

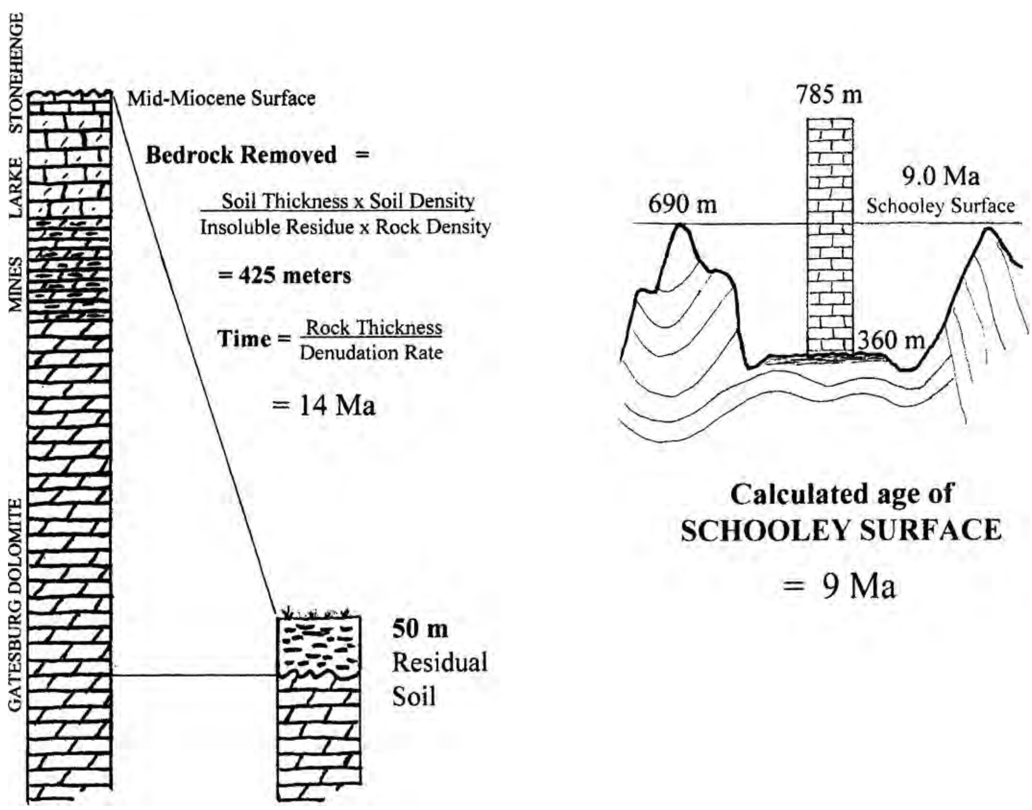

Figure 3. Carbonate rock denudation in the Nittany Valley, Pennsylvania. Estimation of thickness of removed carbonate rock from residual soil taken from Parizek and White (1985).

give a calculated age for the Schooley surface of 9 Ma. Allowing for some denudation of the quartzite would add 30 to 50 meters to the elevation of the ridge tops and thus extend the age to about 10 Ma. Not taken into account is the unknown rate of uplift of the Valley and Ridge.

Another estimate comes from the East Fork of the Obey River on the western margin of the Cumberland Plateau in north-central Tennessee (Fig. 4). The anchor point here is the cosmogenic isotope date for sediments in the upper levels of Xanadu Cave (Anthony and Granger, 2004). This date, 1.64 Ma, and the elevation of the cave above the river give a downcutting rate of 30 mm ka^{-1}. Assuming that this rate has remained constant, on average and extrapolating to the top of the Cumberland Plateau, gives a date of 9.35 Ma for the time that the capping conglomerate was breached at this point in the Obey River Gorge.

The Obey River Gorge has cut about 100 meters below the present-day Highland Rim, which is at an elevation of about 300 m. Using the Big Bone Cave date, the ancestral highland rim would be at an elevation of 450 m, which is the top of the limestone if the karst denudation rate has been maintained. Extrapolating farther back would give the age of the breaching of the plateau as 9.5 Ma.

Cave Mountain Cave, Pendleton County, West Virginia (Dasher, 2001) has the appearance of an old spring mouth. It is located on the crest of the Cave Mountain Anticline, 275 meters above the North Fork River. The crest of Cave Mountain, just above the cave, would also correspond to a remnant of the Schooley surface. Taking a downcutting rate for the North Fork similar to those shown in Table 1, extrapolating to the top of the Smoke Hole Gorge would give an age for Cave Mountain Cave of 9.2 Ma. This would make Cave Mountain Cave one of the oldest caves in the Appalachians, but to the writer's knowledge, no dates have been obtained. Cave Mountain Cave should have functioned as a spring on the bank of the ancestral North Fork when it was just beginning to dissect the Schooley surface. The actual breaching of the surface would have been a bit earlier, perhaps 10 Ma.

Three independent locations give the same 9 to 10 Ma age for the breaching of the Schooley surface. They do not give the age of the surface itself. If, indeed, one can speak of a Schooley surface, it, like the Harrisburg surface, sloped upward toward the drainage divides. Elevations in the Allegheny Mountains of West Virginia are much higher, as are the ridges of the Valley and Ridge, than the corresponding features in Pennsylvania. It appears that the dissection dated to 9 to 10 Ma marks the beginning of present-day topography, rather than the rapid uplift of a low-lying Schooley Peneplain.

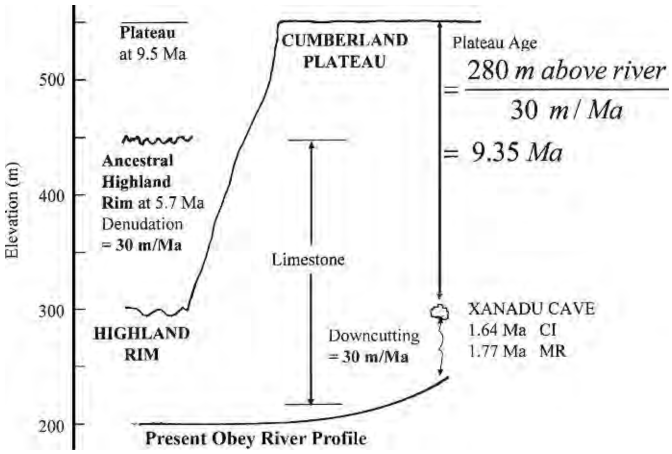

Figure 4. Sketch of Cumberland Plateau showing two estimates of the time of breaching of the caprock.

DEVELOPMENT OF THE APPALACHIAN KARST: LATE MIOCENE TO PRESENT

The hypothesis that Appalachian topography had evolved close to its present form by the early Cenozoic is not consistent with observations of present denudation rates unless long intervals of greatly reduced denudation are inserted. The oldest topographic features that can be linked to present-day topography are the plateau surfaces, especially of the Cumberland Plateau, and the ridge tops of the Valley and Ridge. Although these ridge tops and plateau surfaces have been labeled as the Schooley surface, there is no certainty that it is the Schooley Peneplain as visualized by the early geomorphologists. The dissection of the Schooley surface can be traced back to the Mid-Miocene. Certain features, such as Spruce Knob in West Virginia, with an elevation of 1480 m, and the Cumberland and Crab Orchard Mountains in Tennessee and southwestern Virginia may be remnants from a still earlier time.

During the 5-Ma interval following the initial dissection of the Schooley Surface, there must have been sufficient erosion and denudation to form the Harrisburg surface. The karst denudation data place a severe constraint on the Harrisburg/Highland Rim surface. Because the best development of the Harrisburg surface is represented by carbonate rocks, these will have undergone continuous chemical denudation. It is not appropriate to consider the Harrisburg surface as representing a fixed elevation.

Downcutting of surface streams below the Harrisburg level provided the gradients for the development of large cave systems, particularly in the Greenbrier Valley and along the deep coves of the dissected Cumberland Plateau. Most presently accessible caves range in age from Pliocene to Recent. Most pre-Harrisburg caves have been eroded away with some exceptions of caves in the high ridges, such as Cave Mountain Cave in West Virginia.

The existence of karst surfaces combined with the existence of large master trunk conduits is evidence for a neo-Davisian concept for Appalachian geomorphology. Neither uplift nor downcutting rates appear to have been constant. However, the karst surfaces are lowering continuously, and in this sense, differ from the original peneplain concept.

To end on a note of warning: The foregoing discussion and interpretation should be taken for what it is, back-of-the-envelope number juggling. The hard data are sparse. More cave-sediment dates and more detailed denudation and river down-cutting measurements would certainly help. Other assumptions, such as equating the age of the clastic sediments to the age of the caves and their associated base levels, need more checking. At present, however, the conclusion remains. Present-day Appalachian topography, and certainly the karst topography, can be traced back only to the mid to late Miocene. The shape of the topography at the end of the Cretaceous and its evolution to the mid-Miocene remains lost in the shadows of time.

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B. Schwartz and W. Orndorff – Hydrogeology of the Mississippian scarp-slope karst system, Powell Mountain, Virginia. *Journal of Cave and Karst Studies,* v. 71, no. 3, p. 168–179.

HYDROGEOLOGY OF THE MISSISSIPPIAN SCARP-SLOPE KARST SYSTEM, POWELL MOUNTAIN, VIRGINIA

Benjamin Schwartz[1] and William Orndorff[2]

Abstract: Mississippian carbonates on scarp-slopes of Powell Valley show few surficial karst features, yet host extensive caves (e.g., Omega, Hairy Hole, Rocky Hollow, and Gap Caves) and complex karst hydrogeologic systems. On the limbs of the Powell Valley Anticline, strata dip moderately to steeply into the mountainside, with passage development and flow dominantly along the strike toward water gaps, nickpoints, or structures such as fold axes or faults. Most significant cave development is in the Greenbrier Limestone, which is underlain by Price-Maccrady Formation siliciclastics and overlain by shales, siltstones, and minor limestones of the Bluefield Formation (including the approximately 13-m Little Lime, approximately 100 m above the Greenbrier Limestone). The South Fork of the Powell River, flowing northwest through Powell Mountain at Crackers Neck water gap, defines local base level in the area of recent hydrogeologic studies. Dye traces northeast of Crackers Neck revealed that allogenic recharge sinks into the Little Lime limestone layer and flows southwest beneath the river, resurging on the southwestern bank at the Little Lime Spring. High-flow conditions overwhelm the input capacity of the Little Lime outcrops, and water continues down-slope to sink in the Greenbrier Limestone, then flows southwest along the strike through dominantly vadose cave passages in Omega Cave to the Omega Spring on the northeast side of the Powell River. The stream in Omega Cave is undersized, suggesting that most passage enlargement occurs during high-flow events. Inflows in the upper Greenbrier Limestone near the Crackers Neck water gap drain to a spring on the opposite side of the Powell River. Northeast of the Omega basin, flow is to the northwest, resurging at the nose of the Powell Valley anticline. Springs on the southwest bank of the Powell River receive flow from karstic drainage to both the northeast and southwest, as well as from the river itself. At Powell River Spring, river water includes upstream discharge from Little Lime Spring. This situation resulted in confusing dye-recovery patterns before Little Lime Spring was discovered.

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Introduction

The Mississippian carbonates exposed in southwestern Virginia have long been known to contain significant caves (Douglas, 1964; Holsinger, 1975). Well-known examples include the Cudjos-Cumberland Gap Cave System in Lee County and many other caves along the Powell Mountain and Stone Mountain escarpments in Lee, Scott and Wise Counties. Several of these were extensively mined for saltpetre (Douglas, 1964; Faust, 1964; Holsinger, 1975). Except for interest in the saltpetre caves and intermittent periods of exploration in a few new caves, there has been little systematic or prolonged exploration and study of caves in the region, and even less scientific study of the hydrogeology of karst systems developed in the Mississippian karst of this region. Reasons for this range from the relative inaccessibility of the karst exposures high on steep mountainsides to attention being focused, instead, on other well-known karst regions in Virginia.

However, the early 1990s brought renewed interest in this area when exploration in caves near East Stone Gap in Wise County revealed that many known caves were incompletely mapped and that many unknown caves existed. By the mid-1990s, a historically known blowing pit had been pushed beyond a blocked passage at the base of the entrance shaft to reveal the first pieces of a large and extensive cave system consisting of active and paleo passages. Named the Omega Cave System, it is now both the longest (40.5 km) and deepest (385 m) cave system in the state, and new passages are still being discovered.

Coinciding with the initial exploration and mapping of the Omega Cave System and other caves in the area, was the initiation of hydrogeologic studies in the Mississippian carbonate scarp-slope karst system that contains the cave system. These studies are the first to establish the hydrogeologic significance of the Mississippian scarp-slope system and to determine the relationship between karst systems developed in the Greenbrier Limestone and a stratigraphically higher, but thin, limestone known locally

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as the Little Lime, previously assumed to be hydrologically insignificant.

Scarp-Slope Karstification

In this paper, we present a conceptual model and definition of a scarp-slope karst system. This style of karst has largely been ignored or unrecognized in the literature. We believe it is sufficiently different from other styles of karst that it deserves a new descriptive classification. In brief, a scarp-slope karst system develops in soluble rocks dipping into the mountain side, where the region of major conduit development has occurred internal to the scarp face and may be entirely covered or protected by an overlying mountaintop composed of erosionally resistant, ridge-forming lithologies such as sandstone. Regional structural controls halt down-dip development near the regional base level, and water then flows in a master conduit roughly along the strike toward either springs at water gaps or major fractures that transport water across the strike. In the Mississippian-age carbonates of the Powell Valley, springs are always along the strike because of the insoluble and impermeable nature of the underlying formations. In rocks in other areas, such as the Cambro-Ordovician-age carbonates, springs can be either along the strike or along fractures at high angles to the strike, if underlying formations have appropriate permeability/solubility characteristics. These Cambro-Ordovician systems, while they are similar to the scarp-slope karst described here, will not be discussed further here and will instead be presented in a separate manuscript.

Study Area

Stratigraphically, Mississippian carbonates in Wise, Scott, and Lee Counties in southwest Virginia are composed of the >130-m-thick Greenbrier Limestone (also known as the Newman Limestone and locally known as the Big Lime) and the approximately 13-m-thick Little Lime in the overlying Bluefield Formation (Henika, 1988). Below the Greenbrier Limestone, lies impermeable shale and sandstone of the locally undivided Mississippian Price-Maccrady Formations. Separating the two carbonates are approximately 60 m of thinly bedded shales and thin siltstones, sandstones, and mudstones of the Bluefield Formation. Above the Little Lime is an additional 75 m of Bluefield lithologies, capped at the mountain ridgelines by approximately 50 m of the erosionally resistant Stony Gap Sandstone member of the Mississippian-age Hinton Formation. The Stony Gap member is a cliff former and creates low but prominent cliff lines along many of the ridge-tops.

Structurally, the carbonates are found exposed on steep scarp-slopes along both limbs of the Powell Valley Anticline. The Virginia portion extends nearly 100 km from near Norton, where the northeastern end plunges and forms the head of the Powell Valley, to Cumberland Gap at the far southwestern tip of Virginia in Lee County (Commonwealth of Virginia, 2003). The breached and deeply eroded core of the anticline forms Powell Valley, and, incidentally, is the only location in Virginia where Mississippian, Devonian-Silurian, and Cambro-Ordovician carbonates are exposed by the same structure, which continues to the southwest beyond the Virginia-Tennessee border. Most of the eastern limb of the Powell Valley anticline is not preserved due to regional thrust-faulting and erosion. However, in southern Wise County and western Scott County, the eastern limb remains and is known as Powell Mountain. Powell Mountain contains dramatic outcrops of the Greenbrier Limestone in the form of cliffs several kilometers in length, up to 80-m-high, and 300 to 500 m above the valley floor. The regional dip of the western limb is 30 to 60 degrees to the northwest, while dip angles on the eastern limb are a shallower 5 to 20 degrees to the southeast. It is in this eastern limb near the town of East Stone Gap that the Omega Cave System has formed in the Greenbrier Limestone. On scarp-slopes, outcrops of the Little Lime are commonly covered by Bluefield and Stony Gap colluvium and can be difficult to identify. In steep hollows and water gaps (near Crackers Neck, for example), the Little Lime does form short cliff lines, though overall surface expression is considerably less than the Greenbrier Limestone.

While the exposures of the carbonate units discussed here are regionally extensive, our research to date has been focused on a detailed understanding the hydrogeology of the eastern limb of the Powell Valley Anticline, or the Powell Mountain block, roughly between the towns of Norton and Duffield, Virginia (Fig. 1). Structurally and stratigraphically, our study area is similar to the extensive western limb of the Powell Valley Anticline, with the main difference being the steeper dip of identical strata on the western limb.

Hydrogeology

Regional and local geologic structures and stratigraphy have controlled the development of all major caves known in these scarp-slope Mississippian carbonates. Regionally, cave systems are formed along fractures sub-parallel to the strike on the limbs of the Powell Valley Anticline. Because strata dip into the mountainsides and the base of the Greenbrier Limestone rests on the insoluble and impermeable Price-Maccrady Formation, water perches on the insoluble strata and is forced to follow an along-strike flowpath to the most efficient discharge point. Locally, water gaps and deeply incised hollows or valleys formed by structural flexures and fracture or minor fault zones perpendicular to the strike control the hydrogeology. In turn, these features provide the discharge points for the regional conduit flow through the carbonates and contain all major springs within the Greenbrier Limestone.

On the surface, there is relatively little indication that a regionally extensive active karst system exists beneath these

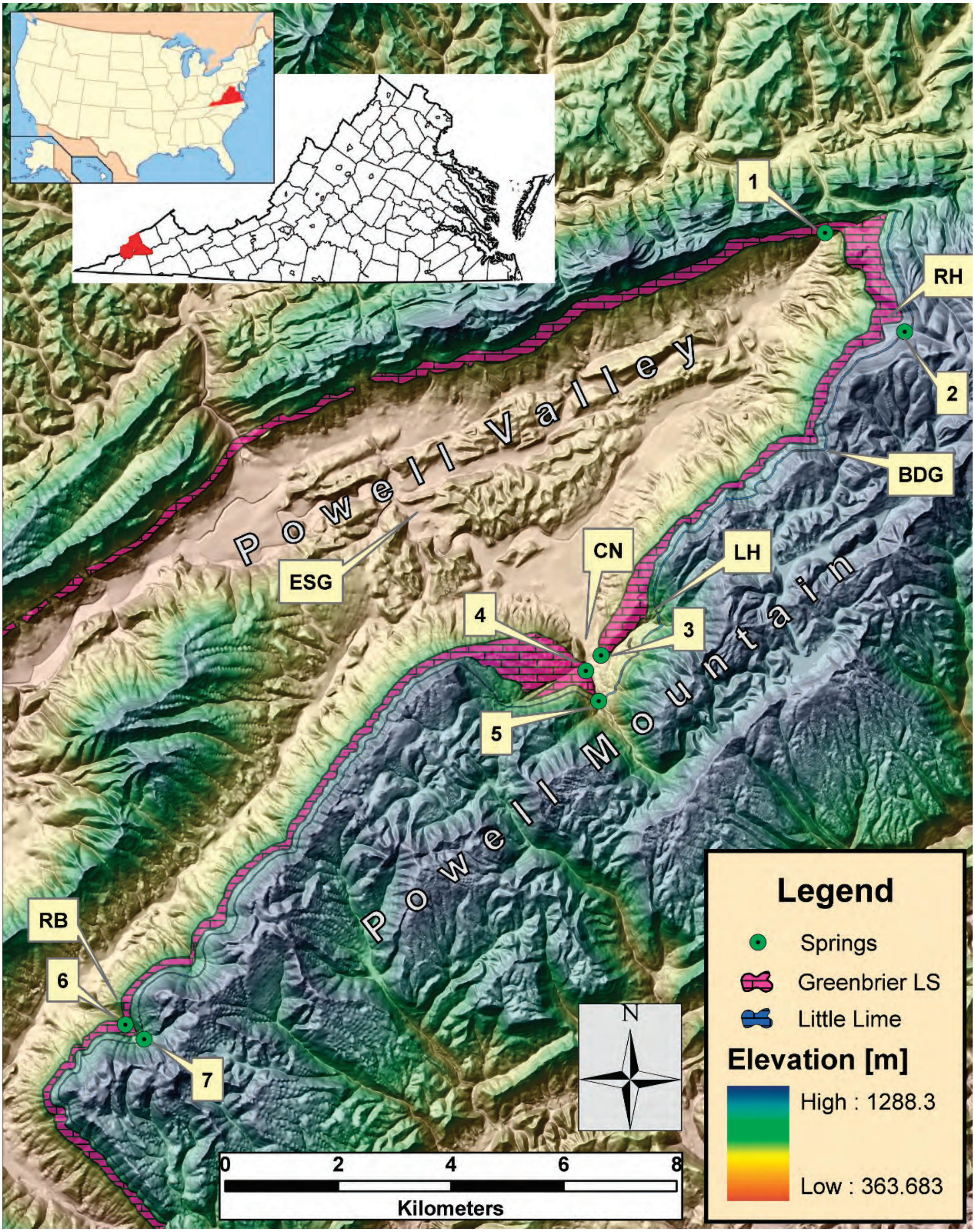

Figure 1. Location, topography (USGS, 2008), Greenbrier Limestone outcrop (USFS, 2008) and Little Lime outcrop of the northeastern Powell Valley and Powell Mountain. Springs are: 1) Bloomer Spring, 2) unnamed Little Lime spring in Rocky Hollow, 3) Omega Spring, 4) Powell River Spring, 5) Little Lime Spring, 6) unnamed Greenbrier spring in Roaring Branch, 7) un-named Little Lime Spring in Roaring Branch. Locations mentioned in text are: CN = Crackers Neck, ESG = East Stone Gap, BDG = Beaverdam Gap, RH = Rocky Hollow, LH = Long Hollow and RB = Roaring Branch. The South Fork of the Powell River flows to the northwest between 3) and 4) and through the community of Crackers Neck. Two minor gaps between Crackers Neck and Beaverdam Gap are (from north to south) locally known as Sheep Gap and Maple Gap.

impressive scarp-slope exposures. With few exceptions, development of karst features such as large sinkholes and sinking streams is very limited, though they are quite common in other nearby carbonates. Due to Greenbrier Limestone outcrops being high on steep scarp-slopes below narrow ridgelines, allogenic contributing areas are narrow (< 1 km wide) and long (> 10 km). With few sink-points or perennial surface streams present above the upper Greenbrier Limestone contact, it might seem that runoff from the mountainsides would simply cascade over the steep cliff exposures without a chance to enter the subsurface. However, except for during large storms, the opposite is true, and much of the overland flow from higher on the mountain actually does enter well-developed karst systems through small recharge features that are frequently buried beneath sediment and colluvium.

In contrast to the few, small karst features on the surface, the Omega Cave System contains an impressively complex and well-developed network of active and fossil conduits. Sequentially abandoned conduits indicate that this system has existed with similar hydrologic inputs for an unknown but extended period of time. Of the four major infeeding streams that join the master stream trunk, three are related to the three known entrances, Blowing, Lori Cori Canyon Cave, and Stingweed. The fourth probably drains Maple Gap, though the source has yet to be determined. Infeeders begin as complex networks of coalescing small tributaries in passages perched on thin, resistant beds within the uppermost 50 to 60 m of the Greenbrier. These infeeders then descend rapidly to the base of the limestone via shaft complexes and short sections of meandering canyons and crawls. The most important shaft complexes have formed along fracture zones that are related to small erosional hollows or surface gaps along Powell Mountain — Sheep Gap, Beaverdam Gap and Rocky Hollow, for example. Ancient infeeder complexes produced now-abandoned passages that were also hydrologically related to gaps on the surface.

Soon after the discovery of the master conduit in the Omega System, dye-trace studies were initiated to characterize the current hydrologic system. Using standard fluorescent dye-tracing techniques, studies began in 1997 and continued until 2005. As part of the initial study, a spring inventory was performed and several springs issuing from the Greenbrier Limestone were located.

Springs in the Greenbrier Limestone

Of the Greenbrier Limestone springs identified (Fig. 1), the Omega Spring is the smallest and appears to be undersized relative to both the internal passage sizes and the large allogenic drainage area that was initially associated with the cave system. With an estimated mean flow between 0.01 and 0.05 $m^3\ s^{-1}$, it discharges from boulders near the base of the limestone at the toe of a ridge on the northeastern side of Crackers Neck.

Powell River Spring rises from a water-filled conduit near the base of the limestone on the southwestern side of the South Fork of the Powell River in Crackers Neck. Compared to Omega Spring, this spring is quite large. However, much of its flow is derived from the South Fork of the Powell River itself. During normal summer flow conditions, the entire river sinks where it crosses the upper contact with the limestone, resulting in approximately 500 m of dry riverbed between the inflow and Powell River Spring. Estimated mean flow is between 0.1 and 0.3 $m^3\ s^{-1}$.

Bloomer Spring has an estimated mean flow of between 0.1 and 0.3 $m^3\ s^{-1}$ and discharges from a fault-controlled cave entrance near the base of the limestone at the northeastern end of Powell Valley where the axis of the Powell Valley Anticline plunges to the northeast. An unnamed spring in the Greenbrier Limestone was also located in Roaring Branch several kilometers to the southeast of Crackers Neck.

Springs in the Little Lime

A second spring inventory later identified several other springs that discharge from the Little Lime (Fig. 1). Little Lime Spring was found along the Powell River upstream from the sink point of the Powell River as it crosses the upper contact of the Greenbrier Limestone. As with the Powell River Spring, Little Lime Spring's discharge also appears to be composed primarily of flow from the Powell River, as it partially sinks in its bed approximately 50 m upstream from this small cave-entrance resurgence. Mean discharge is estimated to be between 0.1 and 0.3 $m^3\ s^{-1}$.

Two other unnamed springs in the Little Lime are relatively large (between 0.1 and 0.3 $m^3\ s^{-1}$) and, unlike the Greenbrier Springs, discharge at high elevations in major hollows (Rocky Hollow and Roaring Branch) from perched karst-aquifer systems on the eastern limb of the Powell Valley Anticline. Discharge from both of these springs flows a short distance across the intervening Bluefield Formation before sinking at the upper contact of the Greenbrier Limestone.

Dye Trace Results and Discussion

Omega System and Associated Little Lime

Fluorescein dye was released in the main stream in the Omega Cave System near the downstream end of the system (Fig. 2). This dye was detected only at Omega Spring. Although the exact travel time is unknown (< 2 weeks between exchange of charcoal traps), the dye probably took less than 24 hours to travel the approximately 1 km between the release point and the spring. This trace proved what had been assumed when the stream was discovered: that the stream observed in the cave discharges from the Omega Spring down-gradient and along the strike.

In an effort to delineate the upstream end of the hydrologic system, a second tracer was injected on the

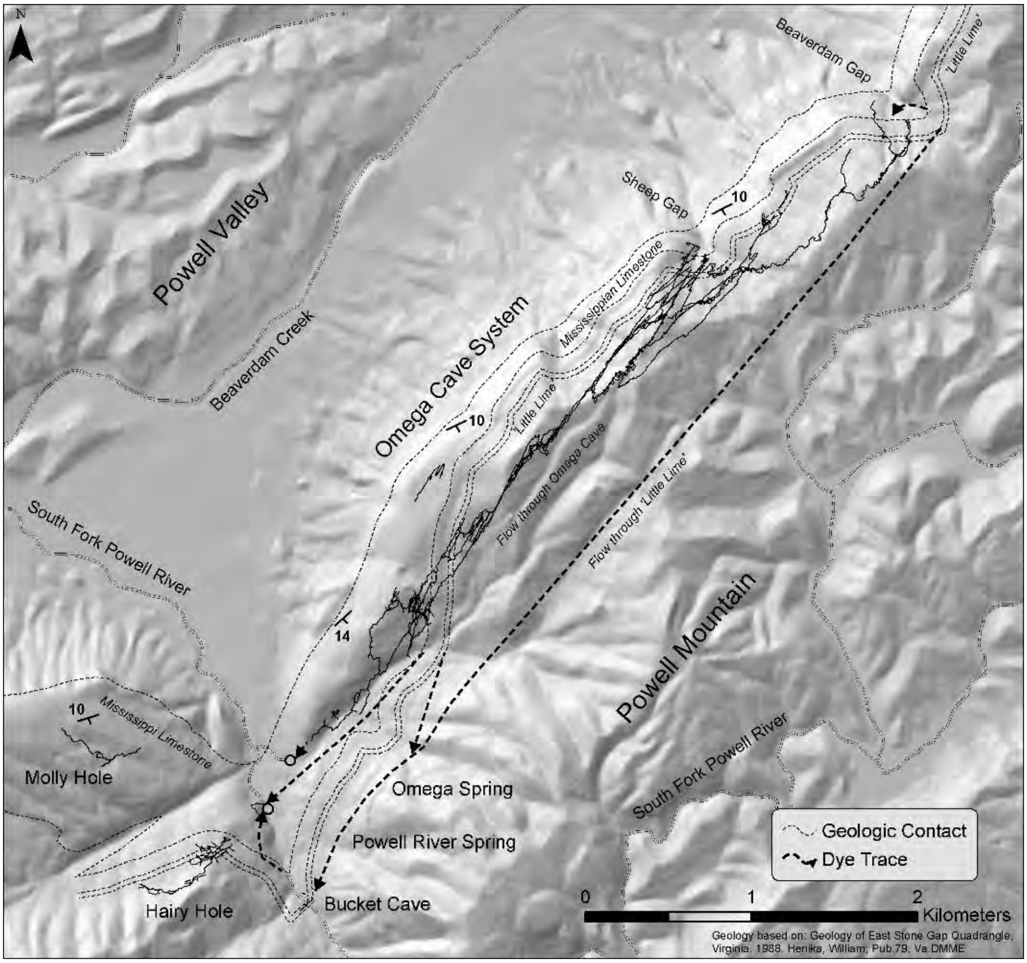

Figure 2. Portion of Figure 1 showing details of the Omega Cave System and other nearby caves. Known flow-paths are noted and dye-trace vectors are shown as dashed lines between injection points and detection points. Approximate surface exposures of the Greenbrier Limestone and Little Lime are shown in dashed outline.

surface in Beaverdam Gap, beyond the northeastern extent of the cave system at that time. Here, a very small perennial surface stream was presumed to sink through gravel and into the Greenbrier limestone. Surprisingly, dye from this trace was detected in very low concentrations at the Omega Spring and much higher concentrations at Powell River Spring.

Because this trace occurred prior to the discovery of the Little Lime springs, only the Greenbrier Springs had been monitored. To explain the unexpected results, two hypotheses were put forth: 1) Two parallel conduit systems exist in the Greenbrier Limestone, one leading to the Omega Spring and the second leading to Powell River Spring. 2) A significant conduit system exists in the Little Lime that is capable of transmitting water to a previously undetected spring up-river from the known Powell River sink point. To test these possibilities, more field work was performed, and Little Lime Spring was found. Merely the existence of this spring proved that some sort of a karst system could develop in the Little Lime, but without knowing the proportion of cave-derived vs. river-derived water, its extent and significance were difficult to predict. With the newly discovered spring being monitored, the Beaverdam Gap trace was repeated. Most of the dye from the repeat trace was detected at Little Lime Spring, where it then flowed down the South Fork of the Powell River to the Greenbrier sink point and flowed underground to reach the Powell River Spring, the location where most dye had been detected during the previous trace. During the repeat trace, a small amount of dye was again detected at the Omega Spring.

The reason both traces resulted in a low dye concentration at the Omega Spring is apparently related to overland flow after small thunderstorms and the fact that dye was injected upstream of the Little Lime contact rather than just above the Greenbrier Limestone. Just after the injection, most dye quickly entered the Little Lime system. Small storms mobilized some of the remaining dye and transported it down-gradient to the upper contact with the Greenbrier Limestone. The limited capacity of the Little Lime to receive water by infiltration through streambed sediments here is easily overwhelmed, and some runoff from storms will flow overland until it sinks at the upper contact with the Greenbrier Limestone and travels through the Omega System. In essence, both dye traces proved two important points: a significant karst system does exist in the Little Lime, and the hydrologic extent of the Omega System extends at least as far as Beaverdam Gap. If the dye had been injected at the Greenbrier Limestone contact as was initially intended, the Little Lime might still be unrecognized as hosting a significant karst system. Since these traces, two caves and a promising dig site have been discovered in the Little Lime. The dig and one of these caves have impressive airflow, indicating that significant air-filled passages do exist in the Little Lime.

In Crackers Neck, a small community near the downstream end of the Omega System, the karst hydrology is more complex. This is the result of the two parallel but hydrologically separate carbonate units, as well as the along-strike dissection of surface drainage on the Greenbrier Limestone outcrop in Long Hollow. Long Hollow extends to the northeast from the Omega Spring and is deeply incised into the upper portions of the Greenbrier Limestone.

Dye injected in a small tributary stream in Long Hollow at the upper contact with the Little Lime proved that during normal summer flow conditions most allogenic water draining from the mountainside is captured by the Little Lime karst system and flows to the Little Lime Spring on the southwestern side of Powell River. During higher flow conditions at this site, some surface water crosses the Little Lime and flows down to the upper contact of the Greenbrier Limestone. In a subsequent trace in the same tributary, dye injected at the upper Greenbrier Limestone contact was recovered at the Powell River Spring, also on the southwest side of Powell River. This trace proved the existence of an adjacent drainage basin in the Greenbrier Limestone near the downstream end of the Omega System. This had been hypothesized because of the fact that much larger amounts of water were observed sinking here than were discharging from the Omega Spring and the fact that no related infeeding streams are observed in the downstream section of the Omega System. Interestingly, water sinking in Long Hollow discharges on the opposite side of the Powell River, proving once again that surface streams often do not represent hydrologic boundaries in karst systems.

Between the sink and rise of the South Fork of the Powell River, additional water is added via conduits draining adjacent basins. The Long Hollow drainage basin has been proven by dye trace to join this underground segment of the river. Additional flow is also derived from Hairy Hole Cave (currently approximately 3 km) on the southwest side of the Powell River. A passage in the cave can be followed until it joins a short air-filled segment of passage containing the underground Powell River. Diving has also proven the connection between Powell River Spring and Hairy Hole Cave. According to survey data and personal communication with diver Ron Simmons, all the underwater passage surveyed in Powell River Spring has developed at depths of between 0 and 10 m.

Well-developed karstic flow systems in the Little Lime on both sides of Powell River drain to the Little Lime Spring on the southwest side of the river. On the northwest side of the Powell River, water that sinks as far to the northeast as Beaverdam Gap has been traced to this spring. Dye travel times (< 2 wk to travel 6 km) and several blowholes in the Little Lime suggest that a well-developed conduit system has developed in this unit. On the southwest side of Powell River, no dye traces have proven the existence of an extensive karst system in the Little Lime. However, Bucket Cave, which is the only significant cave currently known in the Little Lime (approximately 300 m in length), extends to the southwest from near Little Lime Spring and acts as a flood-water overflow route for water that is presumably flowing from the southwest toward the spring. The existence of a flood-water overflow route is further evidence that the Little Lime system is well-developed and captures large amounts of allogenic water at certain times.

Rocky Hollow

Rocky Hollow lies adjacent to and northeast of Beaverdam Gap and the Omega System drainage. The largest known cave in this area is Rocky Hollow Cave (1.7 km), which is largely an inactive fossil remnant of a much older system. Dye traces in a sinking stream in Rocky Hollow and in Rocky Hollow Cave have proven that water that sinks in the Greenbrier limestone flows to Bloomer Spring at the head of the Powell Valley. Interestingly, much of the water that sinks into the Greenbrier in Rocky Hollow has been discharged from a perched Little Lime spring higher in this deeply incised hollow. Due to the absence of surface streams or other recharge features where dye can be injected, the exact location of the drainage divide between the Omega Spring/Powell River Spring and the Bloomer Spring drainage basins has not yet been determined. And there is no reason to assume that the boundary for the overlying Little Lime system will correspond to that of the Greenbrier.

Roaring Branch

Preliminary field observations indicate that the relationships between the Little Lime and Greenbrier appear to be similar in both Rocky Hollow and Roaring Branch. Although no dye-traces have been performed in Roaring Branch, a large perched Little Lime spring discharges high in the hollow. This water then flows down to the Greenbrier contact, where it sinks and likely contributes to discharge from the un-named spring near the base of the Greenbrier limestone in Roaring Branch.

Speleogenesis of the Omega System

The extensive network of passages explored and mapped in the Omega Cave System (currently 40.5 km) provides an

Figure 3. An upper-level passage in which the upper half has formed in a massively bedded maroon red-bed and the lower half has formed in a massively bedded purple red-bed. Note the approximately 60-cm-thick limestone bed separating the two lithologies. Also note the meandering ceiling channel that formed under phreatic conditions. The lower half of the passage has been so severely modified by spalling breakdown that it is impossible to tell how it formed, though there appears to be a narrow vadose canyon below that is nearly filled with breakdown debris.

opportunity to observe features and processes that are related to current and past hydrologic conditions. An observation that was initially quite puzzling is that the active master conduit appears to be significantly larger than would be expected to develop from the very small stream running through it. This was partially resolved when the importance of the Little Lime and its role in capturing allogenic recharge before it reached the Greenbrier Limestone was recognized and understood. However, this only explained the small size of the stream and not the oversized passage. The oversized passage appears to be the result of large amounts of chemically and physically aggressive flood waters periodically pulsing through the system. Observational evidence indicates that flow through the master conduit varies by perhaps as much as four orders of magnitude. Low-flow during dry summer conditions has been measured at approximately 0.0015 $m^3\ s^{-1}$, while flood-water discharge likely reaches 1.5 $m^3\ s^{-1}$ or more. This flood discharge estimate is based on high-water marks after rain events.

Passage development in the Omega Cave System is controlled by a combination of structural features (primarily joints) and lithologic variations. In the upper half of the limestone, several massive to shaly-bedded argillaceous carbonate red-beds influenced the vertical development of multiple stacked fossil passages. The influence of these beds on passage development varies significantly depending on location. In some passages, a red-bed unit acted as an aquitard that perched the passage for long distances, while other passages developed entirely within the red-bed (Fig. 3). Reasons for this are not yet fully understood. In the downstream half of the known cave, the main stream passage is resting directly on the underlying Price-Maccrady Formation. In the upstream half of the system, the active stream is perched on sequentially higher red-beds, with approximately horizontal segments of canyon passage separated by waterfalls. Waterfalls are the locations of active nick-point migration as the more dissolutionally-resistant red-beds dissolve more slowly than the purer limestones between them.

The dominant passage morphology in the system is vadose canyons formed along a joint set approximately 15° south of the southwest strike direction. Many portions of the cave also show evidence of shallow phreatic conditions and vadose-modified phreatic passage (Figs. 4 and 5). These areas are generally associated with certain combinations of structural or stratigraphic controls that created localized shallow-phreatic conditions, or are found at the ceiling level in multi-level canyon complexes where the uppermost passage began as a water-filled conduit at or near the local water table. There is little evidence for deep-seated phreatic development in the Omega Cave System, and the system's evolution generally follows a model of sequential abandon-

Figure 4. The ceiling level of a passage in Omega that initially formed under phreatic conditions (note large scallops on the ceiling), then transitioned into a partially water-filled passage (wall notches mark the level of slow-moving pooled water), and finally turned into a low-gradient, meandering vadose canyon that is still active today (the canyon portion is under the right wall in this view and drops approximately 30 m).

Figure 5. Abandoned vadose canyon passage typical of the Omega Cave System. Passage height in the photo is approximately 15 m.

ment and down-cutting toward a regional base-level that has been lowered over time by landscape evolution and erosion. However, some of the fossil or relict caves (including many of the saltpeter caves, for example) preserve features and morphologies indicative of phreatic conditions and flow. At least one cave, nearby Parsons Cave, may be a fossil resurgence, based on its morphology as a single large tube ascending obliquely updip near, but over 200 m above, the modern water gap at Crackers Neck.

Paleo-Flow and the Age of Caves in the System

Although research is currently underway to more completely understand the speleogenetic history of the area, the paleo-hydrologic conditions and landscape associated with the relict caves, and how the relict and active caves are related, we can reach some general conclusions about flow and landscape evolution based on our current knowledge and understanding of the system.

With the exception of Parsons Cave (Fig. 6), all observed paleo-flow directions in the Omega System are similar to the present flow direction, i.e., generally to the southeast. While it is very close to, and even overlaps passages in Omega, there is currently no evidence indicating that Parsons Cave has any hydrologic or genetic relationship to the modern Omega System. In fact, ceiling features in Parsons Cave indicate that flow was in the opposite direction, to the north-northeast, and the cave likely represents a paleo-resurgence that discharged water from a much older karst system that is now largely eroded away. Regardless of the flow direction, Parsons Cave was formed under phreatic conditions (Fig. 7), and this fact allows us to make a simple calculation as to the approximate minimum age of karstification in the region.

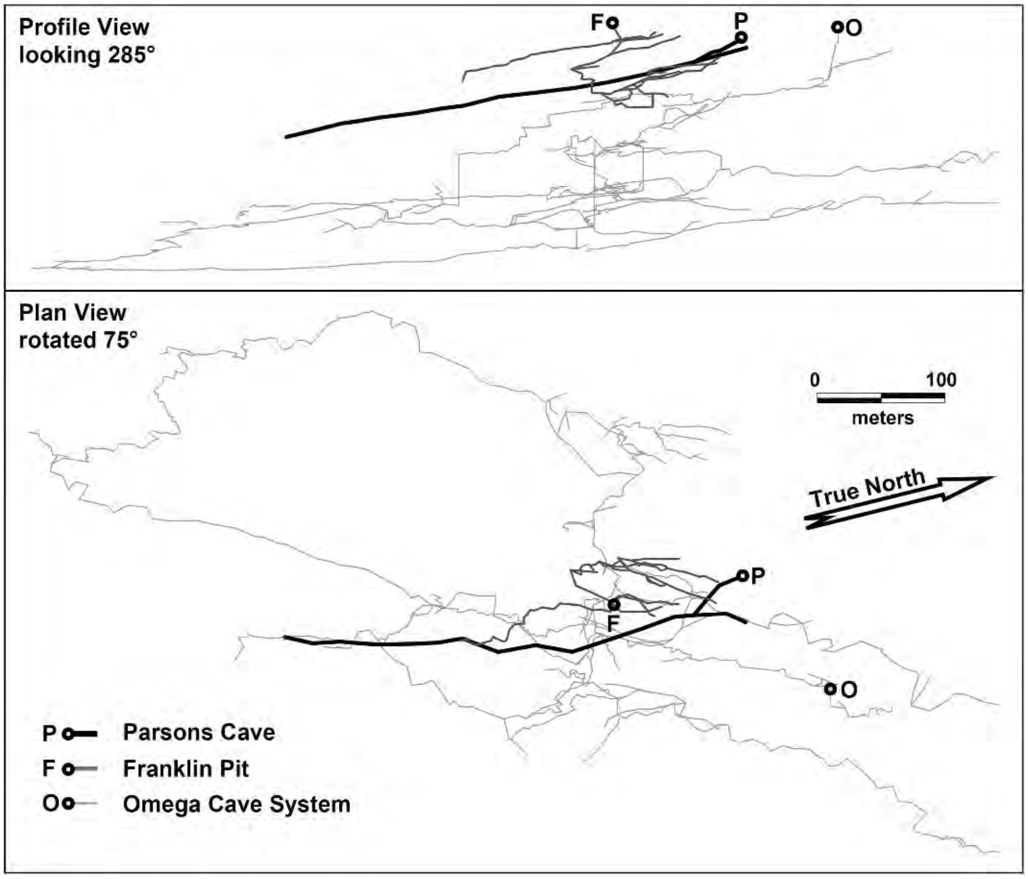

Figure 6. The relationship between Parsons Cave (single bold black survey), Franklin Pit (medium width gray survey) and a portion of the underlying Omega Cave (thin gray survey) near the downstream end of the system (see Figure 2 for complete line plot). Scale is identical in plan and profile views, with no vertical exaggeration. Note that the three caves do not connect.

The entrance to Parsons Cave is currently 200 m higher than the Omega Spring and opens into a single descending trunk passage 400 m in length and 90-m-deep that terminates at a flowstone pinch. If we assume a regional incision rate of approximately 30 m Ma^{-1} (Granger et al., 1997; Ward et al., 2005) and assume that the land surface around the cave entrance has not been eroded (only for the purposes of calculating a minimum age), this means that the entrance of Parsons Cave is evidence of a well-developed, but water-filled, karst system that existed between 6 and 7 Ma. Omega Spring and Powell River Spring are both at similar elevations of approximately 525 m asl, which is approximately 30 m higher than the average elevation of the river bottom in the center of Powell Valley. Unless the stream channel has changed its profile configuration considerably, it is also reasonable to assume that the center of Powell Valley has also been eroded by a minimum of 200 m since the time when Parsons Cave was still in phreatic conditions.

If Parsons Cave actually is a paleo-resurgence, this may indicate a difference between the modern and ancient systems: deeper circulation of water in the past than in the present. There is no evidence of deep phreatic flow in any of the paleo passages currently known in Omega Cave. In fact, all evidence points toward short segments of shallow-phreatic or near-water-table flow that later transitioned into vadose conditions. Deep circulation today is limited to moving farther under the mountain in the down-dip direction by insoluble rock beneath the limestone. This could imply that rates of incision at the surface have varied considerably over time and Parsons Cave represents development during a relatively stable period when deep flowpaths had time to develop. Or Parsons Cave could be the remains of a system with an entirely different hydrologic function than hypothesized here.

Near Parsons Cave lies Franklin Pit (Fig. 6), yet another relict cave. While it appears to be significantly older than the Omega System, there is also no known relationship with Parsons Cave. At this time, very little is understood about the formation and significance of Franklin Pit with respect to speleogenesis in the area, and more work is being done to understand this.

Although more work is needed before the evolution of karst in this system can be considered well understood, none of our observations conflict with the general scarp-slope karstification model that follows.

Figure 7. Passage in Parsons Cave as viewed looking toward the entrance. Note that the passage appears to have formed along the intersection of a joint and a prominent bedding plane. Ceiling features indicative of upward flow are not obvious in this picture. Breakdown has modified the walls in many areas, and the floor has been modified by saltpetre mining. The tan-colored strip down the center of the passage is a thin coating of modern flowstone.

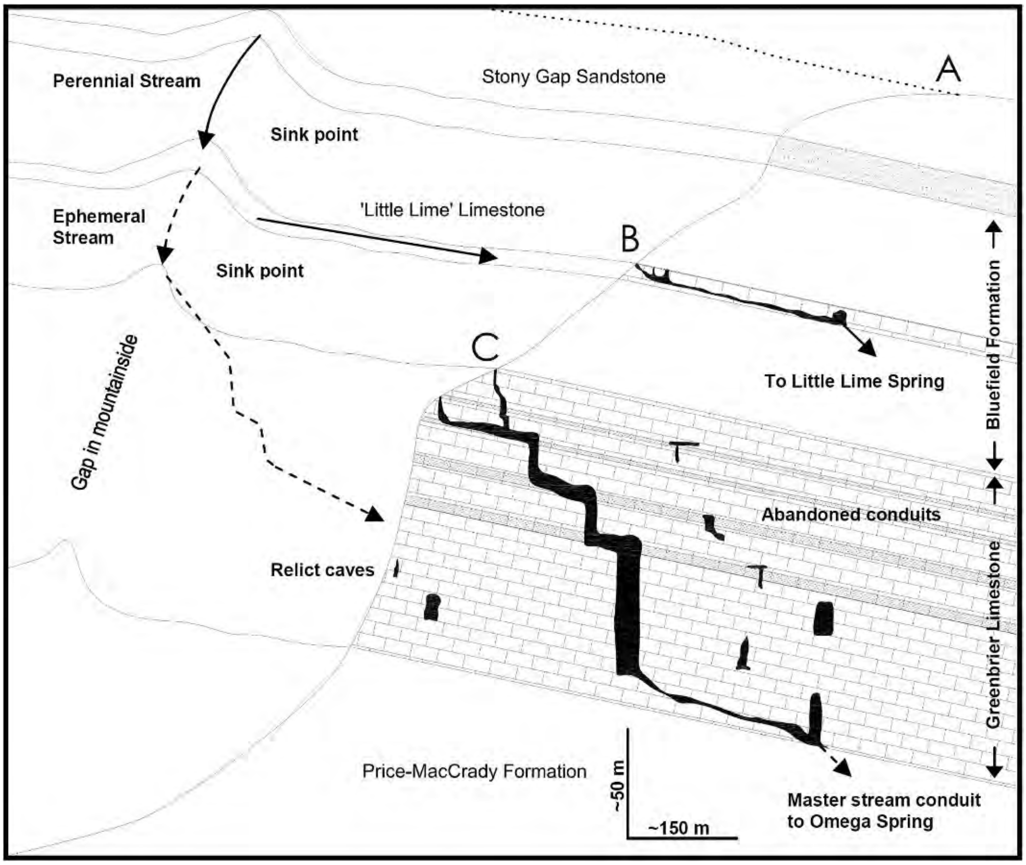

Figure 8. Cut-away conceptual model of Mississippian scarp-slope karst system that generally reflects conditions found near the Omega Cave System A) is the crest of Powell Mountain B) marks a hypothetical entrance passage in the Little Lime that leads down-dip to the Little Lime master stream conduit. Note that the Little Lime system has not been entered by cavers, while the Greenbrier system has C) is a simplified representation of an entrance to the Omega Cave System and a passage that leads to the Greenbrier master stream trunk. Note that the location and thickness of red-beds in the upper half of the Greenbrier are for schematic purposes only and are not precisely represented. Vertical exaggeration is approximately 3:1.

CONCEPTUAL MODEL OF MISSISSIPPIAN SCARP-SLOPE KARST HYDROLOGY

This research resulted in the development of a conceptual model for a scarp-slope karst system (Fig. 8). From a hydrogeologic perspective, the most important components of the general scarp-slope model are: 1) Soluble rocks are exposed on a scarp-slope below a ridge of resistant ridge-forming insoluble rocks. 2) Soluble rocks dip into the overlying mountainside. 3) Allogenic recharge flows off the insoluble ridge and sinks at or near the upper limestone contact before flowing generally down-dip toward a strike-oriented main stream passage deep within and behind the scarp-slope. 4) Water flows along the strike toward a spring in a water gap or deep hollow in the ridge, or toward a major fracture zone where flow crosses the structure and underlying rocks of lower solubility (dolostones, for example) to discharge at a valley-bottom spring. 5) A significant to dominant portion of the water flowing through the system is allogenically derived. 6) Major cave streams are largely undersaturated with respect to calcite, and thus capable of significant dissolution. 7) Steep slopes above the limestone contact result in extremely flashy systems capable of significant physical weathering by abrasive clasts in the sediment load.

The general model can be refined and specifically applied to the Central Appalachian Mississippian scarp-slope system by adding the following essential elements: 1) The stratigraphically higher Little Lime karst system captures most allogenic recharge during normal hydrologic conditions. 2) Excess surface drainage or discharge from high-elevation Little Lime springs crosses the intervening Bluefield Formation and enters the Greenbrier Limestone. 3) The base of the karst system in the Greenbrier Limestone is defined by the contact with the underlying Price-Maccrady Formation, which forces all Central Appalachian Mississippian scarp-slope springs to be along-strike. 4) Significant across-dip flow only occurs in major water gaps where sinking streams enter at the upper contact and discharge from a spring at the lower contact along the same stream channel.

Both the Little Lime and the Greenbrier Limestone systems direct subsurface flow long distances along the strike via well-developed conduit systems toward a regional discharge point formed at structurally controlled water gaps or deeply incised hollows. Both systems can be thought of as gutters at different elevations on the scarp-slopes, with varying capacities for allogenic recharge. Based on discharge, high-elevation Little Lime springs likely have extensive conduit development associated with them, though there is currently little direct evidence, such as mapped cave passages.

With sediment cover and narrow exposures limiting rapid recharge via sinking streams or sinkholes into the Little Lime, the Greenbrier Limestone system is activated during high-flow events that flush chemically and physically aggressive waters through the system. This results in an over-sized conduit. When large or intense storms occur, significant amounts of water bypass both the Little Lime and Greenbrier Limestone and reach the floor of Powell Valley as runoff. No perennial streams cross the scarp-slope carbonate exposures and reach the valley bottom.

Passage morphologies in the Omega Cave System suggest that most development currently occurs during large floods, with major inputs occurring primarily in only a few poorly developed gaps or hollows. This is supported by observations in the cave, where three of the four feeder streams to the main stream trunk are associated with an obvious gap on the surface.

Conclusions

This first thorough hydrogeologic investigation of a Central Appalachian Mississippian scarp-slope karst system has shown that both the Greenbrier Limestone and the thin Little Lime can form well-developed karst systems. In all cases documented, water discharging from the Little Lime almost immediately sinks in the Greenbrier Limestone just down-slope and contributes significantly to discharge from certain Greenbrier Limestone springs. Omega Spring, which receives no hydrologic input from a Little Lime spring, appears undersized in relation to the amount of allogenic recharge available from the slope above the Greenbrier Limestone. Instead, recharge into the Little Lime overlying the Omega Cave System is directed to the Little Lime Spring, where it then contributes to discharge from Powell River Spring. During normal flow conditions, the Little Lime captures nearly all allogenic recharge from higher scarp-slopes. During storm events, the low capacity of the Little Lime recharge zones is easily overwhelmed, and excess flow will recharge the Greenbrier Limestone system directly, as well as contribute to surface runoff into Powell Valley.

Conduit development in the Omega Cave System is dominated by sequential down-cutting and abandonment over time, resulting in a complex network of stacked strike-oriented passages. Modern enlargement of the active master stream passage appears to be dominated by chemically and physically aggressive flood water that enters via narrow canyon passages connected by active shafts. During normal flow conditions or summer drought, discharge from the Omega Spring nearly stops. This is due to a combination of factors, including a narrow surface exposure of the Greenbrier Limestone and the overlying Little Lime system capturing the majority of allogenic recharge during periods with little precipitation.

This research has resulted in a general conceptual model of Mississippian scarp-slope karst hydrogeology that provides a framework for future research in the area. We have also proposed that scarp-slope karst systems develop in a unique manner with unique properties and should be recognized as a category of karst that is common in the folded and faulted sedimentary rocks of the Appalachian Mountains.

Acknowledgments

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CAVE EXPLORATION AS A GUIDE TO GEOLOGIC RESEARCH IN THE APPALACHIANS

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Abstract: Cave exploration and mapping can provide considerable insight into the nature of groundwater flow and geologic processes in soluble rocks. The Appalachian Mountains provide an ideal setting for this exchange of information because their geology varies greatly over short distances. Caves reveal the way in which groundwater flow is guided by geologic structure, and they help to clarify aquifer test data, well yield, and contaminant dispersion. Well tests in karst aquifers often reveal confined or unconfined conditions that make little sense stratigraphically, but which can be explained with the aid of cave mapping. With regard to geologic mapping, many caves reveal structures that are not visible at the surface. Caves also show evidence for underground geochemical processes that cannot be detected from well data. Subtle mineralogical clues are generally erased by weathering and erosion at the surface, but persist in many caves. The information that caves have provided about subsurface geology and water flow is now being used by explorers, even those with no geologic background, to help them find new caves.

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Introduction

A century ago, not much was known about Appalachian caves and their patterns, and the few available cave maps revealed little about the local geology or regional context. As exploration accelerated in the mid-1900s, a wide variety of cave maps became available, and the relation between caves and their surroundings became clearer. In many of the studies from this period, the main application of cave data was its integration with surface geomorphology (see Kambesis, 2007; and White, this volume). In this paper, I follow a different direction to show how cave data can apply to processes that are only indirectly related to the surface. The topics of special interest here include groundwater hydrology, subsurface geologic structure, and geochemistry. Only solution caves are considered here, because they provide the greatest amount of information about water flow and its control by geologic structure.

Appalachian Geology

The Appalachians provide a world-class example of continental-edge tectonic deformation. Their special virtue for cave studies is that they provide a broad variety of geology, landscapes, and cave types. From southeast to northwest, the rocks grade from igneous and metamorphic to sedimentary and from complex to rather simple structure. The region consists of several physiographic regions (Fig. 1): from east to west they are (1) the Piedmont Province, which consists of igneous and highly deformed metamorphic rocks exposed in low hills; (2) the Blue Ridge Province, which is a highland composed of resistant sedimentary and metamorphic rocks; (3) the Ridge and Valley Province, which consists of folded and faulted sedimentary rocks; and (4) the Appalachian Plateaus Province, a broad, high upland composed of mostly flat-lying sedimentary rocks. In the northeast, the New England Province is a broad mountainous region geologically related to the Piedmont, and the Adirondack Province is a local mountainous uplift of relatively old igneous and metamorphic rocks.

Metamorphic rocks of the Piedmont include marbles of Cambrian-Ordovician age. These were deformed by several phases of Appalachian mountain building throughout much of the Paleozoic Era. In hilly regions, the marbles are exposed in narrow discontinuous bands, but they contain significant caves only in the northeastern equivalent of the Piedmont in western New England and eastern New York. The Adirondack Mountains of northern New York include the ancient, highly crystalline Grenville Marble around their perimeter. The metamorphism took place during the massive Grenville deformation (about 1.2 billion years ago during the Precambrian Era) of even older carbonate strata. The Grenville contains many caves, none very extensive, but some with large passage cross sections where there has been abundant groundwater flow.

The Blue Ridge Province is the western boundary of the Piedmont. It consists of resistant sandstones and metamorphic rocks of Precambrian and lower Cambrian age that stand as a high ridge or upland separating the lowlands of the Piedmont from the lowlands of Cambrian-Ordovician limestones of the Ridge and Valley Province to the west. It widens southward into the Smoky Mountains, which include the highest peaks east of the Mississippi. There are no significant soluble rocks in this province.

The Ridge and Valley Province includes strongly folded limestones and dolomites, interspersed with insoluble

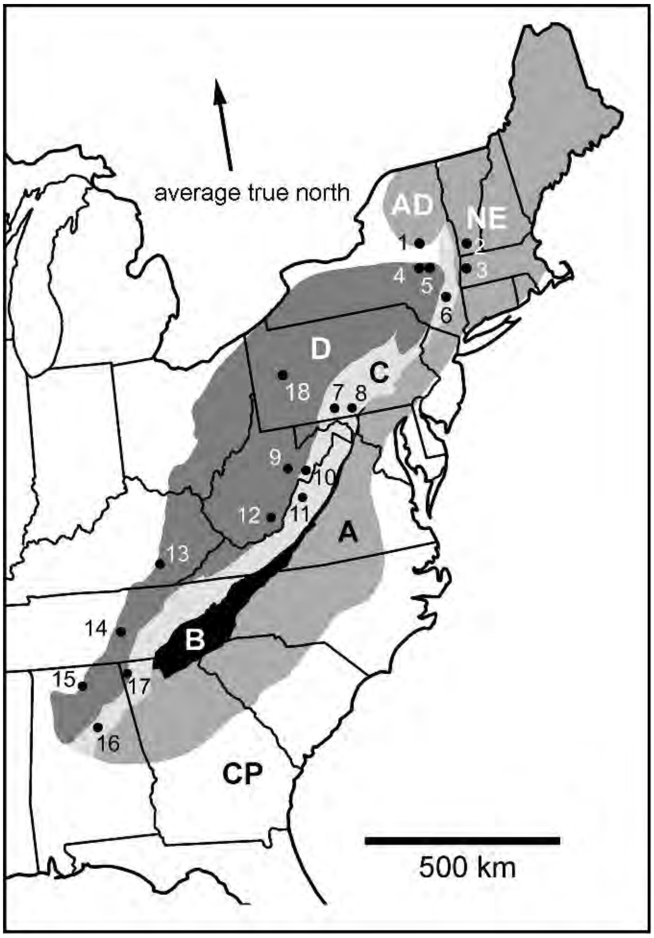

Figure 1. Map of the Appalachian Mountains, showing geomorphic provinces and field sites described in this paper. A = Piedmont Province; B = Blue Ridge Province; C = Ridge and Valley Province; D = Appalachian Plateaus Province; NE = New England Province (continuous with Piedmont Province); AD = Adirondack Province, a window of igneous and metamorphic rocks dating from the Precambrian Grenville Orogeny; CP = Coastal Plain, consisting of gently dipping sedimentary rocks of Cenozoic age. 1 = Carter Ponds Cave, N.Y.; 2 = Morris Cave, Vt., 3 = Convention Cave, Mass.; 4 = caves of Schoharie County, N.Y.; 5 = Knox Cave, N.Y.; 6 = Jack Packer's Cave, N.Y.; 7 = Hipple Cave, Pa.; 8 = Goods Cave, Pa.; 9 = Marshall Cave, W.Va.; 10 = Minor Rexrode Cave, W.Va.; 11 = Butler and Clark's Caves, Va.; 12 = Culverson Creek, Ludington, and Burnside Branch Caves, W.Va.; 13 = Wells Cave, Ky.; 14 = Blue Spring, Camp's Gulf, and Rumbling Falls Caves, Tenn.; 15 = Anvil Cave, Ala.; 16 = Anderson Cave, Ala.; 17 = Ellison Cave, Ga.; 18 = Harlansburg Cave, Pa.

rocks, all of which have been strongly folded and faulted. Differential resistance of the eroded rocks has formed ridges and valleys that are strongly linear and parallel. Carbonate rocks are exposed in valleys and ridge flanks and are highly cavernous. Most of the strata are of early Paleozoic age. A broad valley of thick Cambrian-Ordovician carbonates extends along the eastern edge of the province, but because of the low-relief terrain, most caves are small and scattered.

The Appalachian Plateaus consist of less deformed Paleozoic strata that stand as high, dissected plateaus. Folding and faulting are common along the eastern side and account for a few broad ridges and valleys, but otherwise, the topography is dominated by deep dendritic valleys with intervening uplands. Carbonate rocks range from Cambrian to Mississippian age.

Effect of Geologic Structure on Groundwater Flow Patterns

The way in which water flows through cavernous rock may seem indecipherable from water-well data, but a familiarity with caves can remove much of the confusion. Solution caves form along the paths of maximum groundwater flow, so an accurate cave map shows which of the local recharge patterns and geologic structures have been most important in determining the flow patterns and, therefore, potential routes for contaminant dispersion.

Appalachian cave patterns strongly reflect variations in the geologic structure. Figure 2 shows some representative cave patterns in each province. Comparing caves of vastly different scale can be misleading, but the general tendencies are clear from those shown in the figure.

In the highly deformed marbles of New England and the Adirondacks (Fig. 3), the caves follow irregular paths along several fault and joint systems, sheet fractures parallel to the land surface, and metamorphic foliation. The original stratal dips are indistinct in many areas, and highly varied where detectable. Groundwater moves opportunistically through all of these openings in the direction of the main lines of underground drainage. Small caves of similar, but structurally less complex, patterns are also scattered throughout the other provinces, especially where floodwater has formed or modified them. It is often not possible from the maps alone to make a clear distinction between caves in the northeastern marbles and caves in other provinces.

In the strongly folded Ridge and Valley Province, most caves have greatly elongate patterns that follow the linear trends of the topography, parallel to the strike of the beds (Fig. 4). Examples in Figure 2e–h show that many caves of small to moderate size throughout the full length of the region consist almost entirely of single strike-oriented passages. Many caves, especially large ones such as Figure 2d, are complicated by diversion passages and down-dip tributaries, and yet, the overall strike orientation is still clear. Network caves are also common in this region, mainly in local areas of nearly horizontal, highly jointed beds capped by thin, permeable sandstone.

Most of the Appalachian Plateaus Province contains faintly deformed strata, although folding and faulting are

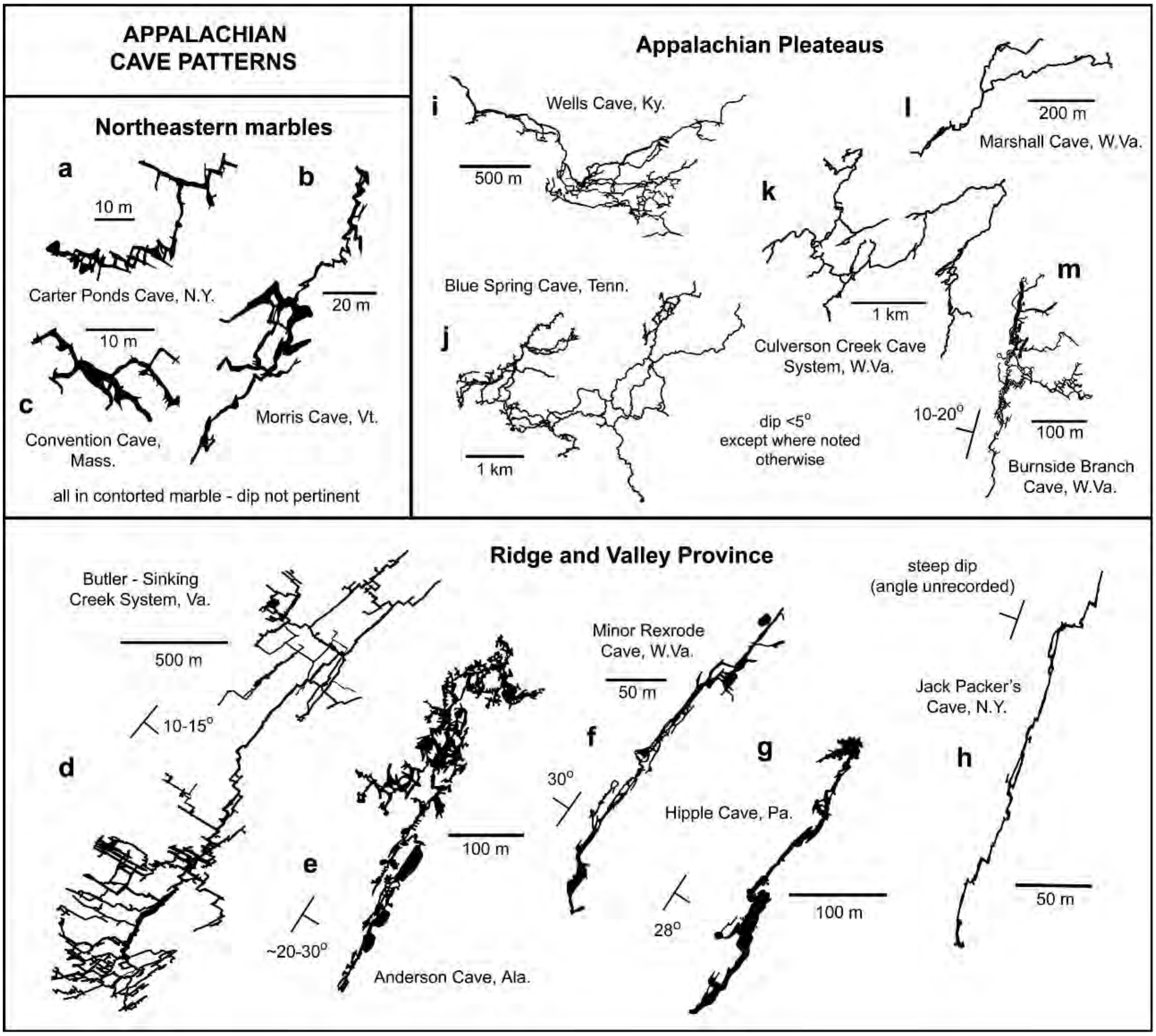

Figure 2. Typical cave patterns in the Appalachians. Maps simplified from the following: Piedmont equivalent (marbles of western New England and foothills of Adirondack Mountains: (a) Porter (2002); (b) P. Quick et al. (Nardacci, 1991); (c) J. Wells et al. (Nardacci, 1991). Ridge and Valley Province: (d) = Butler Cave Conservation Society; (e) E. Breeland et al. (Varnedoe, 1973); (f) W. Davies (1958); (g) B. Smeltzer (Stone, 1953); (h) D. Brison et al. (Nardacci, 1991). Appalachian Plateaus Province: (i) Dayton Area Speleological Society; (j) W. Walter et al. (data from M. Yocum); (k) West Virginia Association for Cave Studies; (l) D. Medville et al. (Dasher, 2000); (m) P. Williams et al. (Haar et al., 1975). See text for discussion of cave patterns.

common along its southeastern border with the Ridge and Valley Province. Mississippian limestones are exposed over broad areas, especially around the margins of the plateaus. Most caves have broad, curving patterns guided mainly by the bedding of the strata (Fig. 2i–l; Fig. 5). Where dips are low, the dip and strike still play an important role in cave patterns, but subtle variations in dip direction make this relationship unclear. In contrast, the cave in Figure 2m is located along the gently folded southeastern edge of the Appalachian Plateaus, where the main passages are tightly aligned along the strike and dip-oriented tributaries join from the east (compare with Fig. 2d). Network caves are also scattered throughout the Appalachian Plateaus; as in the Ridge and Valley Province, they are located in highly jointed, nearly horizontal strata capped by thin sandstone. Maze caves of this sort are described in a later section.

As is typical in many caves throughout the world, passages in Appalachian caves that formed in the vadose zone (above the water table) tend to follow the dip of the rocks, whereas those that formed in the phreatic zone (at or below the water table) are strongly influenced by the strike of the beds. The steeper the dip, the shorter and more linear are the vadose passages, and the longer and more linear are the phreatic passages. From most cave maps it is possible

Figure 3. Contorted marble in an overturned syncline, Eldon's Cave, Massachusetts, in the New England extension of the Piedmont Province.

Figure 4. Opening to a strike-oriented cave along the flank of an anticline, Island Ford Cave, Virginia, in the Ridge and Valley Province. (Cave entrance is the dark triangular area on left.)

to interpret the local rock structure from these criteria alone. In most marble caves, this relationship is insignificant, however, because the original stratal dip is indistinguishable. Large sheet fractures may serve the same function on a local scale.

Although it is easy to overestimate the value of these observations to groundwater science, cave patterns still provide insight that is difficult to obtain in other ways. One who is familiar with them can at least anticipate broad trends in well yield and contaminant dispersion, or find the appropriate words to describe what is observed from well tests.

Directional Permeability (Anisotropy) in Karst Aquifers

In characterizing an aquifer for water supply, one of the first things to consider is whether it behaves in a confined or unconfined manner. In a confined aquifer, the potentiometric surface (level at which water stands in a non-pumping well) lies in low-permeability beds above the aquifer. Water moves parallel to the top and bottom of the aquifer. In an unconfined aquifer, there is no confining bed, and water is free to move vertically, as well as parallel, to the aquifer boundaries. The difference between the two aquifer types, shown in Figure 6, is important in predicting well yield and anticipating contaminant leakage from the surface.

By observing water movement through caves, it is immediately clear that there is a very hazy distinction between the two aquifer types in soluble rock, and apparently most bedrock of any kind. Consider the following examples:

Vertical Anisotropy

In a confined aquifer, the declining water level (drawdown) in a pumping well follows a distinct pattern. After a few minutes of pumping, the drawdown increases linearly with the log of time (see, for example, Fetter, 2001, p. 175). In an unconfined aquifer, the drawdown curve becomes steeper with time and does not plot as a straight line on the graph of drawdown vs. log of time. But in many wells in unconfined soluble rocks, the wells behave as though they were confined. Figure 7 shows an example from a clearly unconfined Silurian-Devonian limestone aquifer, in a well a few hundred meters from McFail's Cave, New York, which lies in the same rock unit. The well terminates at or near the base of the limestone. The straight line on the drawdown graph shows that the flow behaves as though it were in a confined aquifer, i.e., with the flow moving toward the well with no vertical convergence during the period of pumping.

This is unexpected, because many caves in the region are strongly guided by vertical joints, which illustrates the prominent role of fractures discordant to the bedding. But, although there are prominent joints in the uppermost limestone unit (Coeymans Limestone), the cave passages in the underlying Manlius Limestone are sinuous and curvilinear, which shows the strong influence of bedding (Fig. 7). In the Manlius Limestone, this bedding control is interrupted in only a few places by joint-controlled passage segments. The confined behavior of the aquifer, as shown by the well test, is apparently imposed by the prominent bedding. This aquifer is anisotropic; in other words, the horizontal permeability of the Manlius Limestone is greater than the vertical permeability, at least at this location.

Figure 5. Sinuous bedding-controlled passage in Rumbling Falls Cave, Tennessee, in the Appalachian Plateaus Province.

HORIZONTAL ANISOTROPY

In many places in the Appalachians, the elongation of cave patterns suggests that the aquifers are horizontally anisotropic (i.e., that the permeability is greater in one direction than in others). Consider, for example, the elongate patterns of caves in the Ridge and Valley Province (Fig. 2), which suggests that the greatest permeability is in the NE-SW direction. Figure 8 shows an example in south-central Pennsylvania. Like others in the area, it is a maze cave that provides an indication of the relative importance of two sets of fractures. The strike-oriented fractures are dominant. The rose diagram of passage trends vs. length quantifies this relationship.

An extensive well test was performed in a nearby area by a hydrologic consultant, and the resulting cone of depression, drawn by contouring software, is shown in map view in Figure 8. Both the cave and the well are located in steeply dipping, massive carbonates of the Cambrian-Ordovician Beekmantown Group. Relative to the local strike in each area, the elongation of the cone of depression is in the same direction as the rose diagram of the cave. Furthermore, their length-to-width ratios are nearly the same.

The asymmetry of a cone of depression represents an even larger contrast in permeabilities. The permeability contrast (NE-SW vs. NW-SE in this case) equals the square of the ratio of long diameter to short diameter of the cone of depression. In other words, $K_{\max}/K_{\min} = (D_{\max}/D_{\min})^2$, where $K_{\max}$ and $K_{\min}$ are the maximum and minimum permeabilities and $D_{\max}$ and $D_{\min}$ are the maximum and minimum diameters of the cone. In Figure 8, which shows a diameter ratio of about 3.8 to 1, the NE-SW permeability is about 14 times greater than the NW-SE permeability. This has a strong influence on local contaminant velocities. Incidentally, this well showed a time-drawdown curve typical of an unconfined aquifer, which indicates that local fractures allow abundant vertical flow.

The rose diagram of the cave shows a diameter ratio of about 4.3 to 1. If treated in the same way as the cone of depression, it suggests that the NE-SW permeability is about 18 times greater than that in the NW-SE direction. There is no quantitative reason why directional permeability should be revealed exactly by the relative lengths of passages in a cave, but field tests by this author in other geomorphic provinces have verified similar relationships (see also Palmer, 1999, 2007). Anisotropy tests in wells are very scanty, so no other comparisons are possible here. Neither the cave nor the well test are bounded by confining units, and in the massive Beekmantown, the influence of bedding is minor.

Anisotropy is a vital component of any site assessment involving potential contaminants, but detecting it with pumping tests is time consuming and costly, as it requires multiple monitoring wells. Thus, it is often overlooked or assumed not to be present. The use of cave patterns to estimate anisotropy is promising and warrants further investigation.

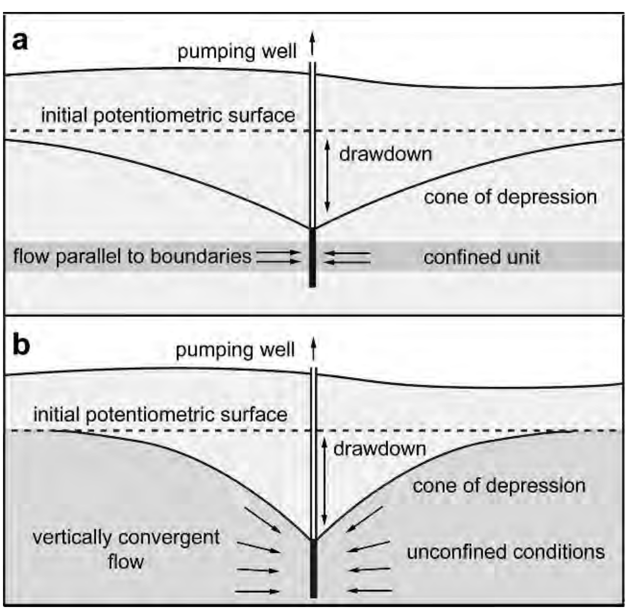

Figure 6. Comparison between confined (a) and unconfined (b) conditions in a pumping well.

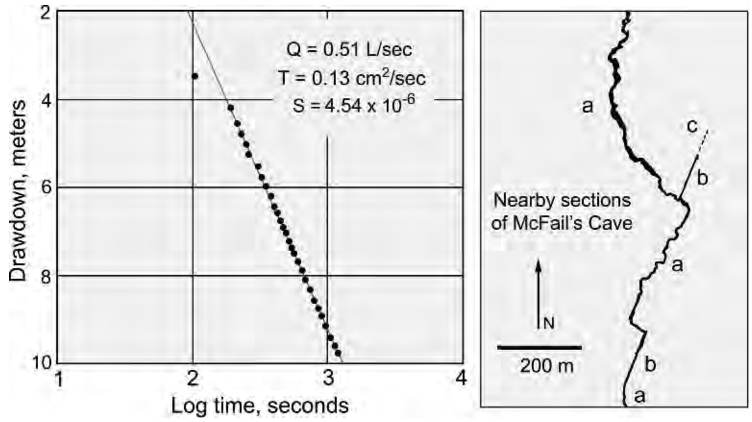

Figure 7. Drawdown graph from pumping of R. Smith well, Schoharie County, N.Y., October 12, 1996. The straight line of the drawdown implies confined aquifer conditions, despite the apparent lack of a confining unit. This shows the vertically constrained nature of water flow in prominently bedded rock. Q = discharge. Transmissivity (T) was calculated from the slope of the line (larger T means higher well yield for a given amount of drawdown). Storativity (S) was calculated from data in an observation well 28.4 m from pumping well (larger S means smaller volume of cone of depression). Both T and S are very low, as is typical of a non-cavernous limestone. On the cave map, a = sinuous bedding-controlled passages in the Manlius Limestone; b = local joint-controlled fissures in the Manlius Limestone; c = joint-controlled fissures in Coeymans Limestone.

Cryptic Geological Structures

Geologic mappers acquire most of their information from surface rock exposures, but exposures are widely scattered in the Appalachians because of the thick soil and dense vegetation. Well logs may help, but they provide only local points of information. Caves provide lengthy and continuous exposures that are well suited for stratigraphic and structural mapping. Some examples are described here.

An Active Fault in Georgia

In 1968, explorers in Ellison's Cave, Georgia, probed into a low, dip-oriented stream passage in the Bangor and Monteagle Limestones (Mississippian age) of Pigeon Mountain, at the eastern edge of the Appalachian Plateaus Province. The passage lies high above the local valleys, partly perched on shale beds. The explorers, led by Richard Schreiber, were approaching the flank of the mountain, but they were still hundreds of meters above the nearby spring outlet. They expected some kind of change in the passage character and were not surprised when the passage floor dropped into a 38-m pit. However, they were stunned when a short distance beyond they found that it plummeted 155 m (510 ft) straight down along a large fault (Schreiber, 1969). They named it Fantastic Pit. Later exploration showed that the total vertical range from the highest ceiling to the floor is about 200 m.

Off the bottom of the pit, nearly all of the passages follow this fault zone. The orientation of wall grooves (slickensides) shows it to be a lateral fault, with mainly horizontal movement. Many of the fault surfaces are recrystallized to thick sheets of calcite or dolomite (Fig. 9). Most unusual are the piles of pulverized rock that have accumulated from grinding along the fault planes (Fig. 10). This fault has been active recently, at least since the cave formed. Some years after the discovery, a mild earthquake occurred with its epicenter in the center of the same mountain (Richard Schreiber, personal communication, ca. 1988). There has been no apparent damage to the cave, because the fault displacements have been small and intermittent. Still, such an intimate view of a semi-active fault is extremely rare near the land surface, especially in the eastern United States.

Structure in Contact Caves of West Virginia

The contact caves of southern West Virginia are located at the base of the Greenbrier Group, mainly limestones, where they overlie the soft, mainly insoluble Maccrady Shale along the eastern edge of the Appalachian Plateaus (Dasher, 2000). Both rock units are of Mississippian age. Passages that follow the contact are common in several long caves, such as The Hole, Ludington, McClung, Maxwelton, and Scott Hollow. Cave streams have cut downward into the shale as much as 12 m in places. The typical contact passage is wide at the top, with a nearly flat ceiling and walls that taper toward each other into a narrow trench (Fig. 11). Any change in dip or displacement along faults is highly visible in the passage ceilings because of the strong contrast in rock types.

At the surface, the low-relief plateaus in which the caves are located show little hint of underlying rock deformation. The caves below display a variety of warps, folds, and faults. For example, in Ludington Cave, many tributaries that enter the main stream passage from the east are developed along the limestone-shale contact. Figure 12 shows the local stratigraphy. An extensive thrust fault cuts

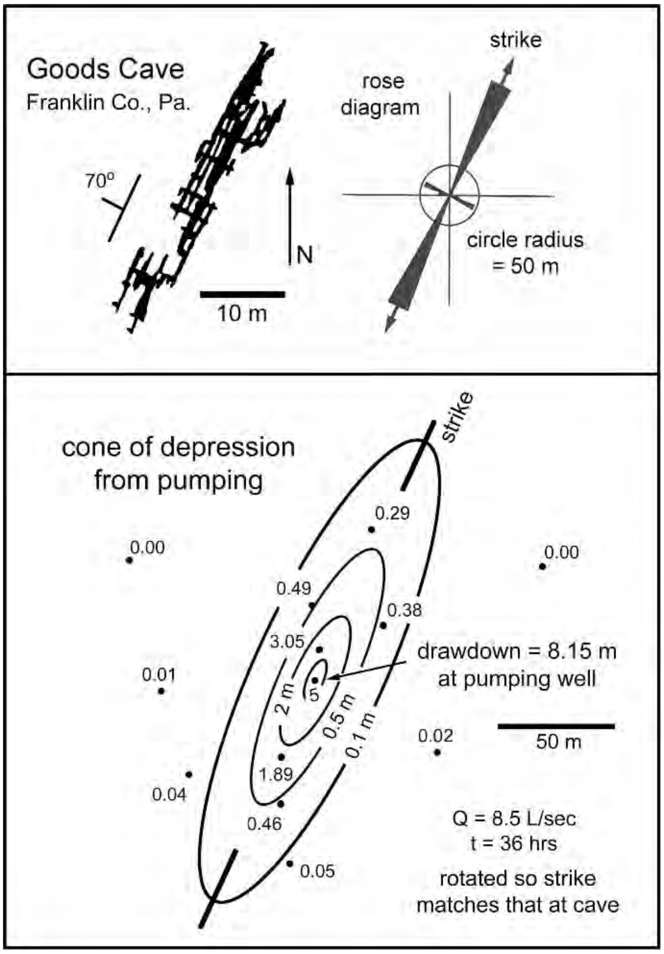

Figure 8. Comparison between the cone of depression in a pumping well in the Beekmantown Group and the rose diagram of passages in a cave in the same strata (Goods Cave, Pa.) in a neighboring county. Cave map by Bernard Smeltzer (from Stone, 1953); well-test data courtesy of John Walker, Doylestown, Pa.

across the beds and follows the contact for a considerable distance (Fig. 13). The underlying beds are shot through with secondary faults, and in places, the beds are dragged upward so that they are vertical or even slightly overturned. The orientations of cave passages in all of the contact caves are adjusted to the structure, with vadose canyons extending down the dipping shale and with local offsets along faults. Few of these structures have been detected at the surface, and their effects on groundwater flow are known only to a handful of speleologists.

KARST GROUNDWATER FLOW ALONG A FAULT IN POORLY SOLUBLE ROCK

Knox Cave, in Albany County, New York, has been well known for centuries. No major discoveries had been made in it since the late 1950s. Except for a few minor trickles and local streams entering during floods, the cave is fairly dry. Digging in 1995–2000 revealed a new crawlway

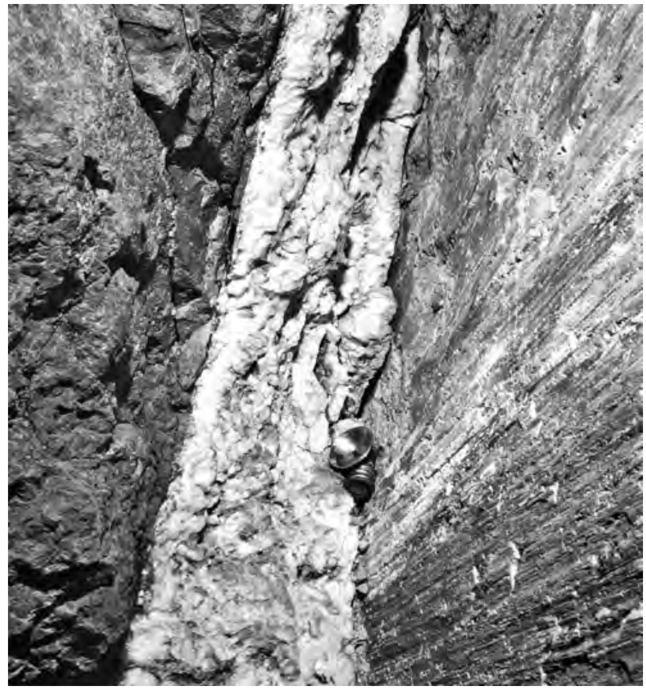

Figure 9. Recrystallized calcite along a fault plane in Ellison Cave, Georgia. Note the slickensides that indicate the direction of fault movement (striations on right). Carbide lamp 12 cm high for scale.

off the far upper end of the cave that eventually dropped into a lower-level stream (McLuckie, 2002). Mapping showed that the new stream passage extended right under many of the long-known passages in the cave. This was quite unexpected, because the formerly known passages are located almost at the bottom of the local Helderberg limestones (Silurian-Devonian). These limestones are underlain by the thin dolomitic shale of the Brayman Formation, and it is in that formation that the stream

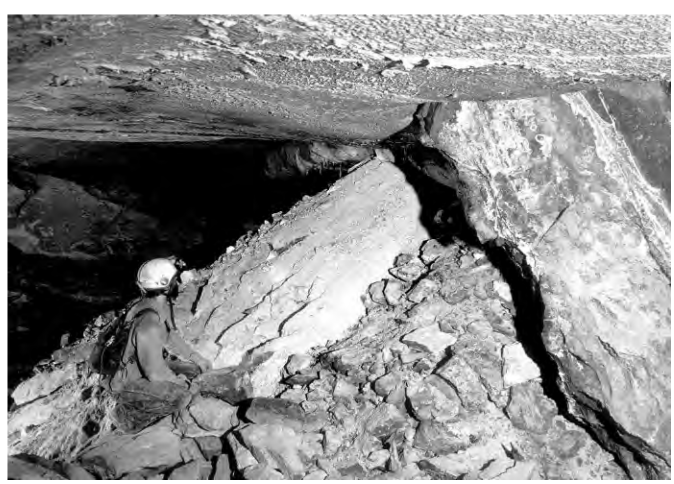

Figure 10. Cone of debris from the active fault zone in Ellison Cave, Georgia.

Figure 11. Example of a "contact cave," Ludington Cave, W.Va. The ceiling is at the base of the Hillsdale Limestone (Greenbrier Group), but the rest of the passage is in the Maccrady Shale.

passage is located. It is the only known cave passage to have formed entirely in the Brayman Formation. A close view shows that the passage formed along a low-angle thrust fault, which allowed enough water to pass through the poorly soluble rock to form a cave passage. The Brayman Formation contains as much as 70% carbonate material in places, but it is soft, crumbly, and generally low in soluble material. The idea that it could be the primary host to an active cave stream had never been considered.

Hidden Geochemical Processes

Camp's Gulf Cave is located in the Cumberland Plateau of eastern Tennessee, near the western edge of the Appalachian Plateaus. It had been long known as a short cave that terminated in breakdown. In the late 1970s, local caver Bill Walter probed through the breakdown and discovered a large stream passage interrupted by a series of huge collapse chambers, three of which were, at that time, the largest cave rooms in the eastern United States. A later discovery in nearby Rumbling Falls Cave (Smith, 2002) contains an even larger room, up to 130 m by 220 m in horizontal extent and 110 m in total height. Figure 14 shows the top of this room, which is entered by a 62-m breakdown shaft.

The large room sizes are the result of local collapse of the Monteagle and Bangor Limestones into wide stream passages. As blocks fell, they were partly dissolved and eroded away by the action of the underground stream.

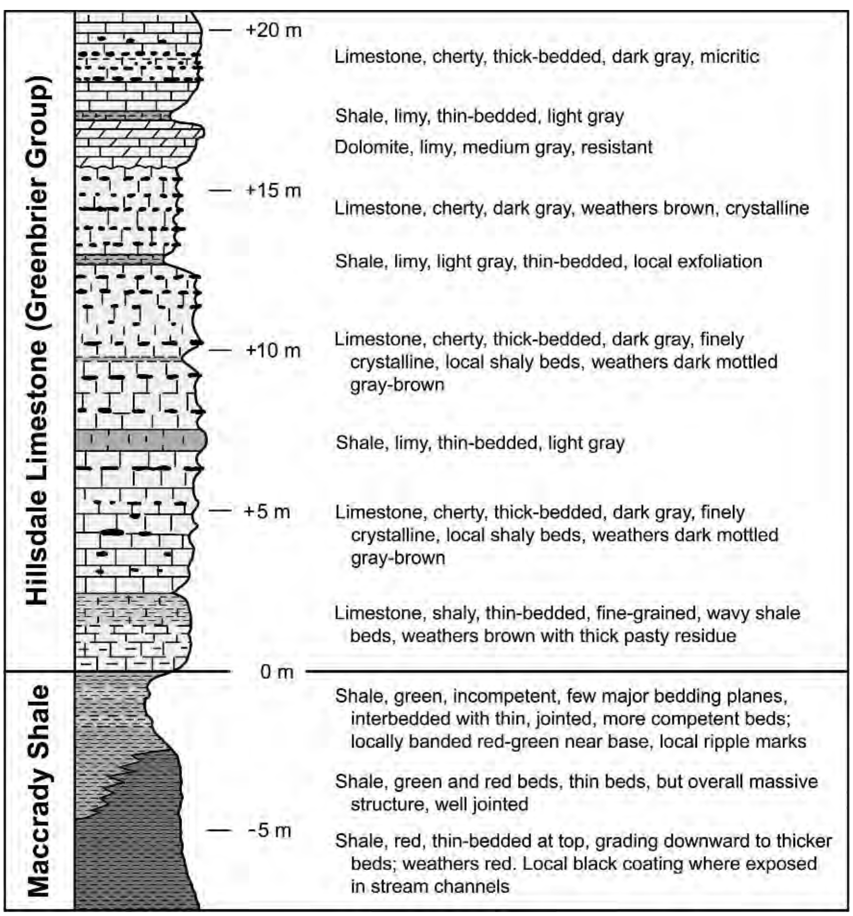

Figure 12. Stratigraphic column measured in Ludington Cave (Palmer, 1974).

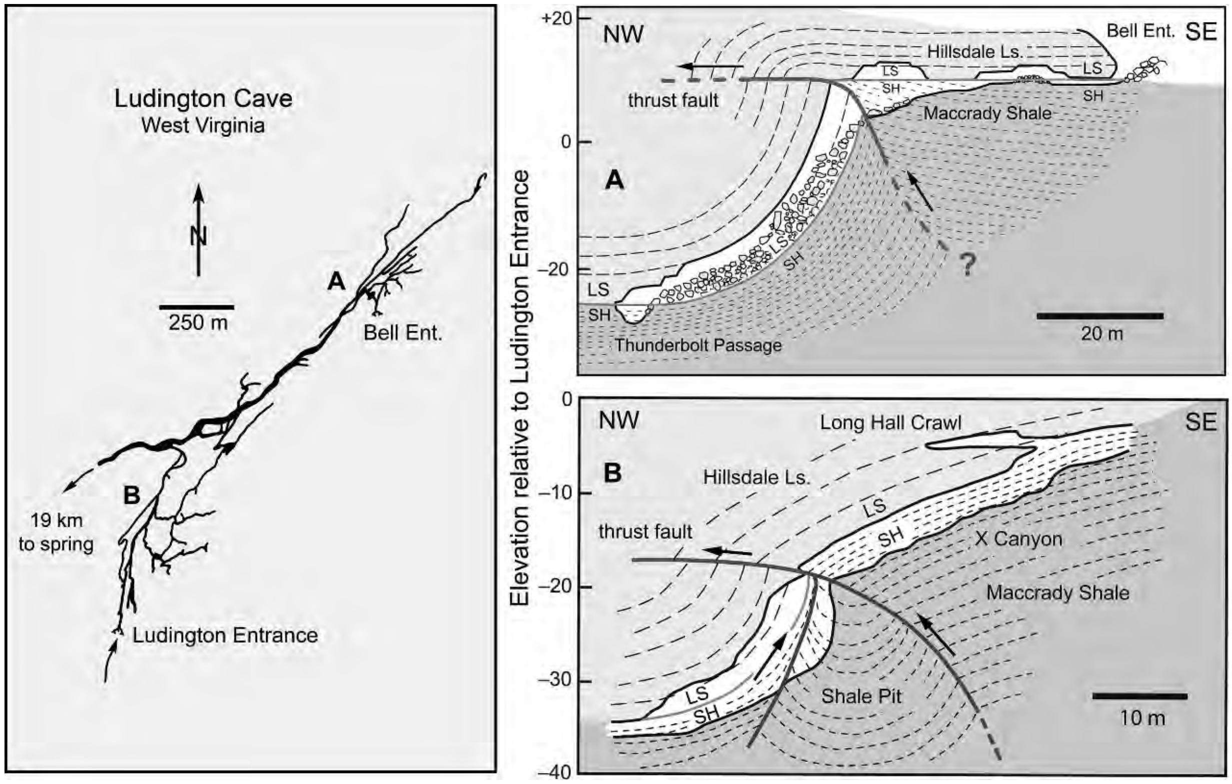

Figure 13. Representative cross sections through Ludington Cave showing fault disruption along the contact between the Hillsdale Limestone and the underlying Maccrady Shale.

Ordinarily, such massive collapse would produce a surface sinkhole. In this region, however, the limestone is capped by up to 100 m of Pennsylvanian sandstones and conglomerates. The net result is that only the very largest collapses are able to penetrate upward through the insoluble cap. These large rooms have been preserved more or less intact, with no surface expression.

The huge amount of collapse in these Tennessee caves has exposed sections of bedrock that have never been attacked by weathering, stream erosion, or even by cave-related dissolution. These sheltered rocks provide a view of geochemical processes that are rarely visible at the surface. Certain beds exposed in the walls of the rooms contain white, ghost-like shapes up to a meter in diameter (Fig. 14). Closer inspection shows that they are rounded nodules and thin dikes of gypsum (Fig. 15). These examples of primary gypsum are of special interest, because they indicate a highly evaporative depositional environment for the host limestone. This gypsum can be dissolved by vadose seepage and deposited lower in the caves as speleothems (gypsum flowers, etc.). On the other hand, most gypsum speleothems in these and other caves are composed of secondary gypsum, that is, gypsum deposited by the reaction between sulfuric acid and limestone. Reduction of gypsum in low-oxygen environments produces sulfides, either hydrogen sulfide gas or solid iron sulfides such as pyrite. When the sulfides are exposed to oxygen-rich water, they oxidize to sulfuric acid, which attacks the neighboring carbonate rock and converts it into secondary gypsum. The bases of many gypsum flowers contain blobs of iron oxide that indicate oxidation of pyrite in the exposed bedrock. Sulfur isotopic ratios ($\delta^{34}S$) tend to be positive in primary marine sulfates and negative in secondary sulfates formed by the sulfuric acid reaction (Palmer, 2007, p. 129).

In many places, the primary gypsum has been replaced by quartz or calcite (Fig. 16; Palmer and Palmer, 1991). Replacement by calcite is well understood. As fresh water infiltrates through the limestone, the water becomes saturated with respect to calcite. When it encounters gypsum, that rock dissolves rapidly. The uptake of calcium ions makes the water supersaturated with respect to calcite, and calcite is forced to precipitate. Replacement of gypsum by quartz is more difficult to explain. Quartz and other forms of silica are most soluble (as silicic acid, H_4SiO_4) at pH values greater than about 8.5. Such high pH values can result from any of several processes, but in this region, it is commonly produced by water seeping through insoluble rocks and encountering carbonate rocks at depth in a rather closed environment; as the carbonates dissolve, the CO_2 content of the water is depleted because there is no ready transfer of CO_2 from the soil (Palmer, 2007). If the water encounters sources of acid, much of the silica precipitates. A possible mechanism to account for this process is reduction of some of the gypsum to hydrogen

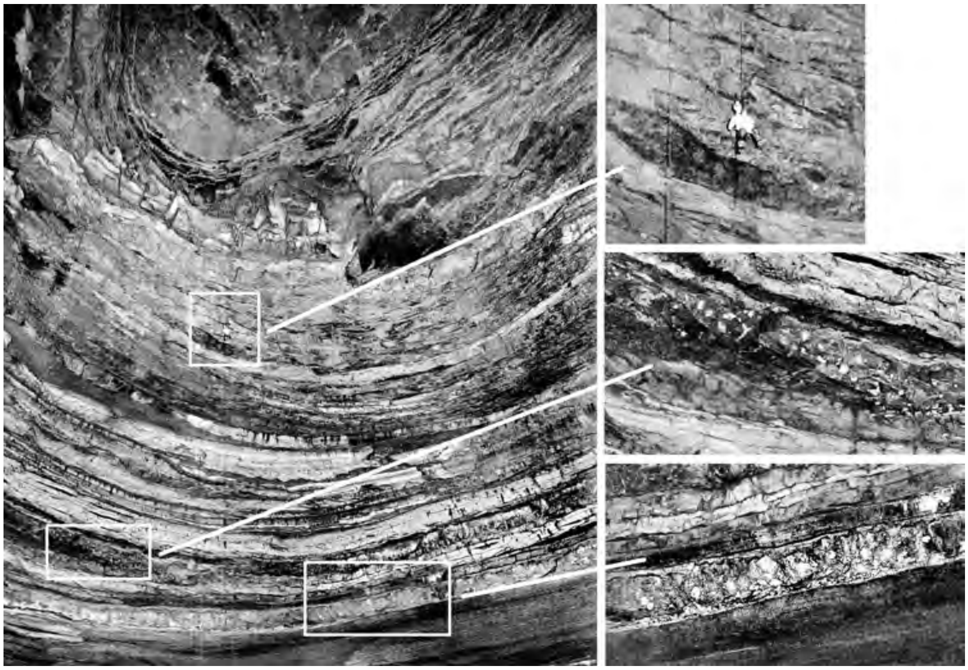

Figure 14. Upper part of a large breakdown room in Rumbling Falls Cave, Tennessee. The insets show enlargements of zones of gypsum nodules and (for scale) a person descending a rope.

sulfide, which produces neutral or possibly slightly acidic conditions. When the silica-rich high-pH water encounters the sulfate zones with their lower pH, silica can precipitate. These are processes that are rarely observed at the surface, especially in humid climates such as those of the Appalachians.

The Maze-Cave Debate

The Appalachians contain some classic examples of maze caves (Palmer, 1975; White, 1976). They are scattered throughout the Appalachians, but the largest are extensive networks in the Plateaus and in low-dip parts of the Ridge and Valley Province. These include America's quintessential network maze, Anvil Cave in Alabama, with 20.4 km of mapped passages. The Ridge and Valley Province in Pennsylvania and Virginia is particularly rich in network caves (see maps in Stone, 1953, and Douglas, 1964; see Figs. 17 and 18). Maze caves have been attributed to a variety of processes, including floodwater, diffuse seepage through an insoluble caprock, and hypogenic processes (Palmer, 1991, 2007). Their origins are of interest because they provide insight about the nature of groundwater

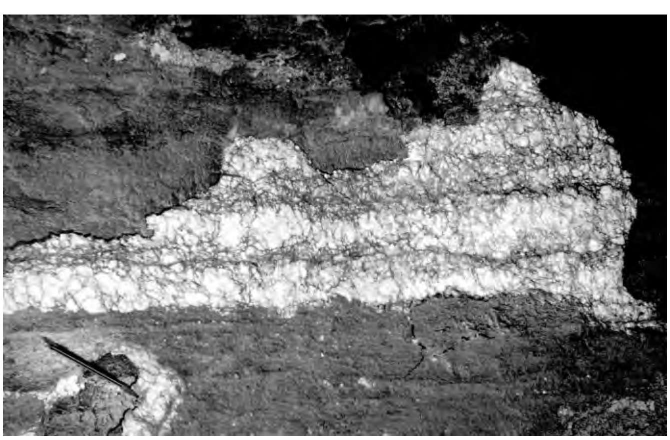

Figure 15. A dike of primary gypsum in Camp's Gulf Cave, Tennessee. Pen for scale at lower left.

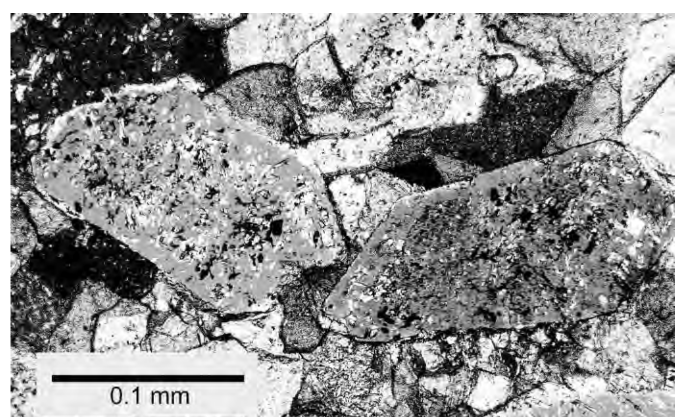

Figure 16. Thin section from a gypsum nodule in Camp's Gulf Cave, Tennessee, showing local quartz replacement. The large blocky speckled crystals are quartz, and the scattered stick-like inclusions in the quartz are residual bits of anhydrite. The crystals surrounding the quartz are gypsum.

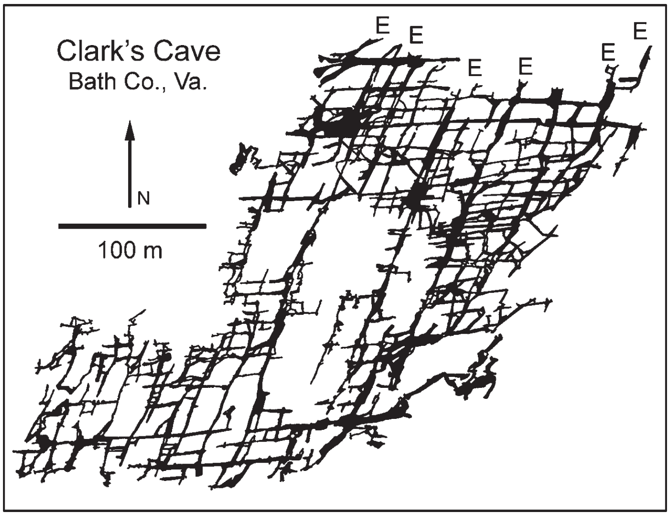

Figure 17. An example of a typical large Appalachian network maze: Clark's Cave, Virginia (based on map by Rod Morris).

hydraulics and chemistry. The largest network mazes in the Appalachians are capped by thin permeable sandstone. In active maze caves, the origin may be clear, but there are many inactive examples in the Appalachians whose origins are a matter of debate. Only the large sandstone-capped mazes are discussed here.

The lower Devonian limestones are particularly rich in network caves where there is a cap of Oriskany Formation (mainly sandstone). Examples include Paxton Cave, Helictite Cave, Clark's Cave, and Crossroads Cave in Virginia and Hamilton Cave in West Virginia. Anvil Cave, Alabama, is the largest example of all, although it is in Mississippian strata. Palmer (1975) proposed that they formed by diffuse, aggressive seepage through the thin, permeable caprock, while the governing effect of the sandstone allocated similar amounts of flow to each underlying fissure, regardless of its size.

Varnedoe (1964) attributed the origin of Anvil Cave to artesian groundwater confined by the local cap of Hartselle Sandstone (Mississippian). Maze caves in the Pennsylvanian-age Vanport Limestone of western Pennsylvania include Harlansburg Cave, longest in the state with 7 km of surveyed passages. White (1969, 1976) explains them as the product of confined flow in the thin carbonate aquifers sandwiched between insoluble beds, combined with seepage from above. For these same Vanport caves, Fawley and Long (1997) and Christenson (1999) suggest a variety of origins, including that of White, as well as the release of sulfuric acid by oxidation of pyrite in organic-rich overlying beds.

The fact that these caves cluster at and just below a sandstone-limestone contact suggests that many or most are produced by water descending through the sandstone. Maze caves are rare or absent where the caprock is thick or contains low-permeability shale. Most such caves also show evidence for backflooding from nearby rivers. Although it is likely that much of the enlargement took place by flooding, it is equally probable that their initial dissolution was the result of diffuse recharge through the overlying caprock. Many such caves receive active drips through the overlying sandstone. This does not prove that the caves formed by this mechanism, but this hypothesis cannot be dismissed without close examination.

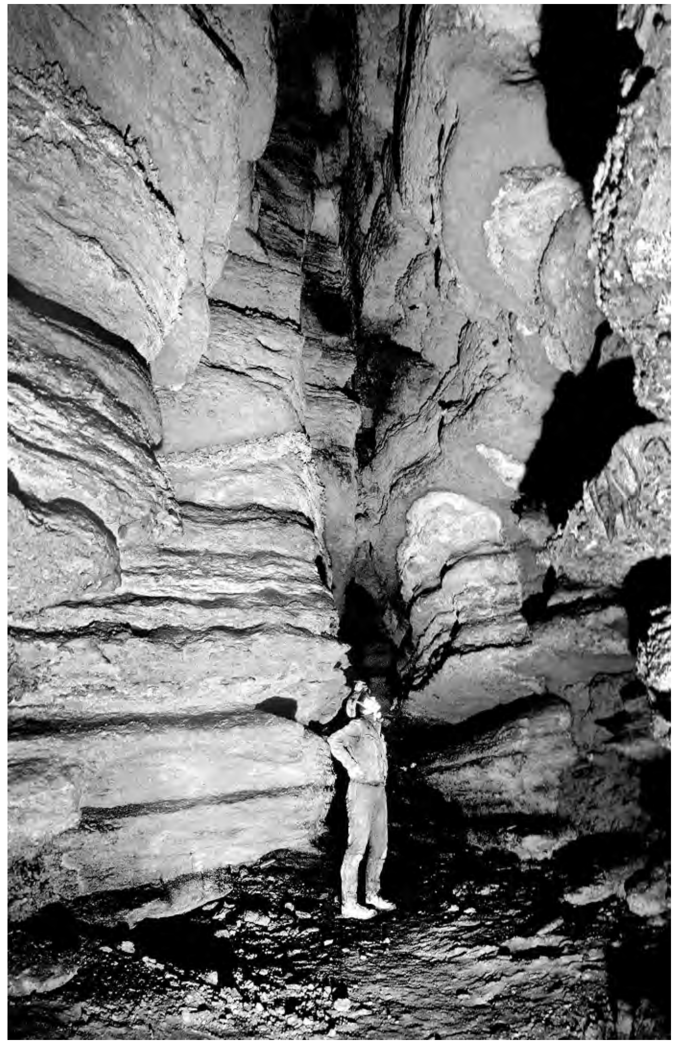

Figure 18. Fissure passage in Clark's Cave, Virginia.

Klimchouk (2007) suggests instead that these caves were formed by water rising across the strata from below (Fig. 19). There is some support for this hypothesis. Rising water is common beneath valleys, and long periods of time are available for this flow to enlarge fissures in the soluble rock. However, this hypothesis has limitations in the Appalachian limestones. The water must pass through large sections of largely insoluble rock, much of which is shale with a permeability many orders of magnitude lower than that of the sandstone caprock. To form significant caves, seepage through shale with typical permeabilities of 10^{-5} to 10^{-9} cm s^{-1} would require hundreds of millions of years.

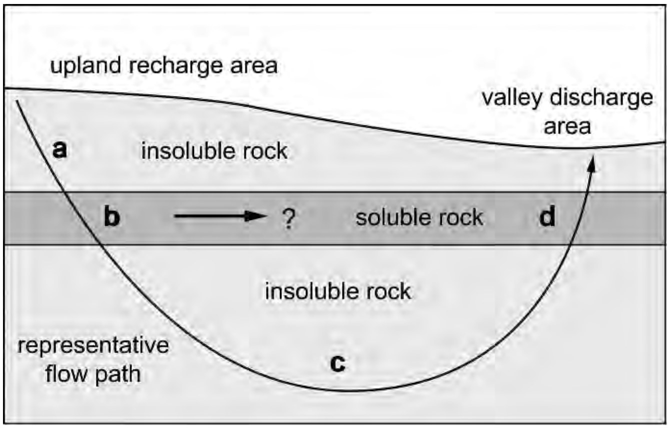

Figure 19. Contrasting models of maze-cave genesis. Palmer (1975) considers that Appalachian network caves capped by thin permeable sandstone are formed by diffuse seepage through the insoluble cap. An increase in flow through the soluble rock results in most water following paths such as a-b-d. Recharge through the cap-rock delivers rather uniform amounts of flow to each fissure. Klimchouk (2007) considers that network caves are formed by deep groundwater rising through the soluble rock (paths such as a-b-c-d), and that the surrounding insoluble rocks limit the flow so that each opening enlarges at uniform rates.

Also, it is unlikely that the rising water would have any solutional capacity at all. In Figure 19, not only has the groundwater passed through the soluble aquifer on the way down, but it has undoubtedly encountered considerable carbonate as impurities in the surrounding sedimentary rocks. Field data from wells and caves show that most seepage water of this kind will have become saturated with respect to calcite while it is still descending. This water becomes warmer with depth because of the Earth's thermal gradient, an average of about 25 °C for every kilometer of depth. Even a small rise in temperature will drive the water to supersaturation with calcite. Some calcite may precipitate, which forces most water to follow shallower paths. The water does not reach equilibrium with calcite, however, because a certain threshold supersaturation must be present to allow precipitation to occur. Thus, as the water begins to rise and cool, time is required for the water to become aggressive toward calcite once more. Whether caves can form by rising water or by descending water requires an analysis specific to each cave, based on quantitative field data.

Debates such as this are welcome, because regardless of their outcome, they lead to a deeper understanding of cave origin. Perhaps no other karst region but the Appalachians offers such a variety of field conditions to fuel these debates.

Changing Strategies in Cave Exploration

Although geologists appreciate the basic data provided by cave explorers and mappers, the benefit also reflects back to the cavers themselves, even to those with no scientific training or interest in abstract hypotheses. Records from fifty years ago show that many people searched for caves without even knowing which rocks are soluble or where they occur. Since then, a basic knowledge of cave-related geology has spread through the speleological community.

Today a great deal of time and effort is invested in digging open new caves or passages, and cavers try to optimize their chance of success by applying geologic principles. Digging is most intense in the Northeastern states, where caves are scant and access points tend to be clogged with glacial deposits. Cave diggers soon become familiar with the local stratigraphy and know the structural properties of the various beds better than most geologists. On one occasion in New York, this knowledge allowed the local diggers to rescue a trapped caver who could easily have perished without their specific skills.

An early digging success was the opening of the spring exit of Single X Cave, in Schoharie County, New York (Mylroie, 1977). This cave contains 900 m of mapped passages and one of the largest cave rooms in the state. It terminates upstream in a single fissure passage almost 600-m-long and up to 30-m-high. Atypically, the excavation was done by a geologist as part of the field work for his Ph.D. dissertation. Most current diggers are non-academic. Their first major success was at Barrack Zourie Cave, also in Schoharie County, New York, and presently third longest in the state at 5.2-km-long (Hopkins et al., 1996; Dumont, 1995). This required a 10-m excavation along a vertical fissure clogged with glacial sediment.

Recently, the digging team of Barrack Zourie fame discovered a large cave farther east that required them to widen a narrow vertical fissure in shaly limestone for about 30 m. The project nearly stalled at a particularly resistant bed, but experience told them that this was the key to entering highly soluble limestone below. A special effort put them through this barrier, and the cave opened into what promised to be one of the largest in the Northeast (Armstrong et al., 2007). To date, several kilometers have been explored. Exploration strategies rely strongly on the team's interpretation of local geology. Fracture patterns, base level, hydraulic conditions, and relationships between cave trends and stratal dip are topics of continual discussion among the group, and some of the most perceptive comments have come from cavers with no formal background in those fields. It is appropriate that they benefit from the geological knowledge to which they contribute so much.

Conclusion

Appalachian caves have provided considerable information about subsurface geologic structure, geochemistry, and hydrology, as well as speleogenesis and karst geomorphology. These topics are useful in a variety of disciplines, including the exploration for new caves. Many of the subjects introduced in this paper have barely been touched. These include the use of cave data as a guide to the development of wells and to the assessment of potential contaminant transport, quantitative validation of hypotheses for maze-cave origin, and interpretation of geochemical processes that are rarely seen at the surface.

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EFFECTIVENESS AND ADEQUACY OF WELL SAMPLING USING BAITED TRAPS FOR MONITORING THE DISTRIBUTION AND ABUNDANCE OF AN AQUATIC SUBTERRANEAN ISOPOD

Ben Hutchins[1] and William Orndorff*[,2]

Abstract: Land-use practices in karst can threaten aquatic subterranean species (stygobionts). However, since their habitat is mostly inaccessible, baseline ecological data such as distribution and population size are not known, making monitoring and risk assessment difficult. Wells provide easy and inexpensive access for sampling subterranean aquatic habitats. Over three years, including a two-month period of intensive sampling, the authors sampled sixteen wells (ten repeatedly) in Jefferson County, West Virginia, USA, for a threatened stygobiont, the isopod crustacean *Antrolana lira* Bowman, in two areas where the species was known to occur. *A. lira* was collected during 21 of 54 sampling events. *A. lira* was collected from 6 wells in which a total of 31 of the sampling events took place. Borehole logs suggest that only these 6 wells intersected appropriate habitat. Using the binomial approximation, the authors conclude that a random well has a 29% to 91% chance of intersecting appropriate habitat. In a well that intersects appropriate habitat, a single sampling event has a 51% to 85% chance of successful capture. The species occurs heterogeneously throughout the aquifer both in space and time, and thus, repeated sampling of multiple wells is needed to confidently establish presence or absence. In a contiguous block of phreatic carbonate-aquifer habitat analogous to that in the study area, at least 6 wells need to be sampled at least one time each to determine absence or presence of *A. lira* with 95% confidence. Additional studies with larger sample size would better constrain confidence intervals and facilitate refinement of minimum sampling requirements. In one well that consistently yielded from 8 to 19 animals, the population was estimated by mark-recapture methods. The limited data only allowed a very rough result of 112.3 ± 110 (95% CI) individuals. Successful recapture suggests that animals are largely stationary when a food source is present. Animals were collected at depths below the water surface from <1m (hand-dug well and cave) to ~ 30 meters in drilled wells. No migration of animals between wells was observed.

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Introduction

The Shenandoah Valley of West Virginia and Virginia is a karst landscape experiencing rapid population growth. Agricultural and urban modification of karst landscapes can lead to contamination and drawdown of karst aquifers, potentially threatening stygobionts (species limited to subterranean aquatic habitats). However, population sizes, ranges, and the distribution of individuals within aquifers is not known for many species, making monitoring and assessment of populations difficult. As international recognition of the significance of groundwater fauna grows, various methods are being developed and tested to sample groundwater habitats and develop predictive models to better understand stygobiont distributions, patterns of abundance, and autecological data (Castellarini et al., 2007, Dole-Oliver et al., 2007, Eberhard et al., 2007, Hancock and Boulton, 2007). For threatened and endangered stygobionts, these data are even more important. The paucity of basic ecological data for most stygobiont species can primarily be attributed to the challenges associated with sampling subterranean habitats. Caves, springs, and wells where biological sampling of karst aquifers is possible are small, isolated points of access into a potentially extensive, complex habitat.

For the majority of stygobionts and troglobionts in the United States, distributional data and population-size estimates have been based on collection efforts in caves (Culver et al., 2003; Fong et al., 2007; Krejca, 2004), while other access points to subterranean habitats, such as springs and wells, have been sampled less thoroughly. However, a large amount of literature demonstrates that

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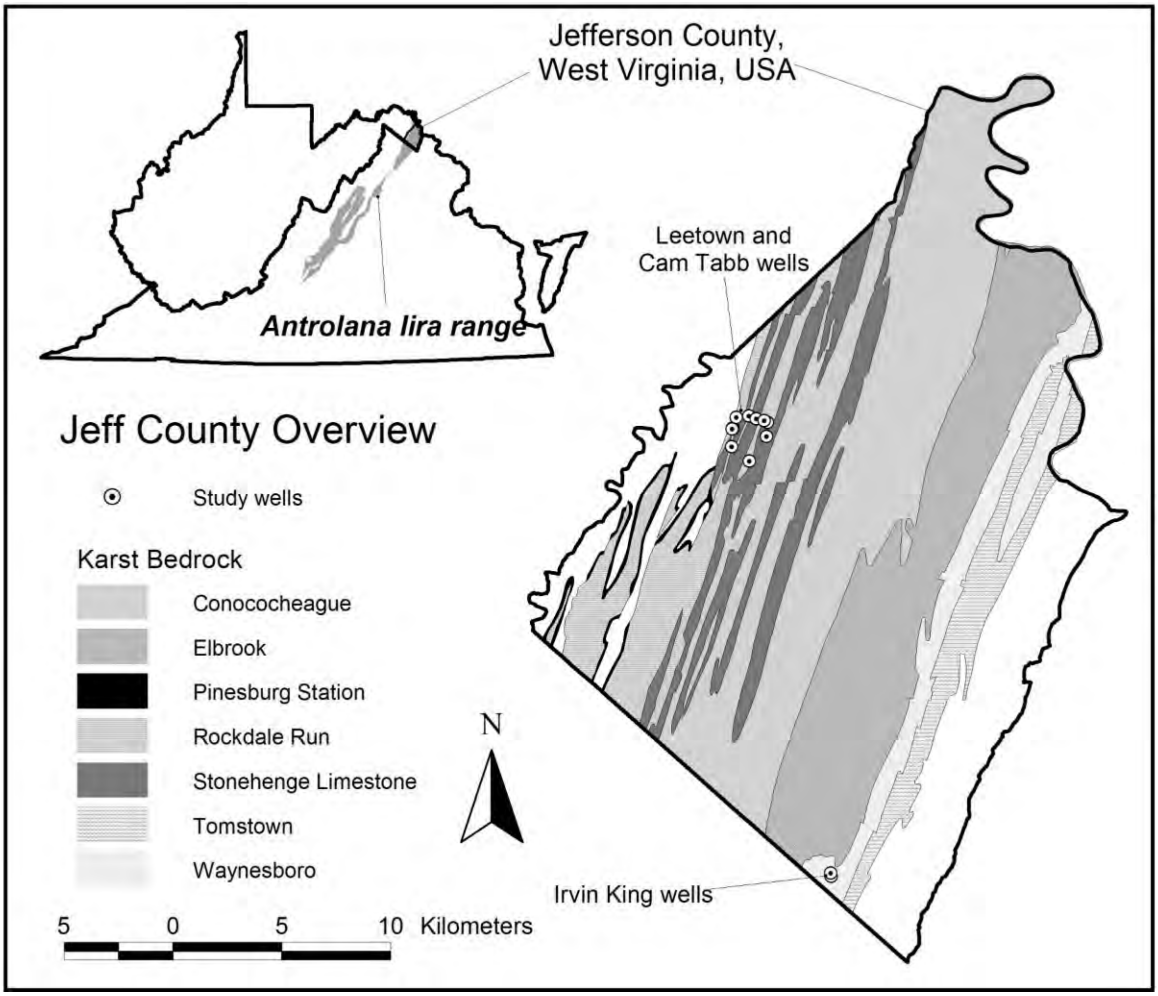

Figure 1. Project location with wells and geology.

wells can be important sampling sites for stygobionts inhabiting the phreatic zone (the saturated zone) in many types of aquifers, including karst (Allford et al., 2008; Culver and Sket, 2000; Eberhart et al., 2007; Hershler and Longley, 1986; Holsinger and Longley, 1980; Malard et al., 1997; Malard and Simon, 1997; Watts and Humphreys, 2003). For carnivorous taxa including amphipods, isopods, and planarians, baited traps can be used (Ginet and Décou, 1977) as an effective and inexpensive, albeit qualitative, sampling method. Wells are more easily accessed than groundwater in caves and are, in some areas, more numerous. This is especially true for the northern Shenandoah Valley, where surface expression of karst is minimal and few known caves extend to the water table.

In 2000, a population of the phreatic stygobiont crustacean *Antrolana lira* was discovered in a small cave in Jefferson County, West Virginia, extending the known range of the federally threatened species 50 km to the northeast. Potential degradation of the phreatic aquifer in this region has prompted concern from the U.S. Fish and Wildlife Service, but baseline ecological data, such as distribution, are needed to assess risk and implement recovery recommendations developed for the species (Fong, 1996).

Here, we present the results of a well sampling effort in Jefferson County, West Virginia, at the north end of *A. lira's* range. This effort included one sampling event in May 2005, one in July 2006, and several over a three-month period during the summer of 2007. Results are used to assess the effectiveness of well sampling for determining presence or absence for *A. lira.* The proportion of wells that intersect habitat where *A. lira* is present was calculated, along with 95% confidence intervals. Furthermore, the probability of capturing *A. lira* at wells where the species is present was also calculated, along with associated 95% confidence intervals. Several wells were sampled simultaneously at multiple depths corresponding to water-bearing fractures or voids to investigate the vertical distribution of the species in these wells. At one well, animals were marked and recaptured to estimate population size. These data are used to develop some preliminary guidelines for future well sampling in other parts of the species range and for efforts targeting other species.

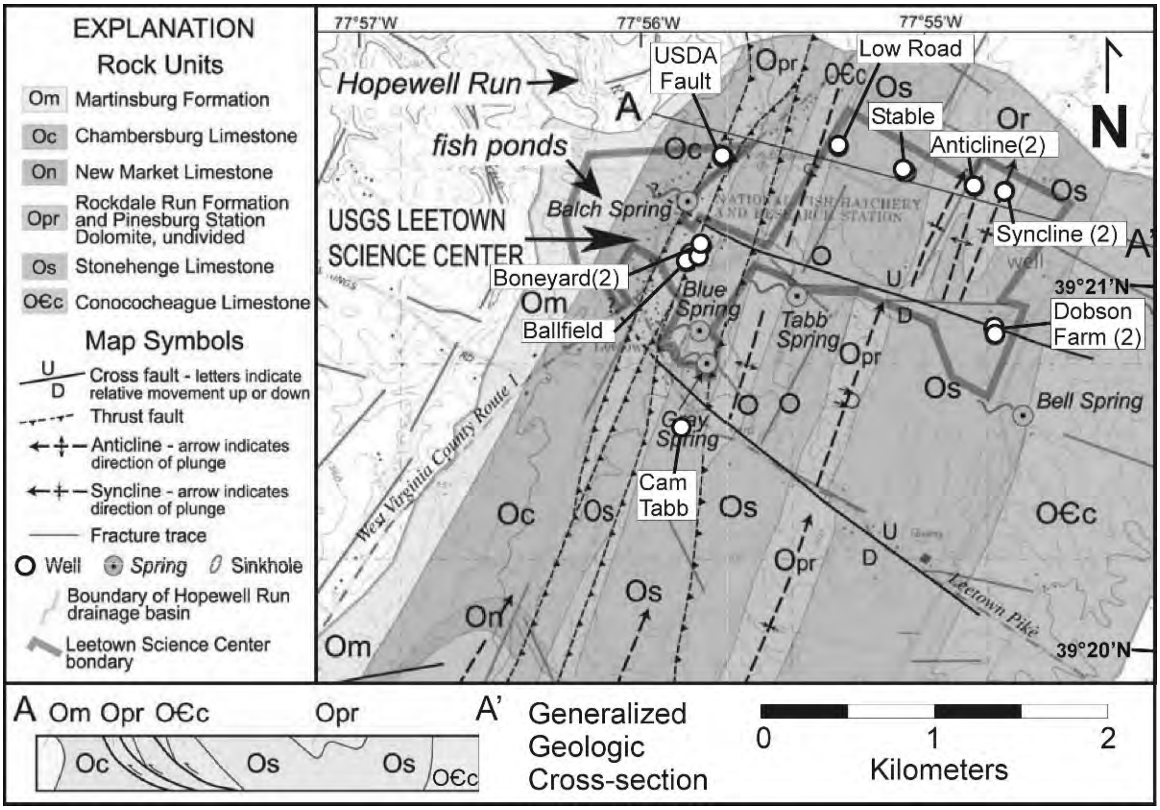

Figure 2. Leetown Science Center wells with topography and geology (modified from Kozar et al., 2007a).

Methods

Sampling Methodology

Seventeen wells were sampled between July 1, 2007, and September 3, 2007. Four of these wells were sampled a single time, and the remaining thirteen wells were sampled between two and seven times. Data from a single sampling event in May 2005 and another in July 2006 were also used for analysis. Wells were located in the karst of Jefferson County, West Virginia, on private property or federal land (Fig. 1). The study area lies within a single contiguous habitat block, here defined as a block of carbonate bedrock bounded by a combination of non-carbonate rocks and base-level streams receiving discharge from the aquifer. The contiguous habitat block involved in this study is bounded to the east and south by the Shenandoah River, to the north by the Potomac River, and to the west by the Martinsburg shale. Analysis of the mitochondrial *CO1* gene in *Antrolana lira* by Hutchins et al. (2010, in press) showed that animals from sites distributed across this bedrock block constitute a single genetic population. Three types of wells were sampled: hand-dug wells, potential production wells, and monitoring wells. Hand-dug wells were usually wide (1 m or more in diameter) and shallow (less than 10 m deep). Potential production wells and monitoring wells had ~15-cm-diameter well casings. All but two of the wells were located on or immediately adjacent to the USGS Leetown Science Center in west-central Jefferson County (Fig. 2) and had been the subject of prior intensive geohydrological investigations (Kozar et al., 2007a; Kozar et al., 2007b). This earlier work provided an unusual amount of detail in terms of the physical characteristics, hydrological properties, and geological setting of the wells used in this study, as summarized in Table 1.

Wells were sampled with a baited trap modified from Boutin and Boulanouar (1983). Baited traps were chosen for this study because they have a history of effective recovery of *A. lira,* as well as numerous other crustacean stygobionts (Collins and Holsinger, 1981; Fong, 2007). Traps were constructed using a 23-cm-long, 1.54-cm-diameter PVC pipe with a cap at the bottom. This narrow design was less likely to get lodged in the well than wider designs. Eight 8-mm holes were drilled around the top six inches of the trap. A piece of raw shrimp, wrapped in pantyhose to minimize ingestion by stygobionts, was used as bait. Traps were lowered into wells using kite string or nylon cord. A surveying tape was used to lower traps to arbitrary depths or to depths corresponding to water-bearing fractures identified in Kozar et al. (2007b). Traps were left for 20 to 28 hours. After animals were counted and possibly marked, they were released using a "release trap" made from a short length of 1.54-cm-diameter PVC pipe (Fig. 3). A piece of panty hose was secured around the bottom opening in the pipe using a rubber band. At the other end, a string was attached for lowering the trap into the well. Traps were

Table 1. Characteristics of wells sampled in study (adapted from Kozar et al, 2007b).

Well Name	Depth, m	Geology	Soil Thickness, m	Regolith Thickness, m	Top of Bedrock, m	Casing Depth, m	Well Diameter, m	Yield, L min^{-1}
Lower Road	125	SH	6.7	1.2	7.9	11.3	15.24	68
Stable Piez	14	SH	3.8	N/A	3.8	11.3	7.62	132
Ball Field	49	RR	3.0	0.0	3.0	11.7	15.24	19
Ball Field Piez	0	RR	…	0.0	…	…	15.24	…
Boneyard Upper	34	RR	4.3	0.0	4.3	13.1	15.24	151
Boneyard Lower	28	RR	3.0	2.0	5.0	5.8	15.24	379
Cam Tabb	~10	SH	…	…	…	N/A	>100	…
USDA Fault	61	RR	1.2	0.0	1.2	29.9	15.24	1135
Syncline	67	RR	3.7	4.9	8.5	28.3	15.24	1135
Syncline Piez	24	RR	5.0	0.5	5.5	18.0	7.62	379
Anticline	79	RR	6.1	1.5	7.6	11.7	15.24	76
Anticline Piez	13	RR	7.3	0.0	7.3	9.4	7.62	26
Irvin King #1	53	WE	…	…	…	…	15.24	38
Irvin King #2	38	WE	…	…	…	…	15.24	57
Old Dodson	19	SH	…	…	…	6.1	15.24	…
New Dodson	51	SH	…	…	…	11.7	15.24	…

Geology: SH – Stonehenge Formation, RR – Rockdale Run Formation, WE -Waynesboro-Elbrook Formations

lowered slowly through the water column until reaching the approximate depth at which the animals were captured, at which point the trap was repeatedly lifted and dropped (causing water to flow through the pipe, dislodging the panty hose and the animals).

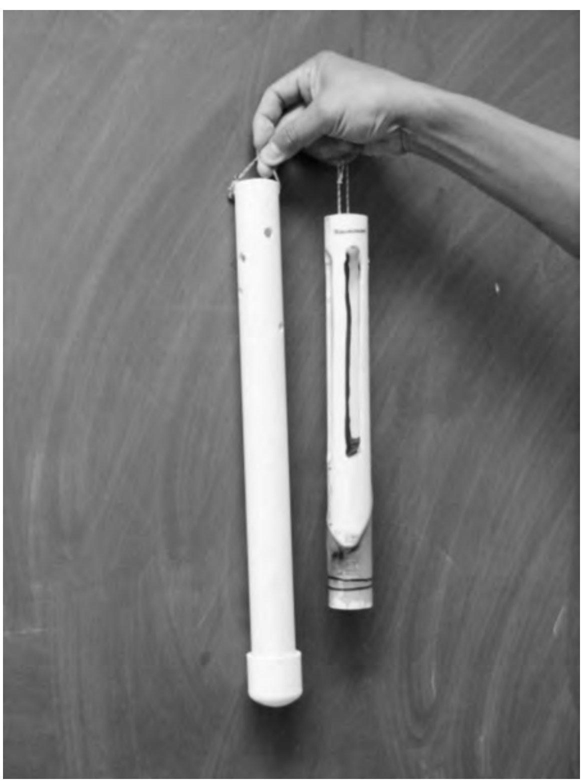

Figure 3. Capture and release traps.

CAPTURE PROBABILITIES

Data on the capture rates of *A. lira* were used to estimate both the success rate at wells where *A. lira* was captured at least once and the proportion of wells in the contiguous habitat block that intersect habitat where *A. lira* are present. Days when multiple traps were used in a single well at different depths were treated as single sampling events, with capture of *A. lira* in one or more traps constituting a positive result.

By approximating capture data as a binomial approximation to the normal distribution, the standard deviation σ_p of capture rates was calculated using

$$\sigma_p = \sqrt{\frac{p(1-p)}{n}} \tag{1}$$

where n is the number of trials and p is the success rate (Lichter, 1999). Standard deviation σ was then used to approximate 95% confidence intervals ($p \pm 2\sigma$, Ott and Longnecker, 2001).

The probability of capture during a single sampling event at a well n in an area where a target species is present was calculated by

$$P_{capture,n} = P_{habitat,n} \times P_{success,n} \tag{2}$$

where $P_{habitat}$ is the probability that the well intersects habitat where *A. lira* is present, and $P_{success}$ is the probability that a single sampling event in a well that intersects such habitat will result in capture. The standard deviation of the product was calculated using conventional error-propagation calculations as described in Lichten (1998).

The minimum number of sampling events T needed to determine if the species was present in an individual well

Table 1. Extended.

Depths to Water Bearing Features, m						Depth to Water (7/2003–10/2005) Mean	S.D.	Min.	Max.	Range
35.1	113.4	121.6	…	…	…	14.1	0.50	12.8	14.8	2.0
…	…	…	…	…	…	10.5	0.85	8.8	11.7	2.9
10.7	…	…	…	…	…	6.0	0.53	4.7	7.2	2.5
…	…	…	…	…	…	…	…	…	…	…
16.8	20.4	…	…	…	…	6.3	0.37	5.5	7.1	1.7
10.4	14.9	19.8	21.6	23.8	…	5.5	0.41	4.6	6.5	1.9
…	…	…		…	…	…	…	…	…	…
9.4	14.3	18.6	35.1	39.9	47.2	5.2	0.28	4.7	6.0	1.2
12.5	31.1	43.3	…	…	…	4.3	1.02	2.4	6.1	3.6
7.6	21.3	…	…	…	…	3.3	0.97	3.6	7.1	3.5
8.8	41.8	51.8	76.2	…	…	5.9	0.99	4.1	7.6	3.5
7.3	…	…	…	…	…	5.5	0.97	3.6	7.1	3.5
…	…	…	…	…	…	…	…	…	…	…
…	…	…	…	…	…	…	…	…	…	…
…	…	…	…	…	…	2.7	0.54	1.7	4.3	2.5
…	…	…	…	…	…	…	…	…	…	…

and the number of wells W that needed to be sampled to determine if the species was present in an area were calculated from the probability of encountering all negative results after a number of trials N using

$$P_{neg,N} = \left(1 - P_{pos}\right)^N \tag{3}$$

where $P_{neg,\ N}$ is the probability of all negative results after N trials and P_{pos} is the probability of a positive result (assumed constant) for any individual trial. For multiple sampling events at a single well, $N = T$ and $P_{pos} = P_{success}$. For a single sampling event at multiple wells in a contiguous habitat block, $N = W$ and and $P_{pos} = P_{capture}$. When $P_{neg,\ N} = 0.05$ after N trials, this means there is only a 5% chance of no positive results (i.e., a false negative) if a species was present in an area. Conversely, this means that there is a 95% chance that all negative results after N trials constitutes a true negative, in our case, no animals present. Plugging in the certainty value of 0.05 and solving for N produces

$$N = \frac{\ln\,(0.05)}{\ln\,\left(1 - P_{\mathrm{pos}}\right)} \tag{4}$$

In general, a species may be absent from a well either because it is not present in the area or because the well does not intersect appropriate habitat. Since this study was confined to a contiguous habitat block where the species is present, a consistently negative result within any individual well most likely reflects a failure to intersect appropriate habitat.

Vertical Distribution

Eight wells were chosen to study the vertical distribution of *A. lira* based on their water yields and the existence of data on the depths of water-bearing fractures or voids (Kozar et al., 2007b). Depending on the number of reported water bearing features in each well, from two to six traps were placed at depths corresponding to these features. In addition to these eight wells, four wells for which no data about water bearing voids was known (Irvin King #1, Irvin King #2, Old Dodson, New Dodson) were sampled. For these wells, traps were placed at 7.6-m intervals starting at the bottom of the well.

Population-Size Estimation

At one well, animals were marked and recaptured to estimate population size. Trapped animals were stored in cool spring water on site for mark and release. To mark animals, we first patted the animal's dorsal surface with a napkin before using a Sharpie brand marker to make an identifiable mark. Population size was estimated using a weighted mean method (Begon, 1979). This method is similar to the traditional Peterson estimate, but employs data from more than one sampling event and uses the equation

$$\hat{N} = \frac{\sum n_i M_i}{\sum m_i + 1} \tag{5}$$

where n_i is the number of individuals caught on sampling day i, m_i is the number of individuals collected on day i that are already marked. $M_i = (r_2 - m_2) + (r_3 - m_3) \ldots + (r_i - m_i)$, where r_i represents the total number of animals marked and released on the indicated days, including those captured that had

Table 2. Summary of sampling results for Madison Cave Isopod (*Antrolana lira*).

Site Name	5/5/2005	7/8/2006	7/1/2007	7/8/2007	7/15/2007	7/29/2007	8/17/2007	8/26/2007	9/3/2007
Lower Road Well	0	···	0	···	···	···	···	···	···
Stable Piezometer	0	···	0	···	···	···	···	···	···
Ball Field Well[a]	1	···	···	0	···	···	0	0	···
Ball Field Piezometer	···	···	0	···	···	···	···	···	···
Boneyard Upper Well	···	···	0	···	···	···	···	···	···
Boneyard Lower Well	···	···	0	···	···	···	···	···	···
Cam Tabb Well[a]	···	68	1	7	1	···	2	2	5
USDA Fault Well[a]	0	···	0	···	···	2	1	···	···
Syncline Well[a]	0	···	2	1	0	1	0	1	···
Syncline Piezometer	···	···	0	···	···	0	···	···	0
Anticline Well	0	···	0	···	···	0	···	0	0
Anticline Piezometer	···	···	···	···	···	···	···	···	0
Irvin King #1 Well[a]	···	0	···	···	···	···	···	0	2
Irvin King #2 Well[a]	···	20	8	20	13	12	9	···	···
Old Dodson Well	0	···	···	0	···	···	···	0	0
New Dodson Well	···	···	···	0	···	···	···	0	0

[a] Captured well.

previously been marked. The standard error is calculated using

$$SE_{\hat{N}} = \hat{N}\sqrt{\frac{1}{\sum m_i + 1} + \frac{2}{\left(\sum m_i + 1\right)^2} + \frac{6}{\left(\sum m_i + 1\right)^3}} \quad (6)$$

Results

Fifty-four sampling events were performed at a total of 18 wells (Table 2). Six wells, referred to as capture wells, yielded *Antrolana lira* at least once. The physical and hydrological characteristics of these wells are summarized in Table 1. Of all the sampling events at capture wells, individuals were captured 21 out of 31 sampling events (68%). Ten wells were sampled between three and seven times to accumulate data on the temporal variation in the presence and abundance of species collected. Table 2 shows results of all sampling events performed during this study. Days when multiple traps were placed in a well on the same day were treated as a single sampling event. Figure 4 illustrates the variation over time of capture rates at each well in which *Antrolana* was captured at least once.

Positive capture rates at wells where *A. lira* was captured at least once ranged from 25% to 100%. In the two wells with relatively high numbers of individuals, Cam Tabb and Irvin King #2, *A. lira* was present 100% of the time. In the other four capture wells, a maximum of two animals were captured during any single sampling event. Furthermore, each of these wells had at least one sampling event in which no animals were captured.

Capture Probability Results

The probability of success at capture wells was estimated at $P_{success} = 0.68 \pm 0.08$, with 95% confidence intervals of $0.51 < P_{success} < 0.85$. Applying Equation (4) to the results for $P_{success}$, the minimum number of sampling events at a well to determine whether it intersects habitat, based on successful capture during one or more event, is three (T = 2.63) using the predicted value of $P_{success}$, and five (T = *4.2*) based on the lower end of the 95% confidence interval. The criteria for use of the binomial approximation as described in Ott and Longnecker (2001) are met for $P_{success}$.

Wells within the study area sampled three or more times can then be used to estimate the habitat intersection rate $P_{habitat}$. *A. lira* was captured at least once in 6 of the 10 wells sampled 3 or more times, resulting in $P_{habitat} = 0.60 \pm 0.16$. Within 95% confidence limits, $0.29 < P_{habitat} < 0.91$.

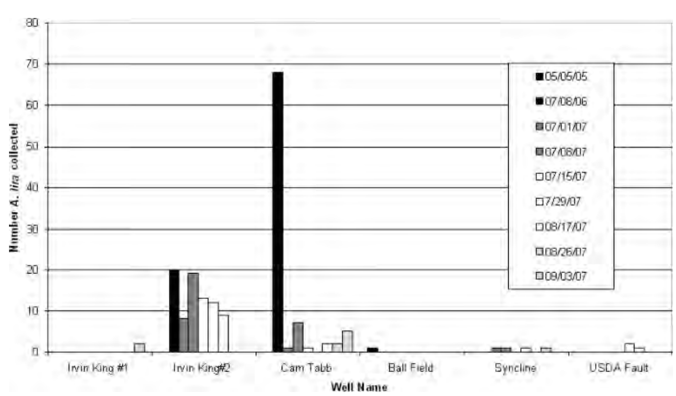

Figure 4. Capture rate variation at wells where *Antrolana lira* was collected at least one time.

Table 3. Vertical distribution of Madison Cave Isopod (*Antrolana lira*) in wells.

		Number of Individuals by Date Sampled			
Well Name	Depth (m)	7/29/ 2007	8/17/ 2007	8/26/ 2007	9/3/ 2007
USDA Fault	9	0	0	···	···
	14	0	0	···	···
	19	0	0	···	···
	35	2	0	···	···
	40	0	1	···	···
	47	0	0	···	···
Syncline	12	0	0	0	···
	31	1	0	1	···
	43	0	0	0	···
Irvin King #1	15	···	···	0	0
	24	···	···	0	2
	37	···	···	0	0
Irvin King #2	23	0	0	···	···
	30	0	0	···	···
	38	10	0	···	···
	44	2	9	···	···

Applying Equation (4), the minimum number of wells necessary to sample to ensure intersection of habitat is four (W = 3.27) for the predicted value of $P_{habitat}$, and nine at the lower end of the 95% confidence interval (W = 8.75). The low number of wells (ten) sampled enough times to determine $P_{habitat}$ limits the significance of these numbers, because the criteria for use of the binomial approximation, as described in Ott and Longnecker (2001), are not met.

Applying Equation (2), the probability of capture $P_{capture}$ for a single sampling event at a single well within the study area is 0.41 ± 0.12. Within 95% confidence limits, $0.17 < P_{capture} < 0.65$. High standard deviation and large confidence intervals are a result of the low number (ten) of wells sampled three or more times combined with the propagation of uncertainty in $P_{success}$ and $P_{habitat}$. For the calculated $P_{capture}$ of 0.41, the corresponding minimum number of unique sampling events (individual wells sampled one time each) necessary to determine whether the species is present in a contiguous phreatic habitat block such as the study area is six (N = 5.68). However, if the lower end of the 95% confidence interval is used, the minimum number of trials for such a determination increases to sixteen (N = 16.07).

Vertical Distribution Results

Table 3 shows the dates and depths at which individual wells were sampled at multiple levels and when and at what level individuals of *A. lira* were recovered. Four of those wells yielded *A. lira.* In USDA Fault Well, *A. lira* was found at water-bearing horizons at 35 m and 40 m. In Syncline Well, *A. lira* was collected from traps at the 31-m water-bearing horizon during four out of six sampling events, while horizons at 12 m and 43 m yielded no individuals. Irvin King #2 yielded multiple individuals at depth of 38 m and 44 m, and no animals at 23-m and 30-m depths. Irvin King #1 yielded individuals at a depth of 24 m. On average, *A. lira* was collected from 31% ± 4% of water-bearing horizons in each of these four wells.

Table 4. Mark recapture data for Irvin King #2 well.

	Number of Individuals by Date Sampled		
Variable	7/1/2007	7/8/2007	7/15/2007
Number captured, n	8	19	13
Number marked, m	···	2	3
Number marked and released, r	8	19	13

Population Size Estimation Results

Marked animals were only recaptured at Irvin King Well #2, and consequently, population size could not be estimated at other locations. At Irvin King Well #2, animals were captured, marked, and released on July 1, July 8, and July 15, 2007. Table 4 summarizes the data used to calculate the population estimate and uncertainty using Equations (4) and (5). The limited population being sampled at Irvin King Well #2 only allowed a very rough estimate of 112.3 individuals ± 110 (95% CI).

Discussion

Capture Probability Discussion

While it is clear that well sampling using baited traps is an effective way to sample for stygobiont crustacean fauna such as the Madison Cave isopod *Antrolana lira*, interpretations of results must be performed conservatively and with caution. At least three conditions must be met for a successful capture. First, the sampling site must be within the range of the species. Second, the well must intersect appropriate habitat, in this case interconnected, permeable voids beneath the water table that are large enough to be traversable by the species. Finally, the trap must effectively attract and retain animals. The efficiency with which a particular sampling method attracts and retains animals must also be considered when comparing data from multiple sampling methods. Allford et al. (2008) tested three different sampling methods on wells in the Yilgarn region of Australia and found differences in the number of species and total number of individuals collected, but no significant difference in the relative probability for capturing a particular species as a function of sampling method.

In Leetown, capture wells appear randomly distributed within a contiguous phreatic habitat block, suggesting that the entire study area lies within the potential range of the species. However, within this range, the species is heterogeneously distributed, depending on the presence of favorable habitat, which is patchy but interconnected (Hutchins et al., 2010, in press). The fact that this habitat is hidden from view complicates any sampling strategy. This study seeks to calculate the probability that the last two conditions are met: a given well intersects favorable habitat ($P_{habitat}$) and that the species is collected during the sampling event ($P_{success}$). We found that $P_{habitat} = 0.60 \pm 0.31$ (95% C.I.) and that $P_{success} = 0.68 \pm 0.17$ (95% C.I.) for *A. lira* in our study area. These values were used to predict that for a unique sampling event for *A. lira* within the study area the probability of capture is 0.41 ± 0.12. (95% C. I.: $0.17 < P_{capture} < 0.65$) and to estimate that approximately six sampling events are necessary to determine if the species is present in a similar contiguous phreatic habitat block (sixteen events if the lower end of the confidence interval is used). Exporting the results to outside the study area assumes that neither $P_{success}$ nor $P_{habitat}$ varies significantly from one contiguous phreatic habitat block to another. Unfortunately, enough data points were not collected in the study to tightly constrain the predicted value of $P_{capture}$ in the study area, although $P_{success}$ and $P_{habitat}$ were moderately well constrained. Our results were similar to those of Eberhard et al. (2007), who used net-haul sampling in the Pilbara region of Australia and found detection probabilities for species to average 33 $\pm$ 5% or 39 $\pm$ 3% (two different methods) and that six samples collect 95% of species present in a well.

For those interested in determining with certainty the absence or presence of a stygobiont in an area, a paucity of sampling locations and low densities of animals presents a high risk of false negatives. Obviously, the best way to reduce this risk is to increase the number of sites sampled and the number of sampling events. However, the number of available sampling sites in a contiguous phreatic habitat block is essentially fixed. This makes desirable a sampling scheme that samples sites on multiple occasions to achieve the desired level of certainty in the presence or absence of a species.

MacKenzie et al. (2002) developed such a technique and applied it to a data-set investigating site occupancy of amphibians in Maryland, USA. Their model considered the probability of the presence of a species at a site, the number of sites, the number of sampling events, and the probability of detection. Such a model could be effectively applied to the stygobiont sampling scenario described in this paper if the probability of the presence of a species was replaced with that of habitat intersection. Unfortunately, the data-set in this study was too small for these methods to be applied.

There was no obvious relationship between physical and hydrological properties of the individual wells (Table 1) and the presence or absence of *Antrolana lira*. While the two highest-yield wells (Syncline and USDA Fault, each 1135 L min^{-1}) both yielded specimens, so did low-yield wells such as Irvin King #1 (38 L min^{-1}), Irvin King #2 (57 L min^{-1}), and Ball Field (19 L min^{-1}), with Irvin King #2 being the most consistent producer of *A. lira*. In terms of geology, specimens were successfully captured from at least one well in all formations in which wells were sampled.

Differences in the May 2005, July 2006, and summer 2007 sampling events suggest that groundwater levels may strongly influence sampling success rates, both in terms of numbers and of stygobiont species. This is in contrast with the results of Eberhard et al. (2007), who found no seasonal turnover in faunal composition in sampling wells over a 4-year period in the Pilbara region of Western Australia. Figure 5 shows water levels in the aquifer at Leetown Science Center over the period of interest. Both the May 2005 and July 2006 sampling events took place during relatively high groundwater levels, immediately after significant recharge events, while the summer 2007 sampling was performed under drought conditions. In May 2005 the water level was more than a meter higher than in summer 2007, and numerous amphipods were captured in the Old Dodson Farm Well and the Ballfield Well, which also yielded a single *Antrolana lira.* Neither of these wells yielded a single crustacean specimen during summer 2007 sampling. The July 2006 sampling event at Cam Tabb Well stands out as well. Sixty-eight individuals were collected in that event, compared with a range of 1 to 7 individuals captured during 2007 sampling events, when water levels were approximately 0.6 m lower than in 2006. This apparent water-level influence on sampling results may have to do with water levels reaching the elevation of specific conduits, allowing the animals to move within the aquifer. Alternatively, the presence of larger numbers of animals following recharge events may reflect flushing of animals from different hydrological realms in the subsurface. A third possible explanation is that the animals may be more active within the aquifer in response to a higher food supply associated with a recharge event. In any case, these results showed that the probability of successful recovery of *A. lira* at wells that intersected habitat varied both from well to well and at an individual well over time.

VERTICAL DISTRIBUTION DISCUSSION

While the depth sampling did not yield enough data to be conclusive, it did suggest that specific water-bearing horizons are associated with the presence of certain stygobiont species, and that many of these horizons are at considerable depths (up to $\sim$ 30 m) below the water table. During all sampling events in drilled wells, *Antrolana lira* was only collected in traps placed at least 25 m beneath the land surface. During July 2007, only traps placed at least 30 m below the land surface yielded specimens. This does not hold for cave or hand-dug-well collections, neither of which generally allow for the trap to be placed more than 10 m beneath the water surface. The risk of false

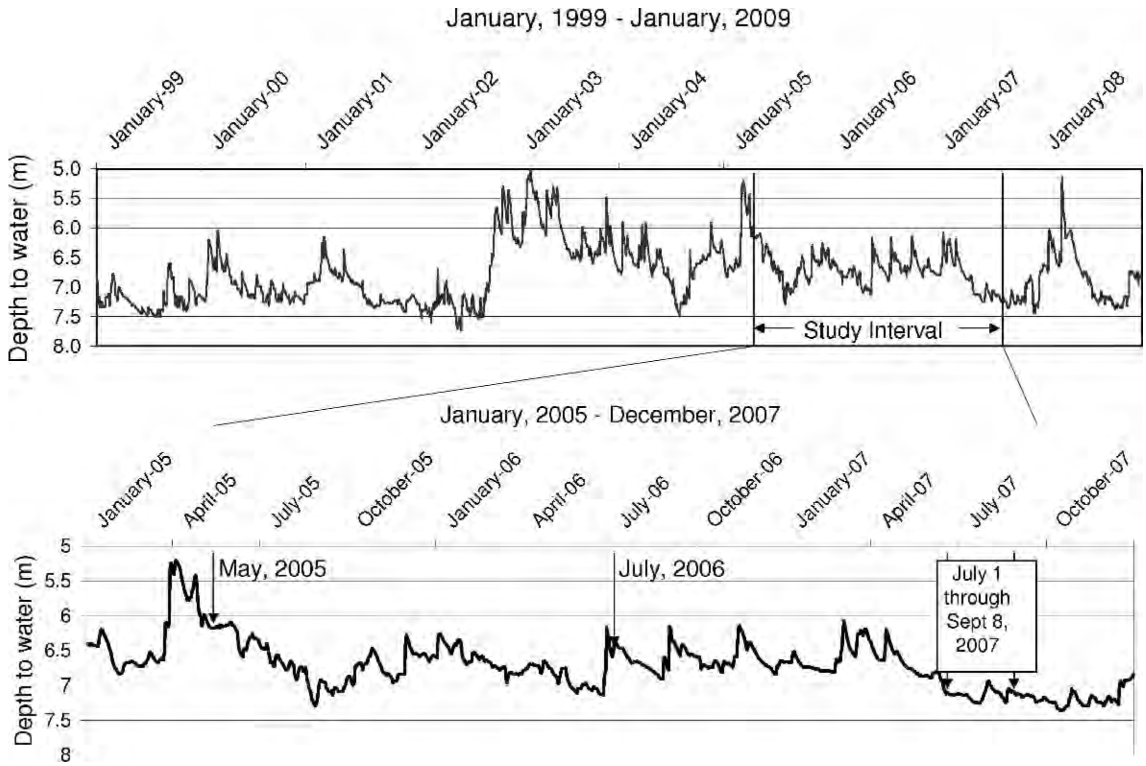

Figure 5. Water levels at Leetown Science Center monitoring wells (USGS, 2009).

negatives for wells is significant, as discussed above, and may be increased when a single trap is placed at an arbitrary depth. However, this risk may be overcome through the use of alternative sampling methods, such as haul nets that sample the entire water column (Allford et al., 2008). Results from Irvin King #2 Well show that trap level will not affect capture rate in all instances. During the mark-recapture phase of the project, a single trap was placed at an arbitrary depth (25 m, approximately 6 m beneath the surface of the water) during weeks one to three, yielding 8, 19, and 13 animals, respectively. During the horizon-sampling phase, traps at 23 m and 30 m yielded no specimens, while traps at 38 m and 44 m yielded combined totals of 12 and 9 individuals for weeks four and five. This suggests that animals were present at lower levels in the well during the mark-recapture phase and swam up the well to reach the bait. It is likely that the reason they are present lower in the well is that they are closer to the intersection of the well bore with water-bearing voids or fractures (e.g., USDA Fault and Syncline Wells, Table 2). Alternatively, Hahn and Matzke (2005) suggest that detritus and sediment that preferentially accumulates at the bottom of wells may act as habitat islands in aquifers, attracting a higher abundance of taxa than elsewhere in the aquifer. This potential relationship depends strongly on the identity and life history of the species involved.

POPULATION SIZE ESTIMATION DISCUSSION

Population size estimation was only possible at one sampling location due to the lack of recaptured specimens elsewhere. At Irvin King Well #2, 112.3 ± 110 individuals were estimated to compose the population sampled during this study. Obviously, this estimation has a large degree of uncertainty. Furthermore, as with other population-size estimation methods, this method makes a variety of assumptions. First, it assumes no births, deaths, immigration, or emigration during the sampling period. This first assumption is probably not significantly violated, given that subterranean organisms have low reproductive potential and metabolic rate and that *Antrolana lira* has no known predators. This method also assumes that capture and marking does not affect an individual's chance at subsequent capture. In another population size estimation study for *A. lira*, one week was found to be a sufficient period of time for previously captured and marked animals to be re-trapped (Fong, unpublished data). Finally, the method assumes that all individuals have an equal chance of being caught. Given the heterogeneous nature of phreatic passages, complex flow routes, and the fact that no ovigerous females have ever been captured, this final assumption may be violated in the case of *A. lira.* Nevertheless, Hahn and Matzke (2005) suggest that taxa may be preferentially distributed near wells that serve as habitat islands, and at least one mathematical model suggests that vagile taxa such as *A. lira* may be able to travel significant distances within aquifers (Eberhard et al., 2007). What these data do suggest is that this is a small population. This is corroborated by low genetic variability within the site (Hutchins et al., 2010, in press). This has implications for the conservation of the species, because low population size that is potentially clustered near the well puts the population at risk.

The only other population size estimates for *A. lira* have been performed using identical methods at Cave Hill in Augusta County, Virginia (Fong, 2007). Population size estimates at Cave Hill were much higher than at Irvin King, ranging from 0.36 to 1.02×10^3 at Madison Saltpetre Cave and 2.24 to 3.42×10^3 at Steger's Fissure (Fong, 2007). Population estimates at other documented sites within the range need to be performed to determine what population sizes are more typical for *A. lira*.

Conclusions

In some areas, the abundance of wells in proximity to one another relative to that of caves and springs allows for more comprehensive sampling across the potential range of a stygobiont species. Some karst areas, like the lower (northern) Shenandoah Valley, are particularly cave-poor, and wells afford a much better way of accessing habitat. This study has shown that if preliminary sampling efforts are sufficient to constrain the probabilities of habitat intersection and successful recovery of animals, then it is possible to develop a meaningful protocol for sampling wells with baited traps to determine presence or absence of a phreatic stygobiont. The results of such sampling are likely to vary with aquifer water levels and in response to recharge events. Use of wells with comprehensive hydrological and borehole descriptions combined with sampling at discrete depths increased understanding of the three-dimensional subterranean habitat structure. Animals were shown to be present at significant depths (up to 30 m) beneath the water table, and they appear to be using specific conduits within the aquifer. Successful completion of a mark-recapture population estimate showed that known populations of *Antrolana lira* in the northern end of its range are at much lower densities than those at the type locality of Cave Hill. Future research on this topic should include extensive additional sampling within the project area to better constrain detection and habitat intersection probabilities, replication of the study in other contiguous habitat blocks of the Madison Cave isopod to test the assumption that detection and habitat intersection probabilities are relatively constant between such blocks, and application of these methods to other phreatic stygobiont species to determine inter-species variations in detection and habitat-intersection probabilities.

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THE VIRGINIA CAVE PROTECTION ACT: A REVIEW (1966–2009)

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Abstract: The Virginia Cave Protection Act was first ratified in 1966, with a major revision in 1979, yet Virginia cave and karst resources are still threatened by vandalism, pollution, and poorly planned development. As public interest in outdoor recreation continues to grow and land development accelerates in the Appalachian Valley and Ridge Province west of the Blue Ridge Mountains, increased pressure will be put on Virginia's limited and fragile cave resources. Over the past thirty years, there have been many important court cases in Virginia, as well as countless state and federal actions. The difficulty of apprehension and prosecution of vandals demonstrates the inadequacy of current penalties. More prosecutions and harsher penalties will invariably serve as a deterrent to future potential vandals. Complex state projects, like highway widening and the construction of new prisons and airports, put additional pressure on karst areas. In order to preserve the unique educational, recreational, scientific, historic, and economic values of Virginia caves and karst, the Virginia Cave Board has been authorized to safeguard these resources.

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Introduction

The first Virginia Cave Protection Act became law on March 2, 1966, when House Bill 24 became Section 18.1-175.1 of the 1950 Code of Virginia. It was the clear the intention of the legislators to protect Virginia Cave resources, especially those found in commercial caverns bringing tourist dollars to the State. With the 1975 recodification of Title 18, the Cave Protection Act was moved to Section 18.2-142 under "Damaging Caverns or Caves" and contained two parts:

> (a) It shall be unlawful for any person, without the prior permission of the owner, to willfully and knowingly, break off, crack, carve upon, write or otherwise mark upon, or in any manner destroy, mutilate, injure, deface, mar or harm any natural material found within any cave or cavern, such as stalactites, stalagmites, helictites, anthodites, gypsum flowers or needles, flowstone, draperies, columns, or other similar crystalline mineral formations or otherwise; to kill, harm or disturb plant or animal life found therein; to discard litter or refuse therein, or; otherwise disturb or alter the natural condition of such cave or cavern; or break, force tamper with, remove, or otherwise disturb a lock, gate, door or other structure or obstruction designed to prevent entrance to a cave or cavern, without the permission of the owner thereof, whether or not entrance is gained.
>
> (b) Any violation of this section shall be punished as a Class-3 Misdemeanor. [Changed in 1975 from a fine not exceeding $500 or confinement in jail not exceeding 12 months.]

In January 1978, members of the Virginia Region of the National Speleological Society, alarmed by the accelerating degradation of Virginia's cave resources, asked Representative Bill Axselle of Richmond to introduce legislation into the Virginia General Assembly that would create a commission to study the conservation of cave resources. An amended House Joint Resolution No. 10 was passed, and an eleven-member Commission on the Conservation of Caves was appointed by Governor John Dalton to "study all problems incidental to cave use, protection, and conservation in Virginia." The members of this Commission were John Wilson, Chairman, John Holsinger, Vice-Chairman, Evelyn Bradshaw, Secretary-Treasurer, Robert Anderson, Roy Clark, Wayne Clark, Robert Custard, Henry T.N. Graves, John Kettlewell, Philip Lucas, and Virginia Tipton.

In December 1978, the commission completed its study and submitted its findings to the governor and general assembly (Commonwealth of Virginia, 1979). This report documented the rapid deterioration of Virginia's caves as geological, archaeological, biological, recreational, and educational resources. The commission recommended that an inventory of archaeological resources in Virginia caves be made, a permanent commission be created, and a new Cave Protection Act giving broader protection to cave resources be enacted (Department of Conservation and Economic Development, 1979).

The 1979 session of the general assembly, responding to the recommendations of the Commission on the Conservation of Caves, created the Virginia Cave Commission and enacted a new comprehensive Cave Protection Act with two basic objectives. The first was to protect Virginia cave resources from vandalism and degradation; the second, to protect the cave owner's property interests. Violations of the act were designated as class-3 misdemeanors, punishable by a fine of up to five hundred dollars.

Under the provisions of this new law it is illegal to remove, mar, or otherwise disturb any natural mineral formation or sedimentary deposit in any cave without the owner's express, prior, written permission. (The 1975 act had not required that permission be obtained in writing.) Although collection of mineral specimens is not completely prohibited, it was the intent of the commission that future collection be as minimal, selective, and scientific as possible. The act was designed to preserve the beauty of Virginia caves and prevent them from being destroyed by indiscriminate collection or vandalism. It also is illegal to sell, or export for sale, speleothems (mineral formations or deposits found in caves). The general assembly felt that by eliminating the market, much of the incentive for theft would also be eliminated.

The commission's report stressed that caves are unique natural laboratories for the investigation of biological processes. Natural organisms found in caves live in fragile environments where even small man-made disturbances can produce major changes in cave ecosystems. Many of the more than two hundred animal species found in Virginia caves are restricted to small geographic areas, occur in very small populations, and have been placed on the Endangered Species List. The Cave Protection Act, therefore, prohibits disturbing or harming any cave organism.

The pollution of groundwater, as a result of the dumping of garbage, sewage, dead farm animals, and toxic wastes into caves and sinkholes, had been a problem in the state. It now is illegal to dump any litter, waste material, or toxic substance in any cave without the express, prior, written permission of the owner.

The new act protects archaeological resources by requiring a permit from the Virginia Historic Landmarks Commission and written permission from the cave owner to excavate, remove, or disturb any fossils, historical artifacts, or prehistoric animals. It also protects gates, locks, and other barriers designed by the cave owner to prevent or to control access to the cave. It is illegal to break, force, or tamper with these barriers or to remove or deface any sign posted by the owner. The cave owner is also exempted from liability for any injury sustained by others in the cave as long as an admission fee was not charged.

Legislative History of the Virginia Cave Protection Act

A brief summary of the legislative history of the Cave Protection Act through 1985:

House Bill 24 created Section 18.1-175.1 "Damaging Caverns or Caves" on March 2, 1966.

"Damaging Caverns or Caves" was moved to 18.2-142.

House Bill No. 1800, introduced by Representative Axselle to create a Virginia Cave Commission, became law (Title 9 Chapter 24.1 Section 9-152.1 through 152.5) on October 28, 1978.

House Bill No. 1220, introduced by Representative Axselle to create the Virginia Cave Protection Act, became law (Title 10 Chapter 12.2 Section 10-150.11 through 10-150.18) on March 15, 1979. Section 18.2-142 was repealed.

House Bill No. 240, introduced by Representatives Murray, Giesen, Axelle, and Michie, reestablished the Cave Commission and amended its powers and duties (January 21, 1980).

House Bill No. 92, introduced by Murray, Axselle, and Van Yahres changed vandalism, pollution, and the sale of speleothems from a class-3 misdemeanor to a class-1 misdemeanor and added a section on paleontology, 1982. (The penalty for a class-1 misdemeanor is a fine not exceeding $2,500 or confinement in jail not exceeding 12 months or both.)

A name change from Virginia Cave Commission to Virginia Cave Board was effective July 1, 1985.

The Virginia Cave Protection Act was amended several more times, as late as 1989, and now defines the Virginia Cave Board and its powers and duties, provides for permits for excavation and scientific investigations, establishes penalties for vandalism, pollution, disturbances, and sale of speleothems, and reduces the liability of land owners.

Examples of Cave Board Actions

Vandalism

The Virginia Cave Board (née Cave Commission) has been involved in several court cases regarding vandalism and has worked with various communities to protect cave resources. In 1981, local students illegally entered the fenced Barterbrook Spring Cave. The owner had the students arrested, but, rather than go to court, their parents paid for a new fence. (Virginia Cave Commission Minutes, March 29, 1981. Copies of Commission and Board minutes can be obtained from the Virginia Department of Conservation and Recreation, 217 Governor Street, 3rd Floor, Richmond, VA 23219.)

In another case, students from James Madison University who had removed speleothems from Fountain Cave argued in their defense they did not know it was illegal because there was no sign at the cave. They were sentenced to complete a special project at the university to benefit caves, including publication of an article in the JMU newspaper about the new Cave Protection Act and the importance of preventing cave vandalism (Virginia Cave Commission, December 6, 1981).

In 1984, a man was apprehended inside Perkins Cave after he had damaged the gate and entered the cave without authorization. The judge sentenced him to ten hours of public service installing cave protection signs in lieu of a $100 fine (Virginia Cave Commission, June 2, 1984).

In southwestern Virginia, two students allegedly entered a cave to collect speleothems for a science project. They saw a sign that said in large letters, "THIS CAVE is protected." They left, found another cave without a sign, and collected their speleothems. Again, the judge sentenced them to community service. As a result of this case, the

Virginia Cave Commission changed its signs from "THIS CAVE in protected" to read "ALL CAVES are protected" (Virginia Cave Commission, June 2, 1984).

In the fall of 1985, there was a break-in at Madison's Saltpetre Cave in Augusta County. The vandals were identified, the cave owner prosecuted, and they were sentenced to twenty hours of community service (Virginia Cave Board, May 10, 1986).

Commercial caves have also had their share of vandalism. In 1981, Grand Caverns was closed for two weeks when six Boy Scouts, camping nearby with their troop from Silver Spring, Maryland, vandalized the cave. They were arrested, released on $500 bond, and sentenced to community service after their hearing. Massanutten Caverns had its steel-plated door smashed in, but there were no arrests (Collins, 1981).

Many, but not all, of the cases involved lack of vandalism-deterrent signage. Nevertheless, out of 370 significant caves in Virginia, only 100 have cave-protection warning signs today.

PROJECT REVIEW

Between 1981 and 1984, the Commission became involved in a long, drawn-out discourse with the Town of Grottoes, via letters, meetings and hearings, regarding a proposed water tank and pipeline on Cave Hill. Planned blasting and other construction activities, as well as possible future failure of the water tank, raised many concerns, including potential damage to speleothems in Grand Caverns, collapse of cavities, pollution and siltation of the Cave Hill Aquifer, or changes in groundwater flow. The number-one concern was the potential impact on the Madison Cave Isopod, *Antrolana lira*, which was on the Endangered Species List of the Fish and Wildlife Service. During this same review period, a sinkhole was inadvertently filled and Federal funding was delayed. Additional studies were conducted, and as a result, all concerns of the Cave Commission were addressed by the town and their engineers and the water tank was built (Shetterly, 1983; 1984).

DEED INTERPRETATION

In 1985, a group of students and their professor from Lincoln Memorial University (LMU) in Tennessee were photographed removing speleothems at Cudjo's Cave (*Home Daily of the Cumberlands*, Middlesboro, Kentucky, November 18, 1985), resulting in a lengthy legal discussion over exceptions in the property deed. On April 3, 1947, the property was deeded from LMU to the Commonwealth of Virginia (Commonwealth), with two relevant exceptions. The first reserved for the grantor (LMU) a parcel of about ten acres that included the entrance to Cudjo's Cave. The second exception reserved for LMU the exclusive right to operate and use Cudjo's Cave, even though the cave extended beyond the ten-acre parcel reserved by the first exception. By a second deed, on May 4, 1950, LMU granted the Commonwealth the ten-acre tract reserved by the first exception to the 1947 deed, and expressly released any further right, title and interest to the cave based on its previous title to the reserved tract. However, in giving up its title to the property, LMU reserved the right to "explore, use, occupy, maintain, develop, operate, and exhibit for profit or otherwise," the caves underlying the tract. On December 1, 1953, the Commonwealth deeded the property, subject to LMU's easement, to the United States for inclusion in Cumberland Gap National Historic Park. The easement reserved by LMU was conditioned expressly upon the fact the property was to be included in the National Historic Park. LMU agreed to the 1950 deed as a condition of the exclusive right to operate and exhibit the cave. The Commonwealth's 1953 deed to the U.S. included the easement reserved by LMU.

In letters received by the Virginia Cave Board, one attorney stated,

> Applying the ordinary rules of construction to the lease terms in question, it appears that the intent of the parties was to transfer all title and rights to the cave to the Commonwealth, subject to the easement reserved to LMU to explore, use, occupy, maintain, develop, operate and exhibit the cave.
>
> The easement, in turn, is limited by the language requiring compliance with all National Park Service (NPS) requirements and regulations, as well as by language indicating a clear intent that the cave be used in a manner consistent with park objectives. Reading the terms together, the lease ensures that LMU's exclusive rights, as set out therein, are not to be barred by the fact that the cave is on National Park property (e.g., LMU does not have to allow public access, cannot be prevented from entering or using the cave, and need not compete with other concessionaires for the privilege of showing the cave for profit). They cannot, however, undertake those activities in a way that would damage, destroy or deface the caves in a manner contrary to park regulations.
>
> This is the only interpretation consistent with the fact that the NPS owns the cave, while LMU owns only an easement giving it certain access and use rights. This is not a typical holding case where the original owner retains the fee or other estate in the land. There is nothing in the language of the easement indicating the property owner intended to allow the easement holder to damage or deface its property, and courts will not construe an easement in such fashion absent express language.
>
> In sum, the deeds construed together require LMU to comply with all NPS cave protection regulations, including 36 C.F.R. § 2.l(a)(1)(iv), which prohibits possessing, destroying, impairing, defacing, removing or disturbing any cave formation or part thereof. The National Park Service has full authority to enforce those regulations against LMU consistent with the term of the deed. (Personal Correspondence from Timothy G. Hayes, Thomas and Fiske, P.C., March 25, 1986.)

Another attorney, Linda Loomis, wrote, "In this opinion, if the language of the deeds is controlling, the National Park Service does not have the authority to prevent resource removal. In brief, the deed granting the land to the United States Government references specific exemptions that benefit the grantor and former grantors of the property. Among those benefits is the use and exploitation of the cave. The language is broad enough to be interpreted to allow the removal of speleothems."

(Personal Correspondence from Linda Loomis, National Parks and Conservation Association, February 24, 1986.)

It was clear to the Virginia Cave Board that, with exceptions, land deeds to the United States needed to be clearly understood before the Federal Government and the Commonwealth of Virginia could consider enforcement actions (Virginia Cave Board, January 18, 1986).

Native American Burial Sites

Bull Thistle Cave, the best preserved example of a burial-pit cave known in southwestern Virginia and listed in the National Register of Historic Places, was used by Native Americans for the burial of their dead during the Late Woodland Period (A.D. 900–1700) and contained archaeological remains in an excellent state of preservation. At least eleven individuals were represented among the bones exposed on the surface of the cave. The structure of the undisturbed talus cone below the pit entrance suggested more human remains and artifacts were probably buried there. Further scientific study of the cave deposits yielded important new information about the paleo-demographic characteristics and cultural practices of the Virginia Native Americans. The removal of remains from the cave was covered under Section 10.1-1003 in the archaeological section of the act, which resulted in the development of a management plan (Virginia Cave Board, September 20, 1986).

In August 2001, there was a break-in at Adams Cave, and human remains were removed. Local students were apprehended, and each was sentenced to ten hours community service (Virginia Cave Board, September 8, 2001).

In 2002, Native American remains removed for research purposes from Bone Cave in Lee County were re-interred at a site in Amherst County on land owned by the Monacan Indian Nation. Unexcavated remains are still in the significant and protected Bone Cave (Virginia Cave Board, November 23, 2002).

Endangered Species

In 1990, it was discovered that the Thompson Cedar Creek and Batie Creek watersheds in the Cedars Karst Area in Lee County had been polluted for more than three years with sawdust debris dumped by the Russell Lumber Company. The sawdust had accumulated in immense ridges 20 to 30 feet deep and 200 feet across, and acres of forest were covered with it. Surface water had become a black, viscous flow that was sinking into Thompson Cedar Creek and eventually the Powell River.

The caves of Lee County host a diverse and abundant fauna of cave-adapted invertebrates. Among them is Thompson Cedar Cave, where in the 1960s cave biologists John Holsinger and David Culver first discovered the Lee County Cave isopod, *Lirceus usdagalun* (Virginia Cave Board, June 9, 1990). Batie Creek was included on EPA's 303(d) list of impaired streams, and through the combined efforts of the Virginia Department of Conservation and Recreation, the Virginia Department of Mines, Minerals and Energy, the U.S. Fish and Wildlife Service, the Tennessee Valley Authority, the Curtis Russell Lumber Company, and the Cave Conservancy of the Virginias a recovery plan was developed. By 2005, the restoration of the Batie Creek watershed was complete. Accumulations of sawdust that had generated toxic leachate were removed and mixed with lime and fertilizer as a beneficial soil additive on nearby coal-mine-reclamation projects. Dissolved oxygen levels that had been near zero returned to normal levels. The Lee County isopod, *Lirceus usdagalun*, listed as endangered due to its extirpation from the cave in the late 1980's, recovered, although not to pre-impairment levels (Virginia Cave Board, March 19, 2005).

A new airport and a prison were planned for Lee County. These projects impacted significant biological resources, including an endemic millipede, several rare cave invertebrates, and rare plants, including a new species of clover found only in Virginia. The Virginia Cave Board wrote letters to the County Board and held meetings, resulting in the airport expansion but not the construction of the prison.

In June 1993, the board recommended a change in the proposed right-of-way for Route 58 in the vicinity of Young-Fugate Cave. With over 5,800 feet of surveyed passages, this cave is considered to be biologically, geologically, and hydrologically significant. A number of rare cave invertebrates, including the trechine beetle *Pseudanophthalmus holsingeri*, a dipluran *Litocampa cooki*, two aquatic crustaceans, and the gray bat, *Myotis grisescens*, have been noted there. The proposed right-of-way could well have led to future subsidence and eventual collapse of the roadbed into the subterranean passages. The result of numerous meetings was a rerouting of the right-of-way (Virginia Cave Board, June 19, 1993).

In 2007, Rocky Hollow Cave, located on the west slope of Powell Mountain and home to the endangered Indiana bat, *Myotis sodalis*, was vandalized. A gate installed at the cave entrance by the U.S. Forest Service in the late 1990s to protect hibernating Indiana-bat populations was breeched via a tunnel near the western end of the cave entrance. Inside were numerous patches of graffiti, including a date and several names in pink, white, and orange paint. Assuming the May 28, 2006, graffiti date was correct, it is unlikely the visit by the vandals caused any disruption or negative impact to the Indiana bat, as it was well past the winter hibernation period. Nevertheless, the Virginia Cave Board requested the assistance of the Wise County Sheriff in apprehending the perpetrators. One individual was apprehended, and based on the recommendation of the board, was ordered by the judge to clean up the graffiti, which resulted in ten hours of community service. Of note, when undertaking an enforcement action, the statute of limitations must always be considered. In Virginia, this statute is one year (Virginia Cave Board, March 24, 2007).

Table 1. State-owned caves.

State Agency	Number of Caves Owned
Department of Transportation	75
Department of Game and Inland Fisheries	53
Natural Tunnel State Park	9
Department of Conservation and Recreation	6
Commonwealth of Virginia	5
New River Trail State Park	4
New Market Battlefield State Historic Park	3
Virginia Polytechnic Institute	1
Total	156

Source: Virginia Speleological Survey Data Files, December 2007

OTHER ACTIONS

The Virginia Speleological Survey, on behalf of the Virginia Cave Board, now gathers and maintains an informational and survey database on Virginia caves.

The board proposed the Virginia big-eared bat as an ideal candidate, because of its name and its status as a federally endangered species, for educating Virginian residents about caves and the animals that inhabit them. Virginia Delegate Jackie T. Stump filed House Bill No. 2579 on January 12, 2005. On February 26, after being approved in both the House of Delegates and the Senate, the bill was signed by the Speaker of the House and the President of the Senate. On March 22, 2005, Governor Marc Warner signed the legislation designating the Virginia big-eared bat (*Corynorhinus townsendii virginianus*) as the official state bat of the Commonwealth of Virginia, effective July 1, 2005. The cave board continues to work with various state departments on environmental reviews and has participated in discussions on state regulations regarding caves and karst and the importance of their protection. The board has also worked with the Department of Historic Resources in granting permits for excavation and removal of archaeological, paleontological, prehistoric, and historic features in caves; worked with the Virginia Department of Transportation, the largest manager of state-owned caves, on the widening of state highways and the gating of significant caves; and worked with the Virginia Natural Area Program and Department of Game and Inland Fisheries on preparing management plans for state-owned caves (see Table 1). Several new species have been identified and listed on both the federal and state endangered-species lists. Board members Dr. John Holsinger and Dr. David Culver reported that the Department of Conservation and Recreation's Natural Heritage Program has recommended to the Virginia Department of Agriculture and Consumer Services that two species of cave beetle be added to the Virginia Endangered Species List under the Virginia Endangered Plant and Insect Act of 1979. The board has suggested that the common name of the mud-dwelling cave beetle be changed to Maddens Cave beetle, and the common name of the thin-neck cave beetle be changed to Hupp's Hill cave beetle (Virginia Cave Board, September 16, 2006).

Ed Wallingford, Virginia Department of Transportation Hazardous Materials Program Manager, and Mark Nelson, the EPA Region III Underground Injection Control (UIC) Program Manager, concurred in correspondence with Department of Conservation and Recreation staff, that only sinkholes whose throats had been significantly modified to accept stormwater runoff were to be registered as Class V Injection wells by the EPA. However, in further conversations with the EPA, UIC staff revealed that Region IV employed a more inclusive definition of Class V injection wells to include any sinkhole to which runoff from converted land has been diverted (Virginia Cave Board, December 4, 2004).

State funding continues to be available for the various Virginia Agricultural Best Management Practices (BMPs). Efforts are underway to inform Virginia landowners about available cost-share and tax-credit opportunities through the programs. This cost-share program is funded through the State Water Quality Improvement Act and is administered by the Virginia Department of Conservation and Recreation through local Soil and Water Conservation Districts. The Agricultural Sinkhole Protection BMP (WQ-11) will pay 75% of the cost of debris removal up to $2,500. In addition to the cost-share payment, the program allows for a tax credit of "25% of the total eligible cost, not to exceed $17,500." Sinkholes with streams that flow into them are given priority under the program (Fagan and Orndorff, 2002; Virginia Cave Board minutes, December 3, 2005).

CONCLUSION

It has been 30 years since the 1979 act became law, and the importance of the confidentiality of significant cave locations and the difficulty of apprehending vandals continue to be addressed by the Virginia Cave Board. The prosecution of vandals demonstrates the inadequacy of current penalties. The Virginia Cave Protection Act should be amended to allow prosecutors to choose between a misdemeanor and a felony charge, similar to the Federal Cave Resource Protection Act. More prosecutions and harsher penalties will invariably serve as a deterrent to future potential vandals (Kramer, 2003).

Virginia cave resources continue to be threatened by vandalism, pollution, and poorly planned development. Unfortunately, many cave owners remain unaware of the immense scientific, historic, and economic value of the unique nonrenewable cave resources they own. As public interest in outdoor recreation continues to grow and land development accelerates, increased pressures will be put on

Virginia's limited and fragile cave resources. The Virginia Cave Board is committed to safeguarding the unique educational, recreational, scientific, historic, and economic values of Virginia cave and karst areas. A board composed of concerned citizens, working in conjunction with other agencies of the commonwealth, appears to be the most effective vehicle for focusing the attention of both government and the public on this important conservation goal.

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ON THE TEMPORAL BEHAVIOR OF KARST AQUIFERS, ZAGROS REGION, IRAN: A GEOSTATISTICAL APPROACH

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Abstract: A geostatistical approach was used to study temporal structures in a time series of discharge and electrical conductivity (EC) in 15 karst springs from the Zagros mountain range, Iran. Two types of temporal behaviors, a periodic structure and nugget effect, plus one or two temporal structures, were identified and interpreted. These correspond to characteristics of karst systems, such as the catchment area, percent of conduit flow, and general degree of karst development. Springs were grouped into three categories based on their ranges (e.g., residence time) obtained by variogram analysis. The first group of springs include those that present the same temporal behaviour in variograms of discharge and EC. These springs are characterized by generally constant EC with increasing discharge suggesting the existence of a large underground reservoir. The second group of springs are those with varying temporal periodic behavior in variograms of discharge and EC. Positive correlation between discharge and EC values is the main characteristic of these springs and is interpreted to result from a piston-flow system in poorly developed karst aquifers. The third group of springs includes those that exhibit different temporal behaviors when compared with the periodic and non-periodic variograms. This group exhibits a negative correlation in scatterplots of discharge versus EC values suggesting a well-developed solution-conduit system that facilitates rapid response of the karst system to precipitation events. This study's results document the role of variogram analysis in delineating temporal structures of spring behaviors by means of time series of discharge and EC. Variogram analysis can be considered as a valuable tool for hydrogeological investigations in karstic terranes.

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Introduction

Karst groundwater is a major water resource in many regions of some countries such as China, Turkey, Iran, the United States, etc. Karstic-carbonate formations cover about 11% of the land area in Iran (185,000 km^2) and 55% of the Zagros Region (Raeisi, 2004). Carbonate rocks become karstic aquifers by dissolution processes, typically referred to as karstification. Karstification creates a significant heterogeneity of the permeability within the aquifer. Karst development processes and several methods intended to characterize karst systems have been extensively presented in Palmer (2007), Ford and Williams (2007), Bakalowicz (2005), White (1988), Gillieson (1996), Mangin (1994), and Milanović (1981).

The hydrogeological study of karst aquifers is particularly difficult because of the complex and heterogeneous character of the karstic massif and the limited number of available wells that permit hydrogeological observation (Padilla and Pulido-Bosch, 1995; Panagopoulos and Lambrakis, 2006; Mohammadi and Raeisi, 2007). As a result, studies on the function and hydrodynamic behaviour of karst aquifers are focused on the analysis of the characteristics of karst springs. Two commonly measured parameters are discharge and electrical conductivity (EC) that are often presented as a time series. These parameters are widely used by karst researchers because these parameters provide reliable results regarding karst-aquifer characteristics and are relatively easy and inexpensive to collect, especially in less developed area such as the Zagros Region in Iran. Time-series variations of physico-chemical parameters of springs have been used by many authors for assessing hydrogeochemical aspects of karst aquifers (e.g., Hess and White, 1988; Scanlon and Thraikill, 1987; Raeisi and Karami, 1997; Lopez-Chicano et al., 2001; Desmarais and Rojstaczer, 2002; Karimi et al., 2005a; and Mohammadi et al., 2007). Generally, these authors focused on temporal variations of discharge and chemical parameters caused by a heavy precipitation event in terms of internal and external factors involving the karst system studied. Many authors (e.g., Mangin, 1984; Moore, 1992; Padilla and Pulido-Bosch, 1995; Larocque et al., 1998; Kovacs et al., 2005; and Manga, 1999) applied correlation and spectral analysis on the time series of springs to extract further information about time lag, periodicity, and residence time.

Variogram analyses are extensively used in hydrology (e.g., Bacchi and Kottegoda, 1995; Holawe and Dutter,

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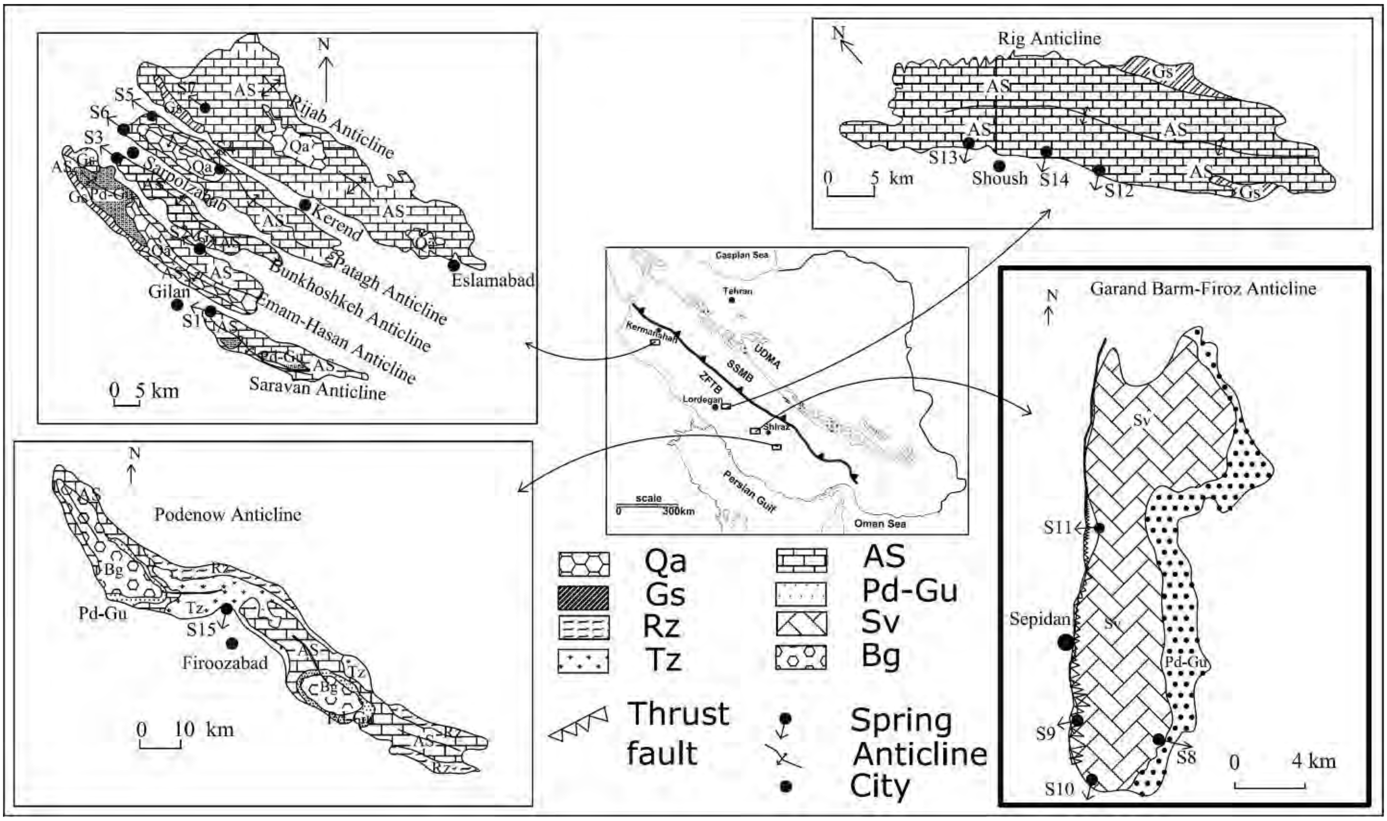

Figure 1. Simplified geological map of the selected aquifers in the Zagros mountain range. See Table 1 for a description of map symbols of depicted geological units.

1999; Berne et al., 2004; Kyriakidis et al., 2004; and Buytaert et al., 2006), but its contribution to a time series of physico-chemical parameters of springs is rare (e.g., Rouhani and Wackernagel, 1990; Goovaerts et al., 1993; Kolovos et al., 2004; and Silliman et al., 2007). Rouhani and Wackernagel (1990) applied variogram analysis to monthly piezometric data at 16 observation wells in a basin south of Paris, France. Two temporal structures were determined by the variogram analysis, including the 12-month seasonal and the 12-year climatic cycles. Multivariate geostatistical analysis was applied to spring-water-solute contents measured in 86 springs situated in Belgium by Goovarets et al. (1993).

In this study, we use variogram analyses to evaluate the time-series data of discharge and EC measurements obtained from 15 springs located in the Zagros Region of southern Iran (Fig. 1). The study is aimed at improving our understanding of the temporal dynamics of the karst system in the area. The objectives of this study, with the expectation that a better understanding of karstic aquifers may be obtained, are (1) determination of the temporal structures in the time series of discharge and EC, (2) exploration of new information on the characteristics of the karst systems, and (3) evaluation of the potential of variogram analysis for studying karst development.

Geological and Hydrogeological Settings

The Zagros Region is located in south-west Iran. The climate is semi-arid in the uplands and arid in the lowlands (south of Iran). Precipitation exhibits large spatial and temporal variability with a mean annual precipitation in the Zagros region of about 450 mm, with a range of 150 to 750 mm. Runoff is an aggregation of several hydrological river basins that discharge to the Persian Gulf. The high elevation areas of the Zagros Region are the zones where the rivers originate; the main rivers flowing in the Zagros region are Karoun River, Dae River, Karkhe River, Hele River, Mond River and Zohre River. Karoun River is the Zagros' most important river with the highest amount of flow and many tributaries in upstream sub-watersheds, the largest of which is the Dez River. Karoun River is a major source of water for 4.5 million inhabitants in the south of Iran (KWPA, 2009).

Iran is geologically a part of the Alpine-Himalayan orogenic belt. The Zagros mountain range extends from the northwest to southeast of Iran and consists of three NW-SE trending parallel zones (Fig. 1): (1) the Urumieh-Dokhtar Magmatic Assemblage (UDMA); (2) the Sanandaj-Sirjan Metamorphic Belt (SSMB); and (3) the Zagros Fold and Thrust Belt (ZFTB). The ZFTB is the study area of this paper.

Table 1. Stratigraphic column of geological units depicted in Figure 1.

Symbol on Figure 1	Geologic Unit	Geologic Age	Composition
Qa	Recent alluvium	…	…
Gs	Gachsaran Formation	Miocene	Gypsum, marl and salt
Rz	Razak Formation	Lower Miocene	Silty marl to silty limestone with interbedded layers of gypsum
Tz	Transition zone	Lower Miocene	Transition between Asmari and Razak Formations
As	Asmari Formation	Oligocene-Miocene	Limestone, dolomite
Pd-Gu	Pabdeh-Gurpi Formation	Paleocene-Oligocene	Marl, shale and marly limestone
Sv	Sarvak Formation	Upper Paleocene	Limestone
Bg	Bangestan Group	lower Paleocene	Limestone, shale, marl

The stratigraphy and structural framework of the ZFTB were studied in detail by James and Wynd (1965), Stocklin and Setudehnia (1977) and Alavi (2004, 2007). The ZFTB is about 12-km thick and consists mainly of limestone, marl, gypsum, sandstone and conglomerate. Since the Miocene age, it has been folded into a series of huge anticlines and synclines. Most of the carbonate rock outcrops are of Cretaceous and Tertiary age. The most important karst features in the ZFTB are karren, grikes, and springs, and to a lesser extent, caves and sinkholes. Most of the springs are permanent with a high percentage of spring discharge from base flow. The ZFTB is characterized by a repetition of long and regular anticlinal and synclinal folds. The anticlines normally form mountain ridges of limestone and the synclines normally form valleys and plains. Most of the karst formations in the ZFTB are sandwiched between two impermeable formations that form broad highland independent aquifers (Raeisi, 2004; Raeisi and Laumanns, 2003). Several anticlines from different parts of the ZFTB were selected for this study. Simplified geological maps of these anticlines are presented in Figure 1 and a description of the geological units in Table 1.

Several Anticlines in the Alvand River Basin

The Alvand river basin comprises seven main anticlines (Karimi, 2003), five of which are considered in this study, and include the Saravan, Emam-Hasan, Bunkhoshkeh, Patagh, and Rijab anticlines (Fig. 1). These anticlines are located ~150 km west of the Kermanshah located in the south-western part of Iran (Fig. 1), follow a northwest-southeast trend, and are mainly composed of the Asmari limestone. The geologic formations in this area are, from youngest to oldest: 1) recent alluvium; 2) Gachsaran gypsum and marl; 3) Asmari dolomite and limestone; and 4) Pabdeh-Gurpi marl and shale with interbedded, thin marly limestone (Fig. 1). The core of the anticlines is composed of the Asmari formation and is situated between the impermeable upper Gachsaran and lower Pabdeh-Gurpi formations (Karimi et al., 2005b). The dense and thick bedded Asmari limestone in the anticlines has numerous joints and fractures with limited solution features and small shelter caves (Karimi et al. 2005b). Generally, the southern flanks are hydraulically disconnected in most parts of the anticlines (Karimi, 2003), except in the plunge areas. Groundwater from the above aquifers discharge from seven main springs including Gilan (S1), Golin (S2), Sarabgarm (S3), Marab (S4), Piran (S5), Gharabolagh (S6) and Rijab (S7) Springs (Fig. 1 and Table 2). There is no hydraulic connection between these springs except the Marab (S4), Piran (S5) and Gharabolagh (S6) which emerge from Patagh Anticline. However, the catchment areas of the springs were mapped without any overlap in their areas (Karimi, 2003).

The Barm-Firooz and Gar Anticlines

The Barm-Firooz and Gar Anticlines are located 80 km northwest of Shiraz on a general northwest trend of the Zagros mountain range. The cores of the anticlines are comprised of the calcareous Sarvak formation, which is overlain by impermeable Pabdeh-Gurpi formations (Fig. 1). The most important tectonic feature in this area is a northwest trending thrust fault (Fig. 1). Groundwater from the Sarvak aquifer discharges mainly from Sheshpir (S8), Berghan (S9), Morikosh (S10), and Tangkelagari (S11) Springs (Fig. 1). The most important karst features in the catchment area of Sheshpir Spring (S8) is the presence of 255 sinkholes (Raeisi and Karami, 1997). Several normal faults and one thrust fault have resulted in an extensive brecciated zone in the catchment area of Berghan Spring (S9). No sinkholes or caves are present in the catchment area of Berghan Spring (S9). It seems that karst is developed as a network of interconnected small fissures and pores (Raeisi and Karami, 1997) and with minimal karstification. No hydraulic connection between the catchment area of Sheshpir spring (S8) and three other springs has been reported.

The Rig Anticline

The Rig anticline is located in Southern Iran near the city of Lordegan. The main formations in this area are the Gachsaran (Miocene), Asmari (Oligocene-Miocene), and

Table 2. Data sets characteristics used in the analysis.

Spring Name	Code on Figure 1	Anticline	Sampling Interval	Sampling Period	Data Reference
Gilan	S1	Saravan	Weekly[a]	09/'00–09/'01	Karimi (2003)
Golin	S2	Emam-Hasan	Weekly[a]	09/'00–09/'01	Karimi (2003)
Sarabgarm	S3	Bunkhoshkeh	Weekly[a]	09/'00–09/'01	Karimi (2003)
Marab	S4	Patagh	Weekly[a]	09/'00–09/'01	Karimi (2003)
Piran	S5	Patagh	Weekly[a]	09/'00–09/'01	Karimi (2003)
Gharabolagh	S6	Patagh	Weekly[a]	09/'00–09/'01	Karimi (2003)
Rijab	S7	Rijab	Weekly[a]	09/'00–09/'01	Karimi (2003)
Sheshpir	S8	Gar and Barm-Firoz	Daily[b]	03/'90–11/'91	Karami (1993); Pezeshkpoor (1991)
Berghan	S9	Gar and Barm-Firoz	20 days	03/'90–11/'91	Karami (1993); Pezeshkpoor (1991)
Morikosh	S10	Gar and Barm-Firoz	20 days	03/90–11/'91	Karami (1993); Pezeshkpoor (1991)
Tangkelagari	S11	Gar and Barm-Firoz	20 days	03/'90–11/'91	Karami (1993); Pezeshkpoor (1991)
Atashgah	S12	Rig	Weekly[a]	05/'02–09/'03	Keshavarz (2003)
Shosh	S13	Rig	Weekly[a]	05/'02–09/'03	Keshavarz (2003)
Enakak	S14	Rig	Weekly[a]	05/'02–09/'03	Keshavarz (2003)
Ghomp	S15	Podenow	Daily[b]	04/'96–09/'97	Karimi (1998)

[a] One week during rainy season and one or two weeks during dry season.
[b] Daily during rainy season and two weeks during dry season.

Pbdeh-Gurpi (Paleocene-Oligocene) Formations (Fig. 1). Rig Anticline is a box fold that mainly consists of the karstic Asmari Formation (Keshavarz, 2003). Numerous joint sets are observed in the Asmari Formation. There appears to be no concentrated recharge points, such as sinkholes or sinking streams, in this aquifer. The Atashgah Spring (S12), having a mean discharge rate of about 900 L s^{-1}, is the largest spring originating from the Rig Anticline (Fig. 1). Two other large springs, Shosh (S13) and Enakak (S14) Springs, emerge from the Rig anticline (Fig. 1).

The Podenow Anticline

The Podenow anticline is located south of Shiraz, Iran. The geological formations in decreasing order of age consist of the Bangestan group (lower Palaeocene), Pabdeh-Gurpi (Palaeocene-Oligocene), Asmari (Oligocene-Miocene), Transition zone and Razak (Lower Miocene), as shown in Figure 1. The core of the Podenow anticline is composed of the limestone Asmari Formation which is sandwiched between the two impermeable Pabdeh-Gurpi (marl, shale and marly limestone) and Razak (silty marl to silty limestone with interbedded layers of gypsum) Formations (Fig. 1). This anticline is divided into eastern, central, and western parts based on the orientations of the anticline. The eastern and western sections follow the general northwestern trend of the Zagros mountain range (Karimi et al., 2005a). The largest spring on the southern flank is Ghomp Spring (S15 in Fig. 1).

Materials and Methods

Data Collection

Electrical conductivity and discharge measurements from the 15 springs from the Zagros mountain range were used for this study. Sampling intervals and sampling periods for each spring are presented in Table 2. Electrical conductivity was measured by a portable ELE EC-meter in the field immediately after sampling. Spring discharge was measured by current-meter or triangular weir related to spring discharge and field conditions. Hydrogeological characteristics of the studied springs are presented in Table 3.

Data Analysis

Exploratory data analysis and variogram analysis were used for database analyses. Use of multiple data analyses techniques provides greater insight into the information contained in a database (Farnham et al., 2000; Silliman et al., 2007).

Exploratory-Data Analysis

Exploratory data analysis is a purely descriptive part of the study that allows for a good preliminary assessment of the collected data (Isaaks and Srivastava, 1989). There is no single statistical tool as powerful as a plot of the data (Chambers et al., 1983). The distribution of continuous variables can be depicted by a histogram with the range of data values discretized into a specific number of classes of equal width and the relative proportion of data within each

Table 3. Hydrogeological characteristics of the studied springs.

Spring Name	Elevation (m)	Watershed Area (km^2)	Annual Precipitation (mm)	Percent of Conduit Flow	Ratio of Recession Coeff. (a_1/a_2)	Ratio of Max. to Min. Discharges (Q_{max}/Q_{min})
Gilan	1413	110	473	8	0.57	1.58
Golin	1526	68.8	492	3.7	1	1.12
Sarabgarm	1191	204.1	454	5.4	75	1.35
Marab	1879	42	515	35	26.7	5.25
Piran	1176	26.7	460	15	1	1.39
Gharabolagh	1576	56.8	515	6.9	4.5	1.36
Rijab	1874	221	552	35	3.08	7.58
Sheshpir	2335	81	1334	24	11.5	4.56
Berghan	2145	19	798	23.7	2	3.92
Morikosh	2450	4.3	1122	31	2.7	15.5
Tangkelagari	2120	4.47	985	28	2.5	15.2
Atashgah	1710	62	930	13.5	4.4	1.91
Shosh	1500	18.2	910	21	2	5.69
Enakak	1750	5	890	25	2.2	7.6
Ghomp	1350	114.2	400.7	29.5	4.4	3.03

class (e.g., frequency) by the height of bars (Goovaerts, 1997). Important features of a distribution are its central tendency and measure of its spread and symmetry. The relationship between pairs of variables can be depicted in a scatterplot, which is the simplest and probably most informative method for comparing data pairs (Deutsch and Journel, 1992). The correlation coefficient is mostly used as a measure of bivariate relationships. Here, exploratory data analyses include histograms and probability plots of the discharge and EC data series for each spring and relationships among pairs of discharge and EC data as scatterplots. The data plots were developed using Statistica Software, Release 6 (StatSoft, 2001).

Variogram Analysis

The variogram approach is extensively used in geological and environmental sciences to assess the characteristics of spatially or temporarily distributed data (e.g., Isaaks and Srivastava, 1989; Goovaerts, 1997; Webster and Oliver, 2001). The variogram measures the spatial and/or temporal behavior of a variable of interest (Deutsch and Journel, 1992). It is easy to interpret the time axis as the location coordinate in the variogram analysis (Holawe and Dutter, 1999). Many papers in various disciplines have been published using variogram methods for different regionalized variables at different time scales of interest (e.g., Holawe and Dutter, 1999; Berne et al., 2004; Buytaert et al., 2006). Variogram modeling and analysis was accomplished using the program, VESPAR (Minasny et al., 2005).

Assuming the studied time series of observations is a realization of a random function Z, so that $z(t)$, $t = 1,2,3,\ldots,m$, where $z(t)$ refers to observed values of discharge or EC at time t and m is the length of sampling period. Given two times, t and $t + h$ inside the period of temporal attribute $z(t)$, the experimental variogram is a measure of one half the mean square error produced by assigning the value $z(t + h)$ to the value $z(t)$, as follow:

$$\gamma(h) = \frac{1}{2N(h)} \sum_{i=1}^{N(h)} [z(t+h) - z(t)]^2 \tag{1}$$

where $N(h)$ is the number of pairs of observations for a time separation, h. The shape of semivariograms is quantified by the behavior of the variogram at the origin (nugget effect), range, and sill.

The nugget effect is a measure of the variability of a variable within small time lengths. Normally the nugget effect is seen as a consequence of the limited number of observations with arbitrarily small time periods (Holawe and Dutter, 1999). Smaller nugget values translate into higher values of influence of small time lags. Therefore, nugget values can be interpreted as altering variables that can play a special role in a simulation model (Holawe and Dutter, 1999). The range, in the case of time dependence, is a measure related to the length of influence of a variable (Holawe and Dutter, 1999). The sill is a value of the covariance that becomes zero when the variogram reaches a constant value. This total sill is equal to the basic variance of the variable. Therefore, the sill is an indicator for the variance in the data field and, in the case of a time series, a measure of the temporal variability.

In order to describe the variogram structure, it is necessary to fit a model to the experimental variogram. The permissible models are presented by Isaaks and Srivastava (1989). The goodness of fit of different models can be

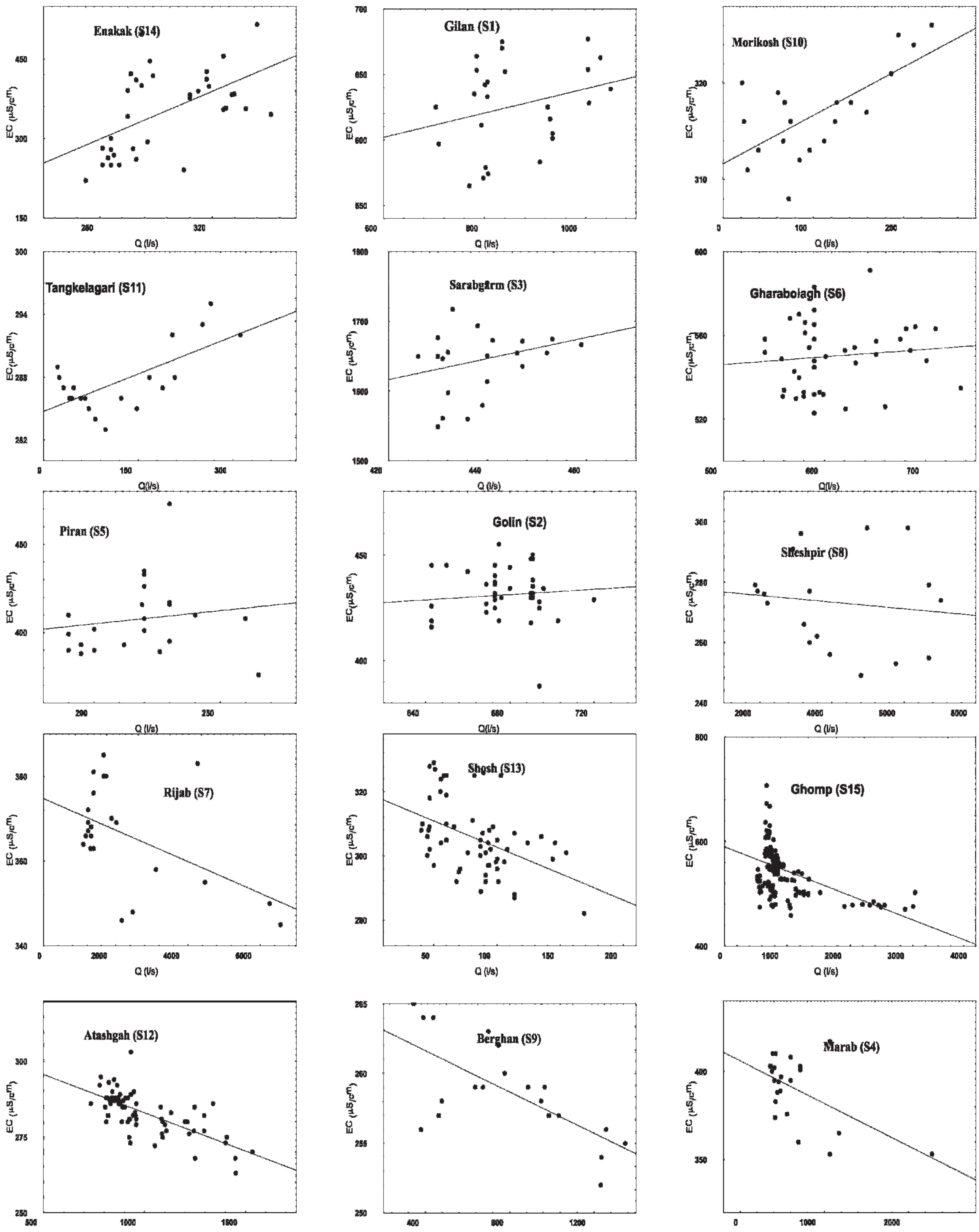

Figure 2. Scatterplots of discharge and EC values for springs.

Table 4. Descriptive statistics of discharge and electrical conductivity for studied springs.

Spring	Number of Samples	Mean	Median	Minimum	Maximum	Range	Std. Dev.	Skewness	C.V %
Gilan									
Discharge (L s^{-1})	40	861.1	820.5	704	1050	346	99.9	0.5	11.6
Elec. Cond. (μ cm^{-1})	40	626.2	630.5	565	677	112	33.8	−0.3	5.4
Golin									
Discharge (L s^{-1})	42	675.4	674	643	720	77	19.1	−0.3	2.8
Elec. Cond. (μ cm^{-1})	42	431.0	430.5	388	455	67	12.2	−0.8	2.8
Sarabgarm									
Discharge (L s^{-1})	41	438.8	438.5	426	459	33	8.8	0.7	2.0
Elec. Cond. (μ cm^{-1})	41	1644.8	1653	1549	1755	206	51.0	−0.2	3.1
Marab									
Discharge (L s^{-1})	42	646.9	481	354	2276	1922	427.1	2.8	66.0
Elec. Cond. (μ cm^{-1})	42	390.4	395	353	417	64	18.6	−0.8	4.8
Piran									
Discharge (L s^{-1})	40	218.3	220	190	265	75	20.9	0.5	9.6
Elec. Cond. (μ cm^{-1})	40	407.6	405	376	473	97	21.1	1.4	5.2
Gharabolagh									
Discharge (L s^{-1})	42	609.8	590	541	735	194	48.5	0.9	8.0
Elec. Cond. (μ cm^{-1})	42	550.1	551.5	523	591	68	15.9	0.3	2.9
Rijab									
Discharge (L s^{-1})	42	2275.0	1670	1096	6560	5464	1597.5	1.8	70.2
Elec. Cond. (μ cm^{-1})	42	366.3	367	345	385	40	11.8	−0.3	3.2
Sheshpeer									
Discharge (L s^{-1})	201	4004.3	3744.2	1493.3	7191.5	5698.2	1351.4	0.5	33.7
Elec. Cond. (μ cm^{-1})	18	273.3	275	249	298	49	15.6	0.2	5.7
Berghan									
Discharge (L s^{-1})	20	796.3	762.5	345	1348	1003	323.7	0.2	40.6
Elec. Cond. (μ cm^{-1})	20	258.7	258.5	252	265	13	3.5	0.2	1.4
Morikosh									
Discharge (L s^{-1})	20	106.0	90.5	21	231	210	64.1	0.5	60.5
Elec. Cond. (μ cm^{-1})	20	317.0	316.5	308	326	18	4.7	0.3	1.5
Tangkelagari									
Discharge (L s^{-1})	21	119.0	82	22	311	289	88.8	0.8	74.6
Elec. Cond. (μ cm^{-1})	21	287.6	287	283	295	12	3.1	1.0	1.1
Atashgah									
Discharge (L s^{-1})	67	998.7	926.61	733	1533	800	194.7	1.1	19.5
Elec. Cond. (μ cm^{-1})	67	283.0	285	263	303	40	7.1	−0.3	2.5
Shosh									
Discharge (L s^{-1})	61	82.6	87	40	169	129	30.4	0.7	36.8
Elec. Cond. (μ cm^{-1})	61	305.5	304	282	329	47	11.2	0.5	3.7
Enakak									
Discharge (L s^{-1})	35	303.4	297	275	341	66	18.6	0.4	6.1
Elec. Cond. (μ cm^{-1})	35	351.3	357	221	515	294	77.8	0.1	22.2
Ghomp									
Discharge (L s^{-1})	218	887.3	771	527	3031	2504	418.9	3.4	47.2
Elec. Cond. (μ cm^{-1})	218	548.2	553	458	707	249	39.1	0.2	7.1

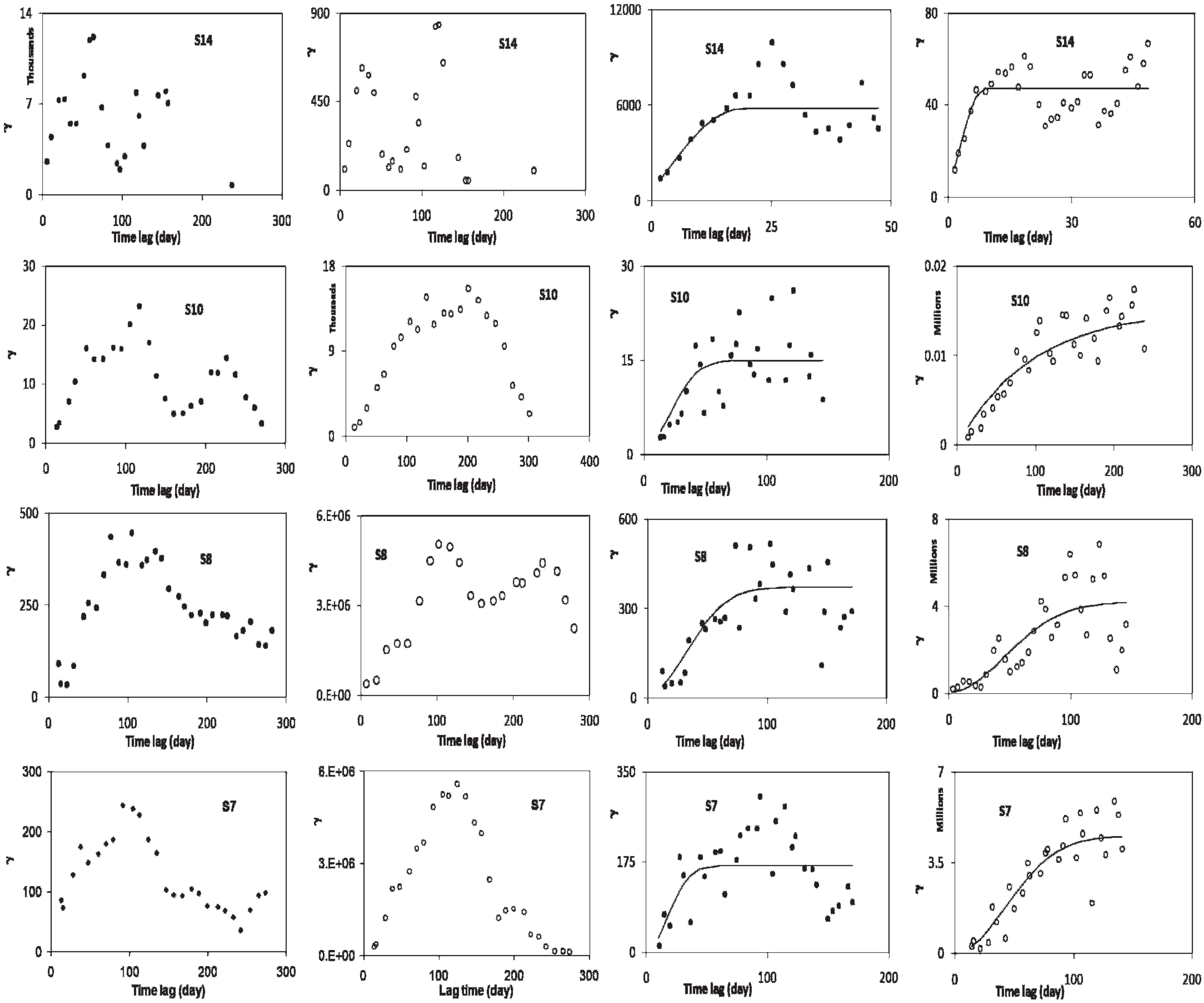

Figure 3. Variograms before (panel 1 and 2 from left) and after removing periodic behavior (panel 3 and 4 from left); solid circles = EC; open circles = discharge.

assumed using the Root Mean Square Error (RMSE), Akaike Information Criteria (AIC) and weighted Sum of Square Error (SSE).

Results

The scatterplots of discharge versus EC and their distributions are shown in Figure 2. Most of the histograms were positively skewed, indicating the presence of large data values for both discharge and EC occurred with low frequency. All histograms were fitted to a lognormal distribution. The descriptive statistics of discharge and EC values for the springs are shown in Table 4.

The correlation among the discharge and EC variables for each spring is shown in Figure 2. Discharge and EC values are positively correlated in Gilan (S1), Sarabgarm (S3), Morikosh (S10), Tangkelagari (S11), and Enakak (S14) Springs, but only partly in Golin (S2), Piran (S5), and Gharabolagh (S6) Springs (Fig. 2). Negative correlation between discharge and EC values is evident for Marab (S4), Rijab (S7), Berghan (S9), Atashgah (S12), Shosh (S13), Ghomp (S15) Springs, but less so in Sheshpir Spring (S8). The strong negative correlation between discharge and EC may be interpreted as an indication of a freshwater recharge signal during the rainy season, which yields a considerable volume of low-EC water. The somewhat constant EC versus discharge values for Piran (S5) and

Table 5. The value of fitting criteria in modeling of periodic variograms after removing the periodicity (the best models are presented by bold numbers).

Spring	Fitting Criteria	Model		
		Exp.	Gau.	Sph.
Sarabgarm				
Discharge	RMSE	23.8	25	**22.7**
	AIC	287	190	**285**
	SSE	17	10	**11**
Elec. Cond.	RMSE	**703**	727	723
	AIC	**484**	486	485
	SSE	**703**	727	723
Marab				
Discharge	RMSE	139,988	**138,152**	139,014
	AIC	791	**790**	790
	SSE	391	**352**	446
Elec. Cond.	RMSE	127	128	**127**
	AIC	384	385	**384**
	SSE	20	21	**20**
Rijab				
Discharge	RMSE	930,689	**784,578**	854,776
	AIC	869	**859**	864
	SSE	172	**31**	102
Elec. Cond.	RMSE	64	**61**	61
	AIC	345	**342**	342
	SSE	125	**86**	124
Sheshpeer				
Discharge	RMSE	1,411,274	**1,267,446**	1,345,383
	AIC	990	**983**	987
	SSE	760	**246**	362
Elec. Cond.	RMSE	102	**92**	93
	AIC	359	**353**	353
	SSE	113	**83**	85
Morikosh				
Discharge	RMSE	1252	**1033**	1167
	AIC	517	**506**	513
	SSE	219	**35**	152
Elec. Cond.	RMSE	11	**10**	10.8
	AIC	190	**186**	187
	SSE	25	**20**	21.6
Tangkelagari				
Discharge	RMSE	**2083**	2011	2008
	AIC	**247**	545	544
	SSE	**38**	11	19.4
Elec. Cond.	RMSE	4.7	**4.6**	4.6
	AIC	172	**170**	170
	SSE	31	**29**	36
Shosh				
Discharge	RMSE	267	**250**	251
	AIC	459	**455**	455
	SSE	53	**43**	55

Table 5. Continued.

Spring	Fitting Criteria	Model		
		Exp.	Gau.	Sph.
Elec. Cond.	RMSE	9.6	9.6	**9.2**
	AIC	252	253	**250**
	SSE	41	40	**32**
Enakak				
Discharge	RMSE	106	79	**9.2**
	AIC	266	253	**250**
	SSE	50	26	**32**
Elec. Cond.	RMSE	1644	1550	**1548**
	AIC	381	378	**378**
	SSE	41	37	**37**
Ghomp				
Discharge	RMSE	640	**377**	569
	AIC	427	**399**	423
	SSE	38	**13**	33
Elec. Cond.	RMSE	87	91	**79**
	AIC	389	392	**384**
	SSE	16	26	**16.8**

Sheshpir (S8) Springs could be caused by (1) a large underground lake that supplies most of the spring discharge water, but also has the capability of damping EC values, or (2) a small or non-rapid recharge component. Alternatively, the generally positive correlation between discharge and EC values may suggest a piston-flow regime in a less developed karst aquifer, which forces water from temporary detention out into a solution conduit for transit to spring outlets during high-flow periods as a result of rising head in the aquifer. In addition, higher mineralized water may be stored in the epikarst, which may contain soluble formations.

Variogram Temporal Structures

The experimental variograms were computed for discharge- and EC-time series data for all springs. Temporal behaviors of springs in terms of variograms is different for discharge- and EC-time series. Two temporal structures are evident in variograms: (1) a periodic behavior and (2) a nugget effect with one or two scales of temporal structures.

Periodicity

Several variograms seem to fluctuate periodically so it is necessary to describe them with a periodic function. One usually observes a variety of temporal periodicities, such as periodic seasonal or annual cycles. The simplest such function is a sine wave (Webster and Oliver, 2001)

$$\begin{aligned} \gamma(h) &= C_1\,\mathrm{Cos}\theta + C_2\,\mathrm{Sin}\theta \\ C_1 &= W\,\mathrm{Cos}\phi \\ C_2 &= W\,\mathrm{Sin}\phi \\ \theta &= \frac{2\pi h}{\omega} \end{aligned} \tag{2}$$

where W, ω and ϕ are the amplitude, length of wave, and phase shift, respectively. The variograms of the discharge and EC values indicate two cyclical trends (Fig. 3). The relative impacts of these cycles may vary from spring to spring, as well as from discharge to EC values. For example, Figure 3 clearly illustrates that the Rijab (S7), Morikosh (S10), and Tangkelagari (S11) Springs have much stronger seasonal components (i.e., effects) than do the Sarabgarm (S3), Sheshpir (S8) and Ghomp (S15) Springs. Discharge variograms display a periodic structure in the Sarabgarm (S3), Rijab (S7), Morikosh (S10), Tangkelagari (S11), and Shosh (S13) Springs while the Marab (S4), Rijab (S7), Sheshpir (S8), Morikosh(S10), and Tangkelagai (S11) Springs reflect periodic processes in EC variograms. Variograms for discharge and EC depict periodicity wavelengths for the springs that range from less than 100 days for Enakak Spring (S14) to more than 316 days for Sarabgarm Spring (S3) and from 82 days in Enakak Spring (S14) to more than 300 days in Sheshpir Spring (S8). These differences may be a result of the catchment areas of the springs. Smaller variogram wavelengths were observed for springs characterized by smaller size of catchment area because of the small distance between the outlet and the hindmost point in the catchment area.

To remove the periodicity and explore the short temporal structures, we use a partial series of discharge and EC values for the springs. These partial series include measured discharge and EC values during a single cyclical period, only. The variograms of these partial time series are shown in Figure 3, and the performance of the different simulation models for temporal behavior of the variograms are presented in Table 5. The variogram parameters (i.e., nugget effect, range(s), and sill(s)) are listed in Table 6 based on the selected best models. The range (A1 in Table 6) varies from 10 days for Shosh Spring (S13) to 114 days for Marab Spring (S4) and from 9 days for Enakak Spring (S14) to 107 days for Sarabgarm Spring (S3), according to variograms of discharge and EC values, respectively. The range depicts a time length (i.e., discharge and/or EC) with minimum correlation. Therefore, it seems that the contribution of a flow component (e.g., conduit(s) or matrix porosity) gradually increases during the range time scale (A1) and its effect disappears in spring water.

Nugget Effect and Temporal Structures

Several of the variograms have a small nugget effect and show one or two scales of temporal structures (Fig. 4). The performance of permissible models are evaluated based on performance indexes (Table 7). The best model for simulating temporal behaviors of experimental variograms are then selected (Table 6). The nugget effect is well pronounced for Golin Spring (S2), Piran Spring (S5), and Gharabolagh (S6) Spring (Fig. 4), and this behavior may be caused by (1) measurement error and/or (2) micro-variability (e.g., variability at a scale smaller than the sampling resolution (Kitanidis, 1997)).

Two ranges with different time scales were obtained during the modeling procedure. One and two temporal structures are observed for Gilan Spring (S1), Golin Spring (S2), Berghan Spring (S9), and Tangkelagari Spring (S11), and for Atashgah Spring (S12) and Sheshpir Spring (S8), respectively (Fig. 4). The different temporal behaviors (i.e., the shape of variograms) are likely caused by different karst systems or subsystems within the aquifers.

Short range variations (A_1) occur from 17 days at Berghan Spring (S9) to 268 days at Atashgah Spring (S12) and from 28 days at Berghan Spring (S9) to 110 days at Atashgah Spring (S12) for variograms of discharge and EC, respectively. Range A_1 could be evaluated as an indicator of the length of time that spring water is dominantly supplied by a part of the karst system that contains well-developed solution conduits. The short range length of time is proposed as a measure of residence time for water stored in large solution conduits as being more important for water movement than for water storage. Long range time values (A2) are estimated to exceed 220 days (Table 6). It would seem that the entire karst system is responsive to A_2. Range A_2 may be regarded as a measure of water residence time in the fissured matrix. Temporal values of discharge and EC are uncorrelated after the A_2 time period, suggesting that the contribution of the entire karst system diminishes after this time period. The results of previous studies on residence time of some springs (Karimi, 2003) confirm our findings.

Discussion

Variations in discharge and EC in the studied springs is complex and exhibit varying temporal behaviors. The exploratory data analysis presents the information in a compact format as the first step for determining temporal structure. Plots of the obtained range (A1) according to variogram analysis versus catchment areas, percent of discharge quick flow (% Q), and ratio of maximum discharge to minimum discharge (Q_{max}/Q_{min}) are presented in Figure 5. From Figure 5, it is apparent that springs with small catchment areas have shorter ranges (i.e., residence time) than those springs with larger catchment areas. The higher values of range (A1) are observed in springs that are characterized by lower percentages of quick flow, as well as the ratio of maximum discharge to minimum discharge (Q_{max}/Q_{min}) (Fig. 5).

The discharge and EC variograms provide different ranges (i.e., residence time) for each spring. Differences between the ranges might be a result of influence of the behavior of the karst system that supplies spring water. Differences between the two ranges obtained, based on variograms of discharge and EC for the springs, vary from seven days in Piran Spring (S5) to 158 days in Atashgah Spring (S12) (Table 8). The springs could be classified into three groups based on the percent of difference between the ranges (Table 8): Group 1 springs are those springs with less than 40% differences; Group 2 are those springs with 40 to 70% differences; and Group 3 are those springs with greater than 70% differences. Group 1 includes Golin (S2), Piran (S5), Gharabolagh (S6), Sheshpir (S8) and Berghan (S9) Springs that are characterized by (1) no obvious trend in scatterplots of discharge versus EC values (Fig. 2); and (2) same temporal behavior of variograms of discharge and EC values (Figs. 3 and 4). It would seem that for these springs, spring water may be supported by a large underground karst reservoir. Heavy precipitation events could be damped by this underground reservoir. Accordingly, the temporal behavior of discharge and EC are controlled by the underground reservoir because large precipitation events do not translate into significant discharge increases and EC fluctuations (i.e., sharp increase followed by a significant decrease below the static condition). Small differences between the ranges suggest the possibility of large karst openings that supply spring water. The existence of an underground karst reservoir (or huge solutional conduits) supplying water to Sheshpir Spring (S8) was reported by Raeisi and Karami (1997).

Table 6. Parameters of selected variograms.

Spring	Model	C_0[a]	C_1[b]	A_1[c]	C_2[b]	A_2[c]
Gilan						
Discharge	Gau.	2138	14381	108		
Elec. Cond.	Gau.	172	1414	70		
Golin						
Discharge	Gau.	343	432	98		
Elec. Cond.	Gau.	103	92	126		
Sarabgarm						
Discharge	Sph.	29	75	107		
Elec. Cond.	Exp.	954	2114	69		
Marab						
Discharge	Gau.	28053	254974	46		
Elec. Cond.	Sph.	58	482	114		
Piran						
Discharge	2 Exp.	134	133	88	545	224
Elec. Cond.	Gau.	208	398	95		
Gharabolagh						
Discharge	Sph.	146	122	51		
Elec. Cond.	Gau.	383	3058	41		
Rijab						
Discharge	Gau.	50000	4458351	61		
Elec. Cond.	Gau.	0	167	95		
Sheshpeer						
Discharge	Gau.	82207	4103478	65		
Elec. Cond.	Gau.	16.5	355	47		
Berghan						
Discharge	2 Exp.	0	48541	28	303380	272
Elec. Cond.	2 Sph.	0	0.3	17	27	269
Morikosh						
Discharge	Exp.	0	15041	94		
Elec. Cond.	Gau.	1.2	14	31		
Tangkelagari						
Discharge	Gau.	56	6021	60		
Elec. Cond.	Gau.	1.5	24	33		
Atashgah						
Discharge	Sph.	2209	50000	110		
Elec. Cond.	Sph.	1.7	19.5	268		
Shosh						
Discharge	Gau.	420	874	39		
Elec. Cond.	Sph.	0	47	10		
Enakak						
Discharge	Sph.	0	47	9		
Elec. Cond.	Sph.	480	5353	19		
Ghomp						
Discharge	Gau.	484	6636	26		
Elec. Cond.	Sph.	193	750	54		

[a] Nugget effect.
[b] Sill.
[c] Range.

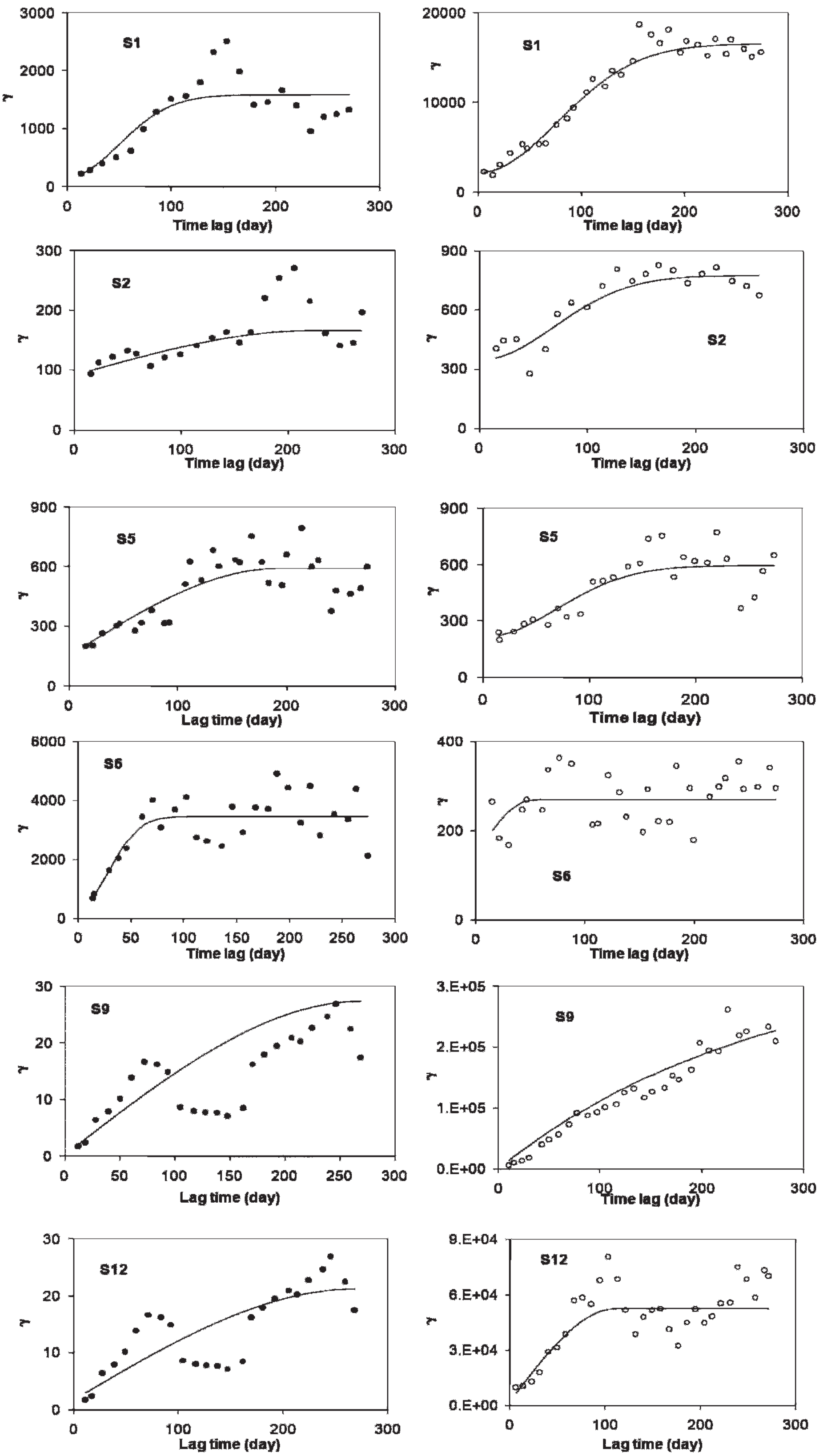

Figure 4. Non-periodic variogarams of discharge (right panel) and EC (left panel) values.

Table 7. The value of fitting criteria in modeling of non-periodic variograms (the best models are presented by bold numbers).

		Model				
Spring	Fitting Criteria	Exp.	Gau.	Sph.	Dub. Sph.	Dub. Exp.
Gilan						
Discharge	RMSE	1742	**1177**	1330	1255	1761
	AIC	575	**551**	558	558	580
	SSE	9.8	**5**	6.3	6.9	12.7
Elec. Cond.	RMSE	422	**357**	373	696	510
	AIC	324	**317**	319	349	336
	SSE	79	**35**	34	34	69.5
Golin						
Discharge	RMSE	85	**68**	71	128	94
	AIC	243	**234**	236	264	252
	SSE	9.5	**6.8**	8	20	10
Elec. Cond.	RMSE	38.7	**32.7**	37.1	34.4	39.3
	AIC	223	**216**	222	222	228
	SSE	5.1	**7.1**	4.7	6.8	13.4
Piran						
Discharge	RMSE	99.4	**90.4**	91	145	112
	AIC	330	**325**	325	353	395
	SSE	16	**13**	14	53	21
Elec. Cond.	RMSE	103	93.6	94	96	**124.3**
	AIC	386	380	381	386	**345**
	SSE	25.4	20	18.5	19.5	**7.3**
Gharabolagh						
Discharge	RMSE	59	62	**59**	96	81
	AIC	353	356	**353**	386	376
	SSE	17	18	**17**	40	30
Elec. Cond.	RMSE	684	**659**	661	1023	851
	AIC	430	**428**	428	455	445
	SSE	11	**9.2**	9.3	27	16
Berghan						
Discharge	RMSE	58,067	73,290	60,766	69,655	**24,986**
	AIC	766	780	769	781	**719**
	SSE	741	7613	856	503	**182**
Elec. Cond.	RMSE	4.9	4.8	4.5	**4.4**	6.5
	AIC	166	165	162	**164**	184
	SSE	70	75	67	**64**	125
Atashgah						
Discharge	RMSE	13,574	11,244	**11,733**	20,497	13,894
	AIC	820.2	806.6	**693.4**	732	707.9
	SSE	181.1	97.8	**116.2**	318	194
Elec. Cond.	RMSE	13.8	16.57	**4.6**	25.4	14
	AIC	275.5	286.5	**163.1**	262.9	281.8
	SSE	42.5	46.3	**70.65**	145.4	47.9

Concentrated rapid recharge through the sinkholes is the dominant recharge mechanism for Sheshpir Spring (S8).

Group 2 includes Sarabgarm (S3), Rijab (S7), Tangke-lagari (S11) and Enakak (S14) Springs that are characterized by (1) a positive correlation between discharge and EC values (Fig. 2); and (2) periodic structures in the variograms but with different temporal behaviors between discharge and EC (Fig. 3). This group may belong to

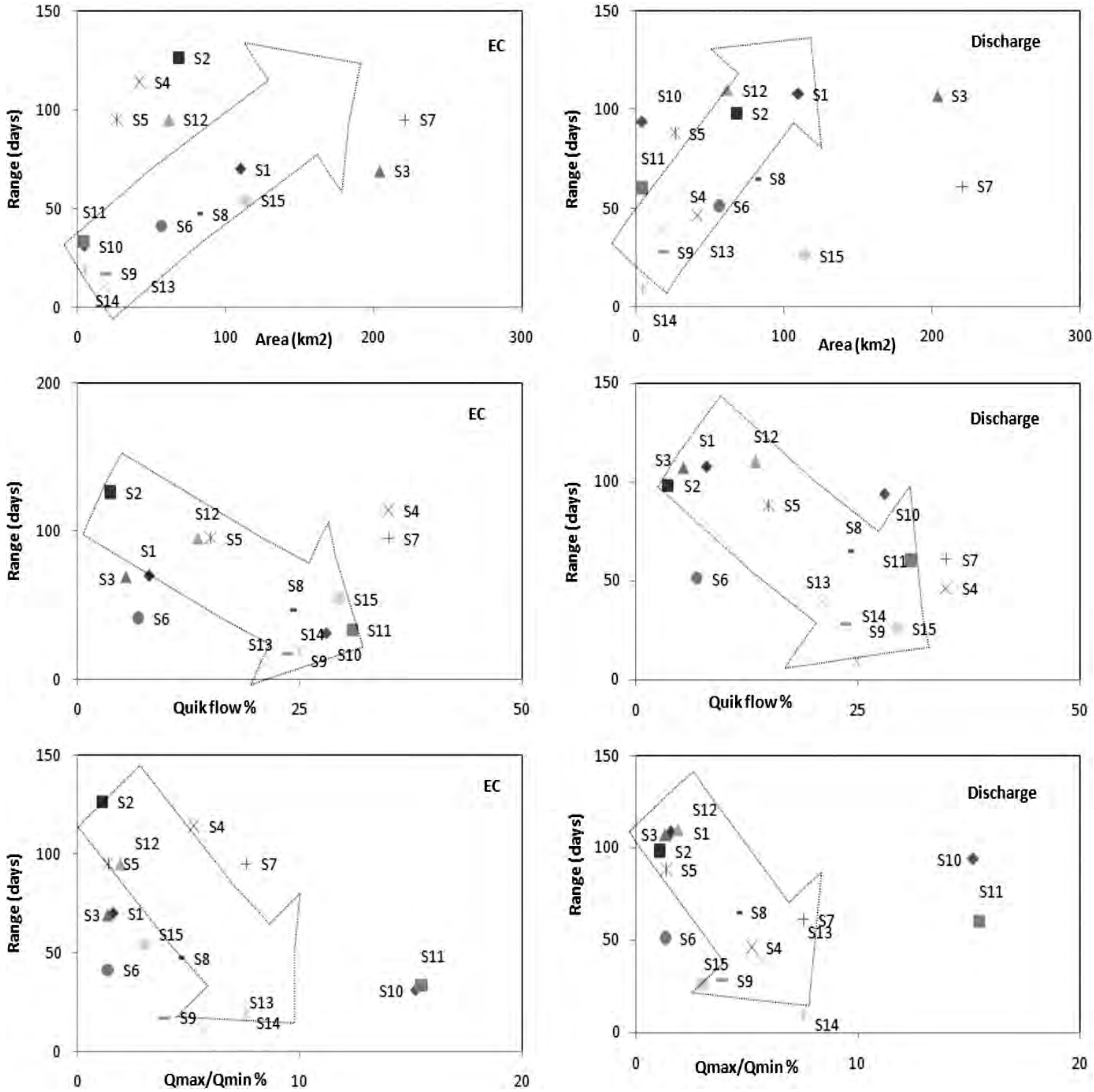

Figure 5. Relationship between residence-time range (A1) and selected characteristics of the springs.

a karst system or subsystem that is poorly developed and is dominantly displaced by a piston-flow regime. Previous findings about Enakak Spring (S14) (Keshavarz, 2003) confirm our interpretation of a karst system or subsystem dominated by a piston-flow regime for Group 2.

Group 3 includes Gilan Spring (S1), Marab Spring (S4), Morikosh Spring (S10), Atashgah Spring (S12), Shosh Spring (S13) and Ghomp Spring (S15), which are subjected to (1) a negative correlation between discharge and EC values (Fig. 2); and (2) different temporal structure in variograms of discharge and EC (Figs. 3 and 4). We believe this group is supported by a well-developed karst system or subsystem that provides higher discharge values that coincide with lower values of EC. Quick response of the karst system or susbsystem to precipitation events causes different temporal behaviors in variograms of discharge and EC.

Table 8. The ratio of difference between obtained ranges based of variograms of discharge (R_Q) and electrical conductivity (R_{EC}).

Spring	Residence Time (d) Elec. Cond. (R_{EC})	Residence Time (d) Discharge (R_Q)	$\lvert R_{EC} - R_Q \rvert$	$\left\{ \frac{\lvert R_{EC} - R_Q \rvert}{\left[(R_{EC} + R_Q)/2 \right]} \right\} \times 100$	Group
Gilan	70	108	38	43	3
Golin	126	98	28	25	1
Sarabgarm	69	107	38	43	2
Marab	114	46	68	85	3
Piran	95	88	7	8	1
Gharabolagh	41	51	10	22	1
Rijab	95	61	34	44	2
Sheshpir	47	65	18	32	1
Berghan	17	28	9	49	1
Morikosh	31	94	63	100	3
Tangkelagari	33	60	27	58	2
Atashgah	268	110	158	84	3
Shosh	10	39	29	118	3
Enakak	19	9	10	71	2
Ghomp	54	26	28	70	3

Note: *EC* is electrical conductivity.
Q is discharge.

CONCLUSIONS

The time series that describe discharge and EC variations at the springs represent aquifer behavior over the time domain. The application of variogram analysis suggests two temporal behaviors characterize the time series of discharge and EC at springs. These temporal behaviors include periodicity and nugget effect plus one or two temporal structures. For the springs studied here, the periodicity ranges from 100 to 316 days and from 82 to 300 days for variogram of discharge and EC, respectively. The temporal structure in one cyclical period is explored by application of variogram on partial data in a cycle.

Some of the variograms are modeled by double exponential or spherical models which introduce two temporal ranges (i.e., A1 and A2). The short range (A1) can be considered as an indication of water residence time in well-developed karst conduits, while the entire karst system is responsive to the long range (A2). The springs are classified into three groups according to differences between ranges obtained by variograms of discharge and EC that belong to the development of karst in each system. The results obtained in this study confirm previous findings of the study area and provide valuable new findings regarding the temporal structure of the aquifers and additional insights into the karst systems. This research also illustrates how variogram analysis can improve our understanding of karst systems by using time series of physico-chemical parameters. The authors propose the application of variogram analysis on time series of physico-chemical parameters as a part of karst spring studies.

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BOOK REVIEW

Enhanced Characterization and Representation of Flow through Karst Aquifers

S. L. Painter, A. Sun, and R. Green, 2007. AwwaRF Report 91139, Denver, Colorado, IWA Publishing (International Water Association), 104 p. ISBN 978-184339979-7, softbound, $240.

Karst aquifers are generally considered difficult or sometimes even impossible to characterize or to model mathematically. This is mainly because of their often extreme hydraulic heterogeneity and problems in detecting and characterizing karst conduits. Still, it is important to deal with karst groundwater quantitatively, since it constitutes important water resources and because the aquifer responses are highly variable in space and time. Aquifer-management issues are of prime importance. The very low storage and high hydraulic conductivity of most carbonate aquifers imply that groundwater storage volume might not last throughout lengthy droughts. Appropriate mathematical tools are required to manage groundwater resources properly by controlled pumping, conjunctive-use schemes, artificial recharge, etc. Also, runoff from extreme recharge events can be rapidly transmitted through karst systems and may contribute to flooding. Therefore, I welcome contributions to the methodology of modeling karst groundwater flow and transport. Most publications dealing with mathematical modeling of karst aquifers to date are based on working codes that are not widely available and are difficult to handle. Developing a code similar to the USGS MODFLOW would be a great step toward quantitative analysis of karst groundwater systems. Although not explicitly stated as a primary objective of this book, this was apparently the authors' intention. The book is written in a report style and includes project planning issues rather than just results. The AwwaRF Report Series has recently been retitled the Water Resources Foundation Report Series.

The Introduction (Chapter 1) describes the motivation, objectives, and structure of projects. While it covers the development of tools for flow simulation, the authors also develop a program package for the modeling of complex transport problems.

Chapter 2 is a short review that illustrates the importance of following a variety of pathways in the modeling of karst aquifers. The short description of karst classification includes descriptions of different types of karst aquifers and types of cave origin. The latter is not especially relevant in this context, since many caves (e.g., those developed under vadose conditions) are not necessarily compatible with the modeling. Few important karst aquifers are dominated solely by conduit flow. The authors confirm this in the section "Groundwater Flow Regimes." The relevance of their Figure 2.1 is not well substantiated: it shows a correlation between permeability and the age of the rocks. It demonstrates the large variability of hydraulic conductivity, but provides little information about how to make generalizations from the data. Better examples, or a schematic, would have better demonstrated the dual-flow response of springs and the coupling between matrix storage and conduit flow.

In Chapter 3, the main approaches for simulating flow in conduits are briefly summarized. The authors describe three different modeling approaches: the "smeared conduit" (an equivalent porous medium, single-continuum approach), the "embedded channel" (a hybrid, discrete pipe–continuum approach), and the "dual conductivity" (a dual-continuum approach). This classification follows that presented by earlier authors. Since no new concept is introduced, I suggest that the existing terminology could have been used to avoid confusion. The problem of employing appropriate flow laws, laminar or turbulent, is addressed, as well as the issue of representing flow in unconfined aquifers. Technical aspects such as the drying and rewetting of cells in MODFLOW and the necessity of specifying the vertical position of conduits for unconfined conditions are also addressed. In a numerical experiment, the effect of the flow law employed on the simulated head is demonstrated.

Chapters 4 and 5 demonstrate with numerical experiments the capabilities and limitations of the two modeling approaches "smeared conduit" and "Dual-Conductivity Model" (DCM). At the time of publication in 2007, development of the "embedded conduit" model was not yet complete. The performance of an equivalent porous medium (single continuum, MODFLOW) model is compared with a prototype discrete model, generated with FEFLOW. The first experiment investigates the response in a single conduit following a recharge event; the second features a two-branch conduit, and the third a dendritic conduit pattern. The authors state that the models perform reasonably well except when the total conduit volume becomes large. These kinds of numerical experiments had been conducted already in the 1990s by U. Mohrlok (e.g., Mohrlok and Liedl, 1996; Mohrlok et al., 1997), who compared a prototype complex aquifer response with a dual-continuum approach. The problem with single-continuum models is that they generally honor flow and the water balance but fail to simulate observed transport velocities. This limitation should be clearly stated. The geometry adopted for the modeling experiments should be closer to that actually observed for karst catchments (e.g., conduits generally drain to surface-water bodies). There is uncertainty about the type of boundary condition at the spring. The authors presumably assume fixed-head conditions.

The experiments with the newly developed double-continuum DCM package consist of simple aquifer-drainage experiments for validation, an imposed recharge pulse on the earlier dendritic conduit network, and an individual conduit in an unconfined karst aquifer. Model results are compared with the FEFLOW prototype and a single-continuum approach. DCM matches well the prototype pulses. To simulate natural aquifers more closely, experiment three should be designed so that the conduits can drain directly to the constant-head boundary.

Finally, in Chapter 6, a model of the Barton Springs catchment of the Edwards Aquifer is built with the double-continuum approach, and a 1-year period was simulated with DCM code. Stability and performance characteristics were tested.

Since this report was published in early 2007, and considerable progress may have been achieved since then, my comments may be outdated. Nevertheless I would like to make a few final conclusions. The authors demonstrate that dual-continuum modeling is feasible for simulating karst aquifers. The data requirement is moderate, and the models mimic the typically observed spring responses. It is also important to provide the community with a tool that is readily available to and usable by consultants. In getting this message across, the authors have contributed a great deal. Regarding the report itself, its preparation could have been more professionally oriented, especially given the high price. It could use a proper review of available karst models and their relative benefits and limitations, including the work of European scientists active in this area since the 1970s; a discussion, with illustrations, of the special characteristics of karst systems and why they require specific modeling approaches; a stronger emphasis on guidelines; and integration of the figures with the text. The title is somewhat misleading, since characterization issues are not addressed.

In any case, this work has potential and I look forward to a new edition of the book, which can be expected to include the recent achievements in karst-model development of the author group, as well as other advances, such as the newly developed hybrid model of the USGS (W. B. Shoemaker, CFP-Model).

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