Appalachian KARST

Proceedings of the Appalachian Karst Symposium
Radford, Virginia March 23–26, 1991
Ernst H. Kastning and Karen M. Kastning, Editors

Sponsored by the
National Speleological Society
In Celebration of
Its 50th Anniversary
Hosted by the
Department of Geology
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Cover photo: Looking out of the entrance, Sinks of Gandy, West Virginia

Cover: The photograph on the cover of this volume is a view looking out from the upstream entrance of the Sinks of Gandy, a 1.5-mile-long cave in southeastern Randolph County, West Virginia. Gandy Creek sinks at this point and resurges at the northern end of the cave. As this picture shows, the realm of karst encompasses both the surface and the subsurface. These zones are linked by water flow, either that now moving through solutional cavities and conduits or that which excavated these openings in the geologic past.

The Sinks of Gandy was an important cave in the early history of the National Speleological Society. Members of the District of Colubia Speleological Society made frequent trips to the cave in the 1930's and Jack Prebble, an early NSS member from Ohio, popularized the cave by founding the Ibinthruthesinks Club in 1937 (*see* membership card, p. 34 of the January 1991 *NSS News*). The caver in this view is Kass Kastning, one of the authors in this volume. *Photograph by Ernst and Karen Kastning.*

Preface

This volume, *Appalachian Karst,* is the official Proceedings of the Appalachian Karst Symposium, held at Radford University, Radford, Virginia on 23-26 March 1991. This was the first professional meeting to address the karst region extending along the Appalachian Mountain Range from Alabama to Maine and including the Valley and Ridge, Appalachian Plateau, New England Upland and Coastal Plain Provinces. The scope of the conference included geologic and geographic studies on all components of karst, from surficial landforms and processes to those in the subsurface. Furthermore, the papers presented herein reflect a balance between basic research of karst processes and application of karst science toward understanding and solving environmental problems of karst terranes.

Thirty-three papers on Appalachian karst were presented at the meeting by fifty-one authors and co-authors. Thirty-one of these papers are published here in their entirety. Manuscripts were not received by press time for the remaining two, however, the abstracts for these are included in this volume. Additionally, the morning session of the first day of papers included several presentations that were either not relevant to the Appalachian region or were proposals of various kinds. These are not included in the Proceedings.

Attendees at the Appalachian Karst Symposium, Radford University, 23-26 March 1991. *Photograph by Ira D. Sasowsky.*

The impetus for the Appalachian Karst Symposium was three-fold. First, the conference was held in commemoration of the 50th Anniversary of the founding of the National Speleological Society (NSS) in 1941. The NSS Board of Governors sanctioned the Symposium as an official event in the 1991 year-long celebration of the Golden Anniversary of the Society.

Second, it was appropriate to devote a meeting specifically to the Appalachians, as the beginnings of the Society were firmly rooted in this region. The District of Columbia Speleological Society (DCSS) and the Spelunkers Club of New England (SCNE) were both independently organized in the 1930s. Members of the DCSS actively explored and studied caves in the mountains and valleys of Virginia and West Virginia. Members of the SCNE went "spelunking" in the caves of New England and New York. The two groups merged by mutual agreement in 1941 by chartering the National Speleological Society in January 1941. Additional information on the formative years of the NSS can be found in commemorative articles in the December 1990 and January 1991 issues of the *NSS News*. A complete history of the National Speleological Society, *Caving in America: The Story of the National Speleological Society, 1941-1991,* has recently been published (June 1991).

The third motivation for the meeting was to convene the Friends of Karst (FOK). The FOK is an informal organization of individuals interested in research in physical speleology. Most "members" are active within the Geology and Geography Section of the NSS. The FOK meets irregularly, but never more than a few years apart. Many members of the FOK living in the eastern half of the United States attended the Symposium and presented papers.

As editors, we express our deep appreciation to all of the contributors to the meeting. It is interesting to note that every paper presented was volunteered and the meeting organizers did not have to solicit a single contribution! We also thank Drs. Chester F. "Skip" Watts and Stephen W. Lenhart of the Department of Geology at Radford University and David Hubbard of the Virginia Division of Mineral Resources for driving vans on the opening-day fieldtrip. Michael and Lori Kilgore contributed considerable time and energy at the registration desk and in making refreshments available during the meeting. The Administration at Radford University fully supported the conference by providing facilities for meeting. The Department of Geology at Radford University (Dr. Robert C. Whisonant, Chairman) hosted the conference. Mr. Jerrold L. Perry of Radford University Food Services made arrangements for the banquet.

We thank Dr. William B. White for his Banquet presentation, *The Appalachian Karst: Historic and Futuristic Impressions*. The NSS Board of Governors approved Society funding for publication of this volume. David McClurg, NSS Special Publications Committee Chairman, served as editorial consultant and as liaison with the printer. We are *especially* grateful for his patience as we struggled with our editorial tasks.

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APPALACHIAN KARST

Proceedings of the Appalachian Karst Symposium

RADFORD, VIRGINIA, 23-26 MARCH 1991

Table of Contents

Karst Erosion Surfaces in the Appalachian Highlands

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ABSTRACT

Among the various "peneplain" surfaces identified in the Appalachian Highlands by the early geomorphologists, the Harrisburg Surface is one of the most important. The surface varies in elevation from 200 meters in the Cumberland Valley at Harrisburg, Pennsylvania to 760 meters in Virginia and West Virginia. The best preserved remnants of the Harrisburg Surface are doline karst plains. Internal drainage through the dolines prevents dissection by surface streams and maintains an easily recognizable surface. Those portions of the Harrisburg Surface preserved as doline karst are continuously lowered by dissolution of the limestone and its removal by internal runoff. The rate of lowering based on present day denudation rates is 20 to 40 meters/megayear with 30 m/Ma selected as a representative value. Various representative segments of the Harrisburg Surface have been reconstructed backward in time by "replacing" the missing limestone. Such reconstructions are often blocked by stratigraphic or structural barriers thus giving a limiting age for the Harrisburg Surface. In general, the Harrisburg Surface appears to be younger than the classical estimates with a late-Tertiary (perhaps Pliocene) age rather than the Early Eocene or late Cretaceous ages proposed in some recent work.

Introduction

Research on karst is turning from descriptive studies of individual karst areas to either small scale analysis of karst processes or to large scale interpretations of entire drainage basins or physiographic provinces. Among the latter investigations the question of age and time scale must arise. That is, we would like to know the ages of present day features in the karst landscape and we would like to know the length of time required for their development. In a very rough way, time scales that must be dealt with by investigators of contemporary karst landscapes (as distinguished from paleokarst) can be divided into three groups.

(i) Karst phenomena that span a few hundred thousand years. These include development of most active karst drainage systems, deposition of speleothems, and transport of clastic sediments. There is a substantial choice of dating methods including U/Th, ESR and thermoluminescence, amino acid racemization, and for the youngest events, radiocarbon (*see* White, 1988, Chapter 10 for review).

(ii) Karst phenomena that span a few million years. This includes development of large cave systems and development of most karst landscapes. The only absolute dating tool is the pattern of magnetic normals and reversals recorded in cave clastic sediments (Schmidt, 1982). Most dating depends on geomorphic context.

(iii) Karst phenomena that relate to the oldest surviving features of the landscape. The time scale could be tens of millions of years and might span the entire Cenozoic Era. Mostly, these features are the doline karsts or sinkhole plains which have descended from equivalent features whose ultimate origins lie far back in Tertiary or even late Cretaceous time.

The present paper is concerned with karst features on the intermediate time scales of a few million years and also on what may be the oldest karst features preserved. These are the doline plains in the Appalachian Highlands. The paper is concerned with their relation to the Harrisburg erosion surface and the time scale for their evolution.

The Harrisburg Surface

Introduction of the Harrisburg Surface

The peneplain concept has its origins in the Appalachian Highlands where it was introduced by William Morris Davis (1889) in his classic *Rivers and Valleys of*

Pennsylvania. The idea that the landscape is eroded down to a flat plain which is then elevated and dissected again has taken some abuse over the years although recently it may be said to have made a modest comeback (*see e.g.* Morisawa, 1989). Of the multitudinous erosion surfaces introduced in the hayday of the peneplain, two have the greatest acceptance and to some extent have survived the turmoil. One is the Schooley Peneplain, which is represented by accordant mountain summits and ridge lines in the Appalachians. The Schooley surface is so highly dissected and fragmented that little can be said about it. The second is the Harrisburg Surface, represented throughout the Appalachians by valley uplands, which is the most pervasive of the erosion surfaces. Much of the Harrisburg Surface is karst. Descriptions of the Harrisburg Surface in the traditional style are given by Fenneman (1938) and by Thornbury (1965).

Tracing the Harrisburg Karst Surface Through the Appalachians

The tracing of the Harrisburg Surface given here is to some extent schematic and depends on the correlations of others. It is quite clear, however, that the valley upland surface that can be traced along the major rivers in the Appalachians rises toward the headwaters and, whatever its regional extent, it is not a horizontal surface.

In the "type locality", the broad Cumberland Valley near Harrisburg, Pennsylvania, the rolling upland surface is at an elevation of about 520 feet (160 m). The Cumberland Valley rises somewhat to the south reaching 700 feet (210 m) north of Hagerstown, Maryland and then falls again to elevations near 500 feet along the Potomac River. West of Harrisburg, there is a pronounced valley upland forming accordant hill tops above the incised valley of the Juniata River. The upland surface reaches an elevation of 1000 feet (305 m) in Huntingdon County and 1200 feet (365 m) at the basin divide in the broad Nittany Valley near State College, Pennsylvania. The upland surface is highly dissected where underlain by shales and well-preserved where underlain by limestone.

In the Great Valley Sub-Province south of the Potomac, the upland surface can be traced along the Shenandoah River as a valley upland that reaches 2000 feet (600 m) at the drainage divide. Cave development in the Shenandoah Valley can be correlated at least roughly with the upland surface (White and White, 1974). The Great Valley Sub-Province is less clearly defined along the transverse valleys of the James River and the New River but an upland surface can be traced at about 2000 feet (600 m). Remnants of valley uplands can be followed into the headwaters of the various tributaries. Burnsville Cove, in the upper reaches of the James River drainage, has a remnant of a karst surface at the upstream divide at an elevation of 2500 feet (760 m). The extensive karst in the Greenbrier River drainage, part of the New River Basin in West Virginia, has been claimed to represent the Harrisburg Surface (Price and Reger, 1929, plate 16B).

In southwestern Virginia, eastern Tennessee, and northern Alabama in the drainage of the Tennessee River are also upland surfaces. These are not considered in the present paper. However, the Highland Rim surface of central Tennessee, below the escarpment of the western margin of the Cumberland Mountains, has been correlated with the Harrisburg Surface by Fenneman (1938).

Recent Estimates of the Age of the Harrisburg Surface

Unlike rocks, which in the first place were formed in a specific depositional event and in the second place carry signatures of their age in the form of fossils, radioactive minerals, and other indicators, "surfaces" or "topography" (being in a certain sense the absence of rocks), are difficult to assign an age. In the first place, topography is continuously evolving; it is not formed in a single discreet event. In the second place, age indicators that exist are fragmentary and often inconclusive.

Some indication of the range of opinion is given by the following quotes, taken from Sevon (1985):

> "Little of the earth's topography is older than the Tertiary and most of it is no older than the Pleistocene."
> Thornbury, 1969

> "....one can conclude that the great bulk of erosion of the Appalachians took place in the Cretaceous or earlier, and that the present mountains have had a relief not much less than the present since well back in the Tertiary, perhaps since the Eocene."
> Rogers, 1967

The Harrisburg Surface itself has the best record of the many Appalachian erosion surfaces in part because of saprolites, mineral deposits, and various residual sediments. Sevon's (1985) thoughtful analysis of the Harrisburg Surface in Pennsylvania concludes that the surface is very old. Sevon (p. 22) says:

> "The Late Cretaceous, Paleocene, and Early Eocene subjected Pennsylvania to an extended period of humid subtropical climate during which chemical weathering was at an extreme while physical appearance of the landscape changed slowly. The very long period of extreme climatic conditions allowed the landscape to achieve a state of adjustment (dynamic equilibrium?) which resulted in a climax landscape -- the Harrisburg erosion surface."

> "The timing of events related to the Harrisburg Surface is not by any means exact. The development of the surface occurred over a period of time which may have been as long as 45 m.y."

Karst Surfaces

Preservation of Karst Surfaces: Doline Karst and River Terraces

Many of the best preserved remnants of the Harrisburg Surface in the Appalachians are on carbonate rock. Because of internal drainage through sinkholes, carbonates remain as undissected plains with an aspect of an "erosion surface", at least to casual inspection. However, these "surfaces" are being continually lowered by dissolution of carbonate rock at the soil/bedrock contact and transport of dissolved material into an underlying karstic drainage system.

Figure 1 illustrates the essentials of the problem. In regions of non-carbonate rock, rapidly down-cutting valleys take on the characteristic V-shape. During extended periods of stable base level, downcutting slows down and the valleys widen, forming a wide flood plain. Renewed downcutting results in a rejuvenated V-shaped valley cut into the old flood plain, remnants of which remain behind as a terrace or valley upland. The usual assumption, not always stated specifically, is that the rate of lowering of the valley upland is much less than the rate of downcutting of the rejuvenated valley.

In contrast, the karst plain need not change its form. To be sure, sinkholes become filled, new sinkholes form, and there is a continual evolution of the detail of the landscape. Viewed in the large, however, the visual appearance of the landscape remains much the same. On the average, the sinkhole plain is simply lowered at a rate determined by the rate of dissolution of the underlying carbonate rock. Unlike the valley upland on clastic rock, it should not be assumed that the sinkhole plain marks a fixed elevation.

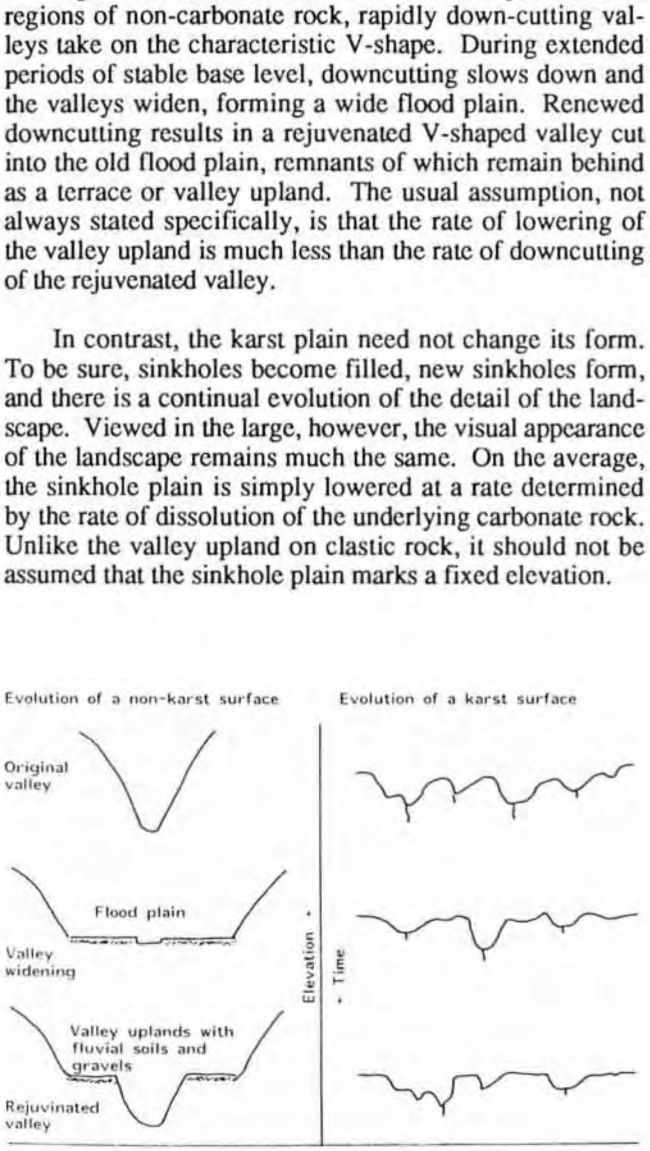

Figure 1: Sketch showing the contrast in landscape morphology between an eroding valley and an internally drained karst surface.

Denudation Rates

In an earlier investigation (White, 1984), the rate of lowering of a karst surface was calculated from first principles using a model which assumed that infiltrating groundwater came into equilibrium with calcite at the soil-bedrock contact. The equation is

$$D_n = \frac{100}{\rho\sqrt[3]{4}} \left(\frac{K_c K_1 K_{CO_2}}{K_2}\right)^{1/3} P_{CO_2}^{\;1/3} \quad (P - E)$$

The denudation rate, Dn, is given in this equation in units of mm/ka. This is numerically equal to the units of µm/a preferred by those who make direct microerosion measurements and m/Ma often used for denudation rates calculated from clastic load in non-karstic river basins. ρ is bedrock density in g/cm^3, the various K's are equilibrium constants for the usual carbonate reactions (*see* White, 1988 and many other sources), P_{CO_2} is CO_2 partial pressure in units of atmospheres, and (P-E) is precipitation minus evapotranspiration (or runoff in karstic basins with no surface streams) in units of mm/a.

The equilibrium denudation equation is plotted in Figure 2 in comparison with the regression line for denudation rates in temperate climate karsts compiled from world sources by Smith and Atkinson (1976). The two lines have similar slopes but the regression line for the world data does not pass through the origin. There is an offset of

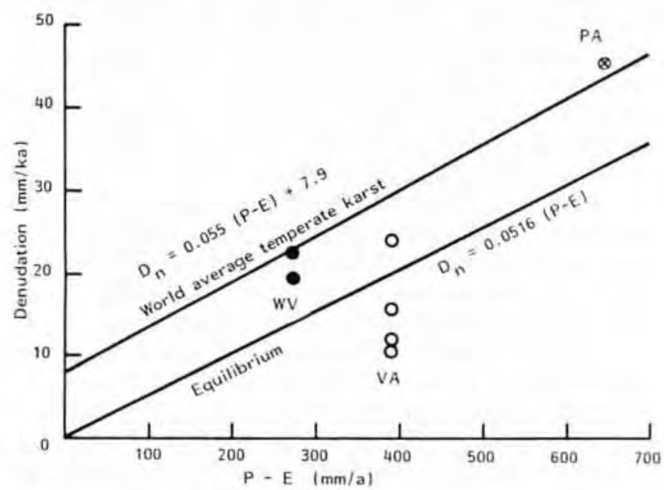

Figure 2: Denudation rate as a function of precipitation minus evapotranspiration. The "world average temperate karst" is the regression line of Smith and Atkinson (1976) for many experimental measurements. The "equilibrium" line is a first principles calculation based on the model of White (1984). Points show experimental measurements from the Appalachians: WV (Ogden, 1982); VA (calculated from Harmon and Hess, 1982 and Davis and Hess, 1982); PA is calculated from Thompson Spring in central Pennsylvania.

8 mm/ka at zero runoff. This may be merely an artifact of the scatter in the measured denudation rates.

In principle, the denudation rate in an actual karst drainage basin can be determined if one knows the area of the basin, the basin runoff, and the total dissolved carbonate load discharged by the springs which drain the basin. In practice, data are usually sparse and there are some problems with relating dissolved load to long term average lowering of the land surface. The dissolved carbonate load derives in part from dissolution at the land surface which penultimately results in a lowering of that surface, but also includes contributions that are carried into the basin by sinking streams as well as contributions from the dissolution of bedrock at depth in the subsurface. Bedrock removal in the subsurface includes dissolution along fractures and bedding planes and enlargement of conduit drainage systems. Furthermore, most limestone spring waters are known to be undersaturated with respect to calcite because of the relatively sluggish kinetics of carbonate dissolution in open conduit systems. Thus, the measured denudation rate should be less than the equilibrium denudation rate. However, the regression line based on observed denudation rates for world-wide data in Figure 2 is actually higher than the equilibrium line.

Few denudation measurements have been made in Appalachian watersheds. Data from Monroe County, West Virginia and Burnsville Cove, Virginia are plotted on Figure 2. The West Virginia data were obtained from time series data on three conduit flow springs receiving most of their catchment from carbonate rocks with some contribution from surface streams (Ogden, 1982). The Burnsville Cove denudation rates were obtained by combining chemical analysis data for four springs (Harmon and Hess, 1982) with some low flow discharge data (Davis and Hess, 1982). Because of the use of low flow runoff, these values are probably less than mean annual denudation rates.

Data are rarely available to determine the flux in response to different runoff conditions. However, by a stroke of good fortune, the small karstic basins of Thompson Spring, a diffuse-flow system, and Rock Spring, an open-conduit-drainage system, both in Centre County, central Pennsylvania, were instrumented for discharge and regularly sampled for chemical analysis during an eight-month period that included the Hurricane Agnes storm that swept across Pennsylvania in June of 1972. From these data (Jacobson, 1973; Jacobson and Langmuir, 1974) it is possible to calculate the flux of dissolved carbonate rock being removed from these two small basins (White, 1990). What was immediately obvious was that although the concentration of dissolved carbonate is diluted during storms, the decrease in concentration is more than compensated by increased discharge. The net result is a large increase in the rate of denudation during high-runoff events. Because of smaller fluctuations in discharge during storm events, diffuse-flow springs give a more reliable measure of karst denudation, at least if one has only short periods of record.

The rate of denudation can be calculated by integrating the flux of dissolved rock over the time period of observation. Because the Thompson Spring data are discrete observations taken at a few-day intervals, the denudation rate is expressed as a summation

$$F_i = Q_i \, Hd_i/\rho A$$

$$D_n = \frac{31.55}{t_{rec}} \sum_{i=1}^{n-1} F_i \, (t_{i+1} - t_i)$$

F_i is the instantaneous flux of carbonate rock lost from the basin in units of $cm^3 sec^{-1} km^{-2}$; Dn is the denudation rate in mm/ka. The time unit here is in days. Q_i is discharge in m^3/sec, Hd_i is hardness as $CaCO_3$ in g/m^3, ρ is rock density in g/cm^3, and A is basin area in km^2. Thompson Spring receives its entire recharge on a limestone and dolostone valley upland which is the local expression of the Harrisburg Surface. The catchment area is 17.6 km^2 (estimated upward from the 11.1 km^2 given by Jacobson) giving a denudation rate of 45 mm/ka. Average precipitation in the Centre County area is 980 mm according to the University Weather Station. Depending on estimates of evapotranspiration losses which vary strongly with season, plant cover, and topographic location, runoff is about 650 mm, allowing the Thompson Spring data to be plotted on Figure 2.

Based on Figure 2, a denudation rate of 30 mm/ka has been selected for the Appalachians. The uncertainties are large and a finer tuning does not seem to be warranted. The number is in the same range as denudation rates based on other basin measurements and calculations summarized by Sevon (1985) and with the value of 27 mm/ka determined by Sevon (1989) for the entire Juniata River Basin.

Age Dating of the Karst Surfaces

One approach to the dating of erosion surfaces is to examine residual soils. If these can be demonstrated to be residual, let-down soils, rather than alluvium or colluvium, then a comparison of soil thickness with insoluble residue in the underlying bedrock gives a minimum rock mass that must have been removed to produce the soil.

An example from the Harrisburg Surface in Pennsylvania was described by Parizek and White (1985). The Harrisburg Surface cuts across the broad, multiply-folded arch of the Nittany Anticlinorium and exposes the Cambrian Gatesburg Dolomite at its center. This is an interfluve area with no evidence for transported soils. There occurs on the Gatesburg a residual soil with a thickness of 30 to 100 meters for which a representative thickness was estimated to be 50 m (165 feet). To get an idea of the amount of carbonate rock removed to produce the residual soil, we take the measured average insoluble residue of 7.27 percent, a bulk density of the soil of 1.76 gcm^{-3}, and

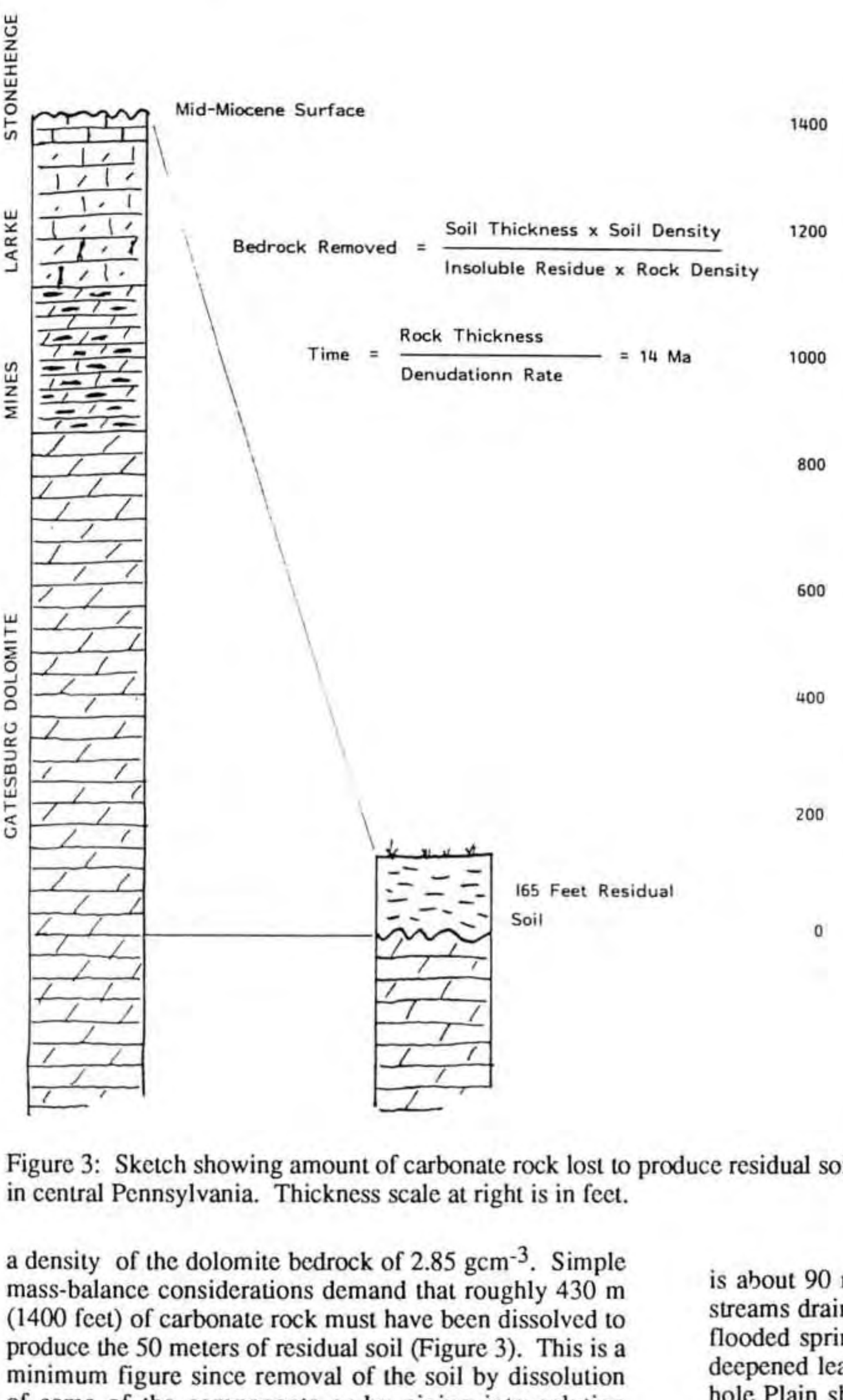

Figure 3: Sketch showing amount of carbonate rock lost to produce residual soil in central Pennsylvania. Thickness scale at right is in feet.

a density of the dolomite bedrock of 2.85 gcm^{-3}. Simple mass-balance considerations demand that roughly 430 m (1400 feet) of carbonate rock must have been dissolved to produce the 50 meters of residual soil (Figure 3). This is a minimum figure since removal of the soil by dissolution of some of the components or by piping into solution cavities in the bedrock would demand an even larger bedrock thickness.

Using 30 m/Ma as the denudation rate gives 14 million years as the time required for the bedrock column to be dissolved. Some predecessor of the present surface with enough stability to preserve the residual soil must have existed since mid-Miocene time. This is consistent with some of the younger estimates for the age of the Harrisburg Surface, but it immediately poses a different problem.

The present day elevations of the mountain ridges that border the Nittany Valley are 2000 to 2500 feet (610 to 760 meters) with a relief of 820 to 1310 feet (250 to 400 meters). One of the assumptions in the interpretation of the Appalachian landscapes is that the resistant quartzites that support the mountain ridges have lowered relatively little during the Cenozoic. Adding 430 meters to the valley floor moves it above the ridge tops, which seems unlikely. There is carbonate rock available as shown on Figure 3, but because of the curvature of the anticlinal arch, the mid-Miocene ridges must have stood farther out into the valley and they must also have eroded substantially. This begins to look like the dynamic-equilibrium model proposed by Hack (1960).

One additional result can be deduced from this number-juggling. The present day Harrisburg Surface in central Pennsylvania is dissected by present-day streams producing a local relief on the surface, particularly evident on shale-floored valleys, of about 100 meters. Taken as a proportion of the 14 Ma that the valley uplands have been lowering, about 3 Ma has been required for the recent dissections. That also gives a late Pliocene or Pleistocene age for the oldest cave fragments that have formed in the remnants of the upland surface in response to the downcutting.

The other piece of ground truth comes from the Mammoth Cave area. The Mammoth Cave System occurs beneath the sandstone caprock of the Mammoth Cave Plateau. The present-day relief between the Sinkhole Plain to the south and east and the sandstone contact is about 90 m. Recharge into the sinkholes and sinking streams drains beneath the Plateau to some slightly backflooded springs on the Green River. As the cave system deepened leaving abandoned high-level trunks, the Sinkhole Plain should have been lowering to keep pace with the active drains at the base of the cave system.

There have been three different attempts to estimate the age of Mammoth Cave from geomorphic evidence (Deike, 1967; Miotke, 1975; Palmer, 1989) all of which more or less agree that the highest abandoned cave levels date from late Pliocene or early Pleistocene. The second line of approach is Schmidt's (1982) measurements of magnetic reversals. Down in the floodwater zone of the cave, the magnetic mineral grains in the sediments are

pointing to the present-day north pole. Near the middle of the system, around the elevation of Cleaveland Avenue, some of the sediments are reversed, then they become normal again, then reversed again. Comparing these results with the known paleomagnetic time scale also gives an age of about two to three million years.

If we consider the relief of a passage like Collins Avenue, which is right up against the sandstone 90 meters above present-day base level, and ask when it was an active drain carrying water to the paleo-Green River, it was presumably getting its water from the paleo sinkhole-plain when it was 90 meters above the present-day sinkhole plain. Dividing 90 meters by the denudation rate of 30 m/Ma again gives an age of 3 million years. The Mammoth Cave area checks out. Three independent assessments: a geochemical argument, a paleomagnetic argument, and a geomorphic drainage-basin development argument all come out with the same answer to plus or minus something in the first decimal place. This is also the best evidence that scaling karst surfaces backward in time with the denudation rate as a scale factor is not a completely erroneous exercise.

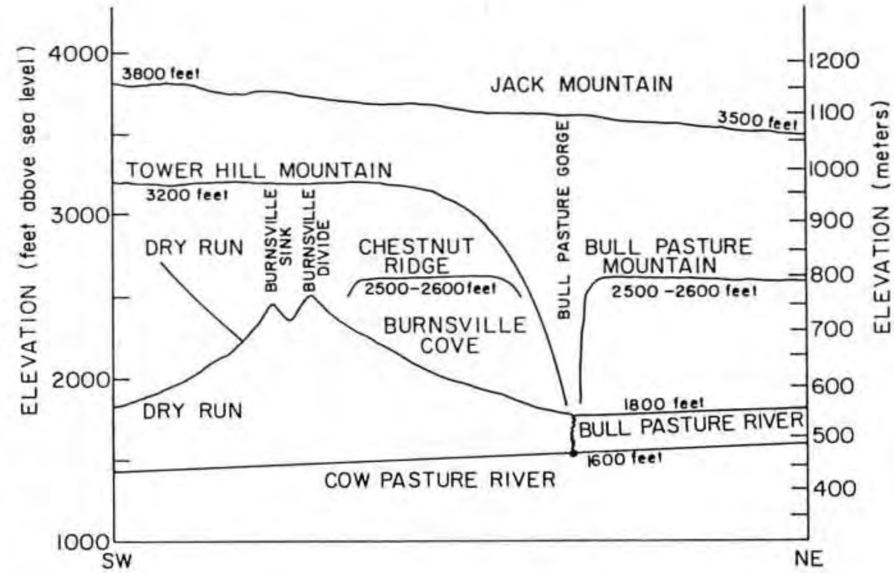

Figure 4: The Harrisburg Surface in Burnsville Cove, Bath County, Virginia.

Surface Reconstructions

The Valley and Ridge of Virginia: Burnsville Cove

On the drainage divide at the head of Burnsville Cove is a little remnant of the upland surface which is characterized by large closed depressions (White and Hess, 1982). There are fairly steep valleys falling away on both sides. There is something rather curious (Figure 4). The Burnsville karst is at an elevation of 2500 feet (760 m), the same as the top of Bullpasture Mountain and Chestnut Ridge which are remnants of the erosion surface capped with clastics. Although the Harrisburg here is a ridge-top surface, it is much lower than Tower Hill Mountain or Jack Mountain which would represent the Schooley surface. Present-day base level is down at the Bullpasture River.

Now what happens if this surface is pushed back in time? There is a karst erosion surface which is equal to sandstone-capped mountain ridges and things are staying in place. It really does look like an erosion surface because it cuts across bedrock of different lithologies, but some mechanism must be found for keeping the carbonates and the clastics in accordance.

The Greenbrier Karst of West Virginia

In the Greenbrier River drainage of West Virginia there's another remnant, called the Little Levels, around the town of Hillsboro in Pocahontas County (Figure 5). The Little Levels are at an elevation of 2300 to 2500 feet (700 to 760 m), about the same as in Burnsville Cove. The upland karst surface is a doline plain with large closed depressions. The present-day surface drainage, the Greenbrier River, is cut down to about an elevation of 2100 feet (640 m). Another fragment of this representation of the Harrisburg Surface occurs in the Swago Creek Basin to the north and a larger fragment appears as the Big Levels, also called the Great Savanna, a major karst surface in Greenbrier County to the south (Jones, 1973) (Figure 6).

The southwestern boundary of the Little Levels is the sandstone-capped Droop Mountain, which separates it from the limestone-floored Hills and Bruffey Creek valleys to the northwest. Much of the Little Levels is backed up against the Yew Mountains where the limestone dips away beneath the Allegheny Plateau. The top of Droop Mountain is 3000 feet (915 m), or a little higher, but the top of the limestone is lower on its flanks. If the karst surface is reconstructed by replacing the lost limestone, it requires only a few million years for the western edge to butt up against the Droop sandstone and run out of limestone. Meanwhile the eastern edge of the karst surface slides up the dip slope of the western flank of the Browns Mountain anticline. However, if the Greenbrier River has not shifted its course, the eastward migrating surface will be cut off by the river as it is today.

It is difficult to see how the karst erosion surface on the Greenbrier Limestone can be much older than 2 to 5

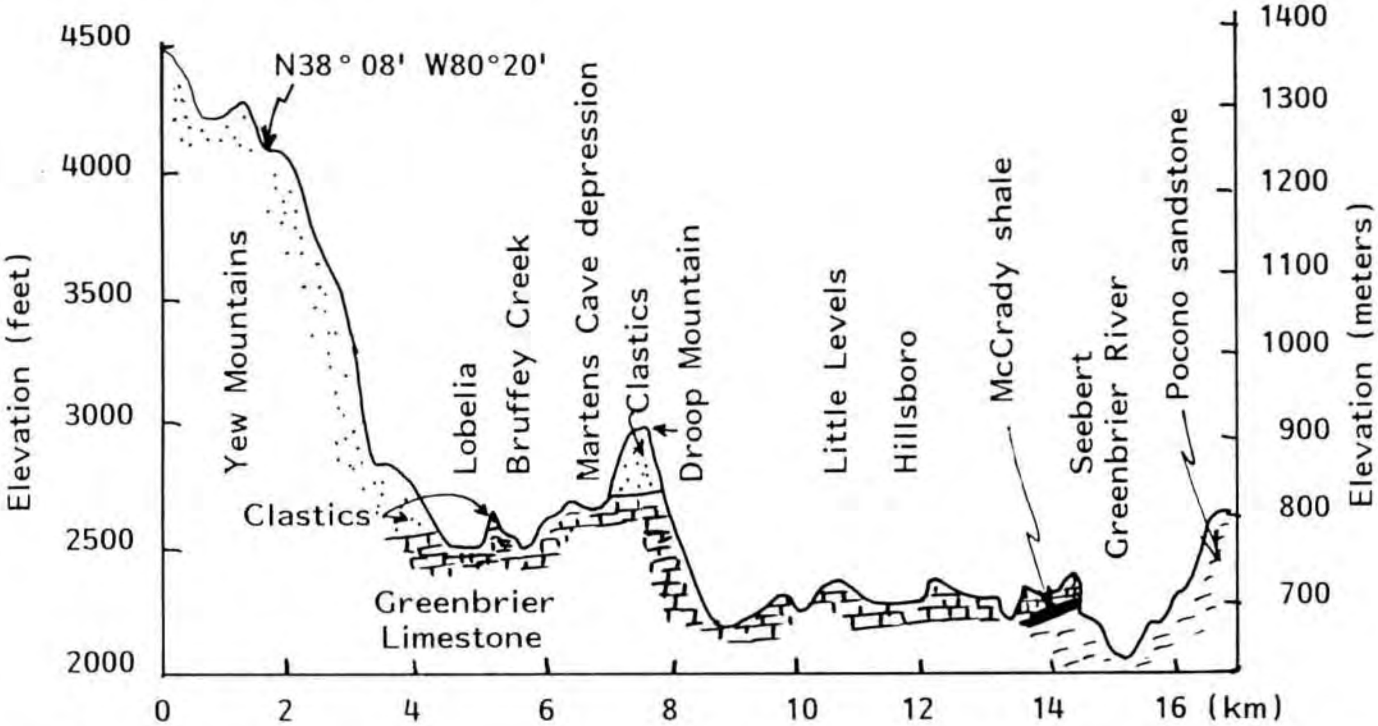

Figure 5: Cross-section drawn east-west through Lobelia and Hillsboro 7.5-minute quadrangles, Pocahontas County, West Virginia. Profile is along N 38°08' latitude.

Ma, unless the assumed value for the denudation rate has changed dramatically over this time span. Obviously, there was a precursor landscape but it is very questionable if the earlier landscape had the appearance of the present surface.

The Highland Rim

A third and final example is the western margin of the Cumberland Plateau in the Cumberland River drainage (Figure 7) (*see also* White and White, 1983). There the Harrisburg Surface is represented by the Highland Rim which is at about 1000 feet (305 m) elevation. The Cumberland Plateau itself varies somewhere between 1800 and 2000 feet (550 to 610 m). The elevation at the top of the Bangor limestone marks the stratigraphic horizon at which development of a karst plain could have started. That is as far up as there was any limestone to dissolve. Dissolution of the limestone surface down to the present-day Highland Rim requires removal of about 150 meters through the whole section. Again we arrive at about 5 megayears as the time at which the Highland Rim butted up against the base of the clastics that make up the top of the Cumberland Plateau, and this defines a maximum age at which anything resembling the present karst surface could exist.

Conclusions

It seems that this brief excursion into the relationship between karst and the Harrisburg Surface has raised more questions than it has answered. Judging from the available evidence, most of the caves of the Appalachians have ages that extend maybe back through the Pleistocene but not much more than that. Many are related to valley systems that formed by the dissection of the valley upland (Harrisburg) surface. The Appalachians have been eroding for a long time but the recognizable form of the Harrisburg Surface appears to be much younger than the traditional geomorphic literature has claimed. Rather than Eocene, a very late Tertiary (maybe late Miocene or Pliocene) age seems to fit the data best for this guiding surface for the Appalachian karst.

The oldest record described in this investigation was 14 Ma, derived from residual soils on the Harrisburg Surface in Pennsylvania. These soils required, however, removal of such a thickness of rock that one must question what changes took place in the other topography, particularly the quartzite-capped mountain ridges, during this period of time. Interpretation of denudation rates on other karst surfaces reveals severe stratigraphic constraints if these surfaces are reconstructed farther back than a few million years. If one means by "age of the surface", the time at which the surface evolved into something recognizably similar to what is seen today, the Harrisburg Surface cannot be older than a few million years. However, there is also the possibility that in spite of recent neo-Davisian interest in erosion surfaces, a Hackian model of continuous evolution may better fit the Appalachian landscape.

Returning to the two quotes cited earlier, can these be reconciled? The Appalachians have been eroding since the

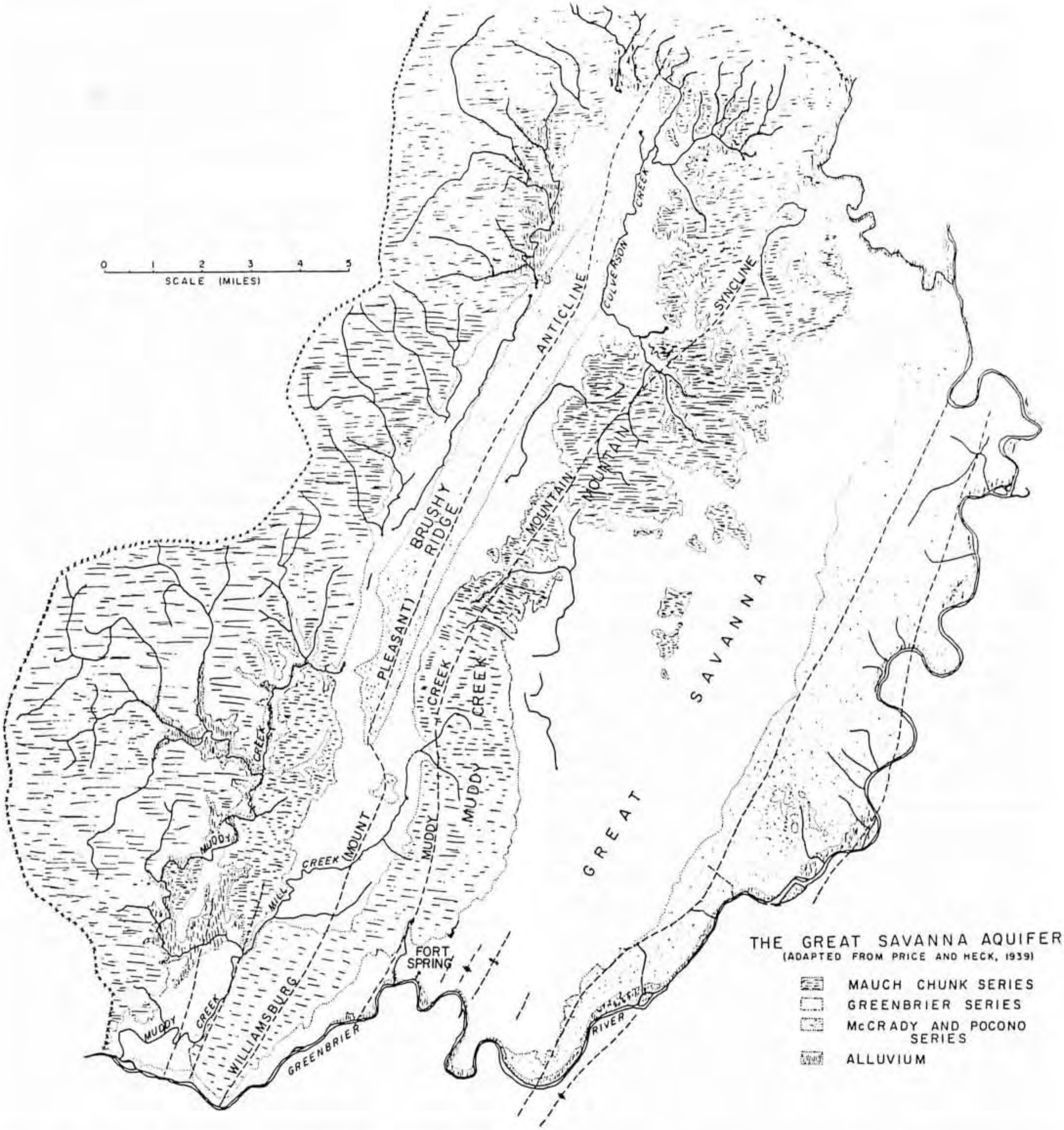

Figure 6: Map showing the "Great Savanna" karst on the Greenbrier Limestone, Greenbrier County, West Virginia.

close of the Paleozoic but there has also been a huge quantity of rock removed. The conflict appears to concern the landscape itself and to what extent present-day topography can be followed backward in time. Evidence from karst investigations supports Thornbury's point of view. The recognizable details of the landscape are relatively young.

References

Davis, N.W. and Hess, J.W., 1982, Hydrogeology of the drainage basin, Burnsville Cove, Virginia: *National Speleological Society Bulletin*, 44, p. 78-83.

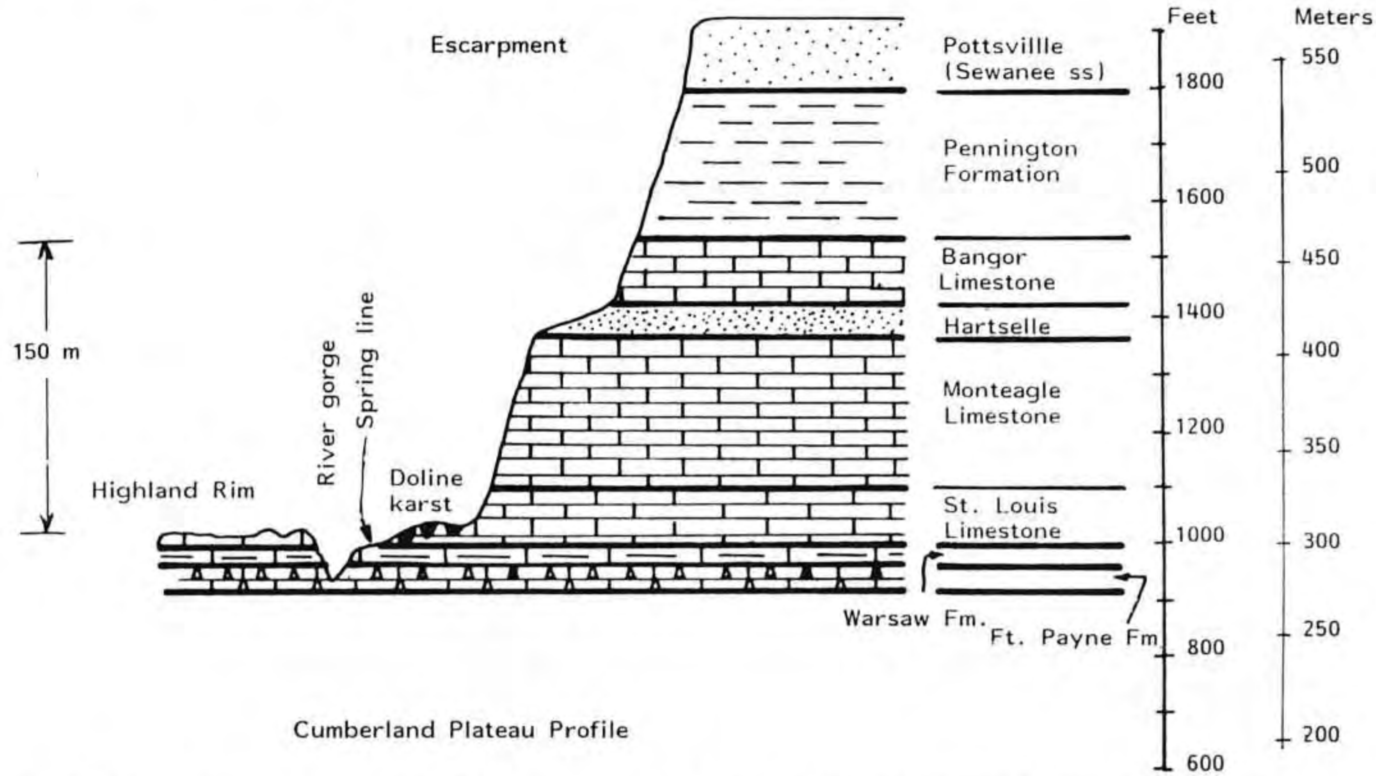

Figure 7: Schematic cross-section through the western margin of the Cumberland Mountains, north-central Tennessee in the Cumberland River drainage.

Davis, W.M., 1889, The rivers and valleys of Pennsylvania: *National Geographic Magazine*, v. 1, p. 183-253.

Deike, G.H., III , 1967, *The Development of Caverns of the Mammoth Cave Region*: Ph.D. dissertation (unpublished), Pennsylvania State University, 235 p..

Fenneman, N.M., 1938, *Physiography of Eastern United States*, McGraw Hill, New York, 714 p.

Hack, J.T., 1960, Interpretation of erosional topography in humid temperate regions: *American Journal of Science*, v. 258-A, p. 80-97.

Harmon, R.S. and Hess, J.W., 1982, Ground water geochemistry of the Burnsville Cove area, Virginia: *National Speleological Society Bulletin*, v. 44, p. 84-89.

Jacobson, R.L., 1973, *Controls on the Quality of Some Carbonate Ground Waters: Dissociation Constants of Calcite and $CaHCO_3^+$ from 0 to 50 C*: Ph.D. dissertation (unpublished), Pennsylvania State University, 131 p.

Jacobson, R.L. and Langmuir, D., 1974, Controls on the quality variations of some carbonate spring waters: *Journal of Hydrology*, v. 23, p. 247-265.

Jones, W.K., 1973, Hydrology of limestone karst: *West Virginia Geological Survey Bulletin*, no. 36, 49 p.

Miotke, F-D., 1975, Der Karst im zentralen Kentucky bei Mammoth Cave: *Jahrbuch der Geographischen Gesellschaft zu Hannover für 1973*, 360 p.

Morisawa, M., 1989, Rivers and valleys of Pennsylvania, revisited: *Geomorphology*, v. 2, p. 1-22.

Ogden, A.E., 1982, Karst denudation rates for selected spring basins in West Virginia: *National Speleological Society Bulletin*, v. 44, p. 6-10.

Palmer, A.N., 1989, Geomorphic history of the Mammoth Cave System (Chapter 12) *in* White, W.B. and White, E.L. (editors), *Karst Hydrology: Concepts from the Mammoth Cave Area*: Van Nostrand Reinhold, New York, p. 317-337.

Parizek, R.R. and White, W.B., 1985, *Application of Quaternary and Tertiary Geological Factors to Environmental Problems in Central Pennsylvania*: Guidebook to 50th Annual Field Conference of Pennsylvania Geologists, p. 63-119.

Price, P.H. and Reger, D.B., 1929, *Pocahontas County*: West Virginia Geological Survey, 531 p. (*see* plate 16B).

Price, P.H. and Heck, E.T., 1939, *Greenbrier County*: West Virginia Geological Survey, 846 p.

Schmidt, V.A., 1982, Magnetostratigraphy of sediments in Mammoth Cave, Kentucky: *Science*, v. 217, p. 827-829.

Sevon, W.D., 1985, *Pennsylvania's Polygenetic Landscape: Guidebook for the 4th Annual Field Trip of the Harrisburg Area Geological Society*, 55 p.

Sevon, W.D., 1989, Erosion in the Juniata River Drainage Basin, Pennsylvania: *Geomorphology*, v. 2, p. 303-318.

Smith, D.I. and Atkinson, T.C., 1976, Process, landforms and climate in limestone regions (Chapter 13) *in* Derbyshire, E. (editor), *Geomorphology and Climate*, John Wiley, London, p. 367-409.

Thornbury, W.D., 1965, *Regional geomorphology of the United States*: John Wiley, New York, 609 p.

White, E.L. and White, W.B., 1983, Karst landforms and drainage basin evolution in the Obey River Basin, north-central Tennessee: *Journal of Hydrology*, v. 61, p. 69-82.

White, W.B. and White, E.L., 1974, Base level control of underground drainage in the Potomac River Basin, *in* Rauch, H.W. and Werner, E. (editors), *Proceedings of the Fourth Conference on Karst Geology and Hydrology*: West Virginia Geological Survey, p. 41-53.

White, W.B. and Hess, J.W., 1982, Geomorphology of Burnsville Cove and the geology of the Butler Cave - Sinking Creek System: *National Speleological Society Bulletin*, v. 44, p. 67-77.

White, W.B., 1984, Rate processes: chemical kinetics and karst landform development (Chapter 10) *in* LaFleur, R.G. (editor), *Groundwater as a Geomorphic Agent*: Allen & Unwin, Boston, p. 227-248.

White, W.B., 1988, *Geomorphology and Hydrology of Karst Terrains*: Oxford University Press, New York, 464 p.

White, W.B., 1990, Surface and near-surface karst landforms (Chapter 7) *in* Higgins, C.G. and Coates, D.R. (editors), *Groundwater Geomorphology: The Role of Subsurface Water in Earth Surface Processes and Landforms*: Geological Society of America Special Paper 252, p.157-175.

Structural Controls on Drainage beneath Droop Mountain, Pocahontas County, West Virginia

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ABSTRACT

Over 8.7 miles (14 kilometers) of surveyed cave passages are found in a 590-foot (180-meter) thick sequence of Mississippian-age carbonates around the northern end of Droop Mountain in southern Pocahontas County, West Virginia. Regional dip is 4 to 5 degrees to the northwest. Whereas primary subsurface flow paths along both the western and eastern sides of Droop Mountain are generally aligned along strike (N 30 E), other flow routes, passing beneath the mountain and crossing the dip, are used as well. This paper compares flow-path gradients, describes relationships among caves, and discusses the influence that faults subparallel to the dip have on flow-path direction. Inconsistencies among measured dip elevations and the top of the Union Limestone on both sides of Droop Mountain are also noted.

Introduction

Droop Mountain is a north-south trending ridge in southern Pocahontas County, West Virginia. Local relief is about 1000 feet with a maximum elevation of 3100 feet (Figure 1). The ridge is capped with Mississippian-age sandstones and shales. It is bounded on the east by the valley of Locust Creek and on the west by the Hills and Bruffey Creek valleys, flowing from higher elevations to the northwest. The northeast flank of Droop Mountain is bounded by north-trending valleys that open to the southern margin of a sinkhole plain to the north (Little Levels).

Middle Mississippian-age carbonates of the Greenbrier Group crop out on the ridge's western, northern, and eastern sides. These limestones, along with interbedded shales and sandstones, are about 650 feet thick. The main limestone sequence in the Greenbrier extends from the Union Limestone to the basal Hillsdale Limestone. The thickness of this sequence is 600 feet, as measured by the Sun Oil No. 1 Droop Mountain well drilled at the northern end of Droop Mountain and described in Leonard (1968). Regional dip is N 60 W, strike N 30 E, and measured dips on both sides of the ridge are 4 to 5 degrees (Worthington, 1984).

The entire limestone sequence is exposed along the mountain's eastern side with the base of the Hillsdale Limestone seen at an elevation of 2100, feet at a spring forming the head of Locust Creek (Locust Spring). To the western side of Droop Mountain, in the Hills/Bruffey Creek valleys, only the upper 50 or so feet of the Union is exposed.

Streams sinking on the western flank of Droop Mountain, *e.g.* Hills Creek, Bruffey Creek, and Rush Run, flow entirely or in part, depending on flow conditions, to the southeast, beneath the Droop Mountain ridge, crossing the dip of the limestones, and rising at cave entrances and springs along the northeastern and eastern sides of the mountain. The area has been well studied since the late 1950's with the basic hydrogeologic setting summarized by White and Schmidt (1966) and early stream tracing carried out by Zotter (1963, 1965). Additional stream tracing has been carried out by Coward (1970), Williams and Jones (1983), and Jones (1991). Speleogenesis of the major cave in the area, the Friars Hole System just to the south, was studied by Worthington (1984) and the structural geology of the Droop Mountain area has been summarized by Jameson (1985) and Kulander and Dean (1978).

Within the past ten years all caves at the northern end of Droop Mountain that have any hydrological significance have been resurveyed. These surveys, totalling nine miles, have been carried out in eight caves to the north and east of the 43-mile Friars Hole System. The areal extent and relative locations of these caves are shown in Figure 2. Profiles from the survey data demonstrate relationships among cave-stream gradients and can be used as indicators of inferred flow routes where such routes have yet to be demonstrated through stream tracing.

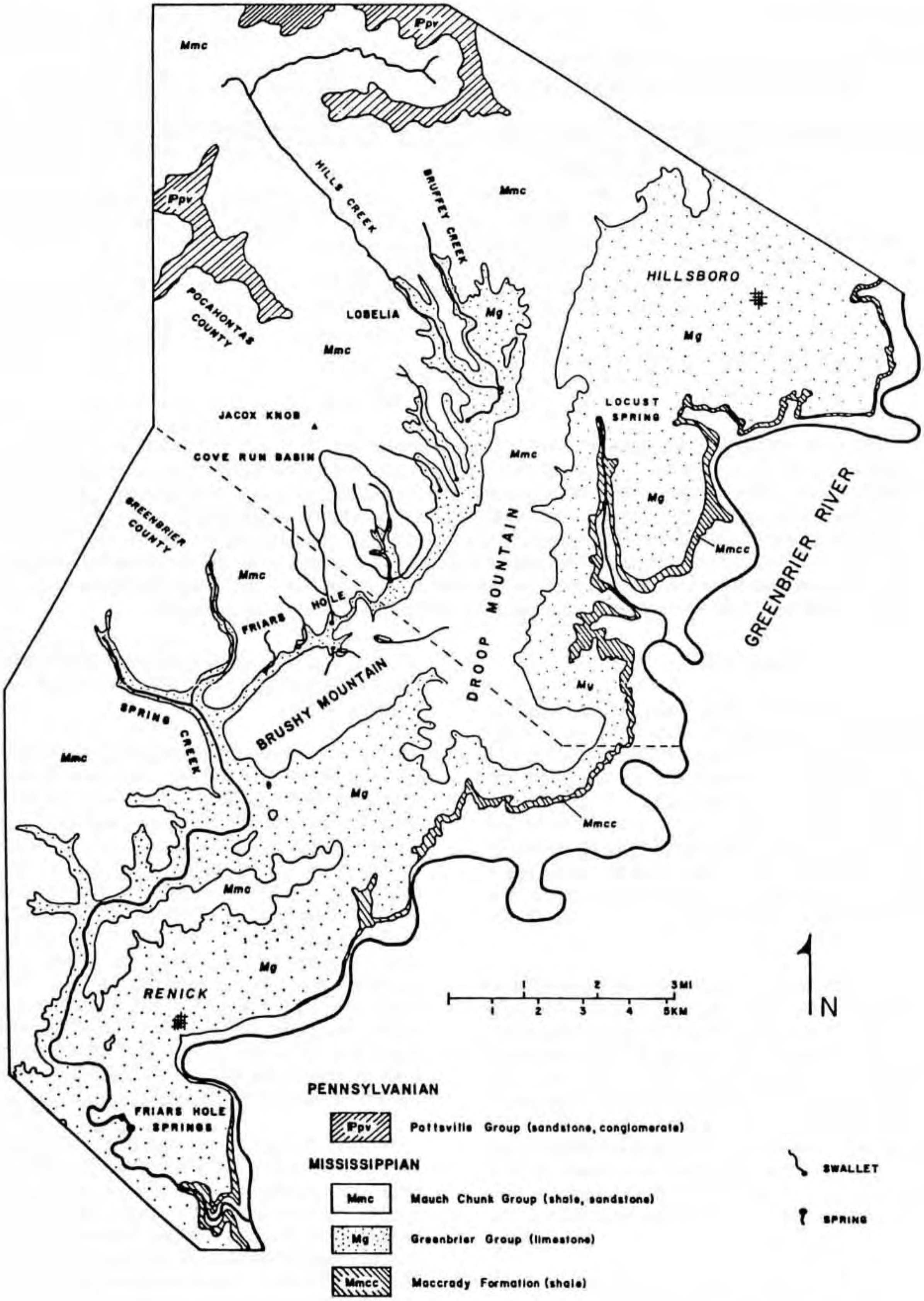

Figure 1: Geologic map of Droop Mountain and vicinity. From Jameson (1985).

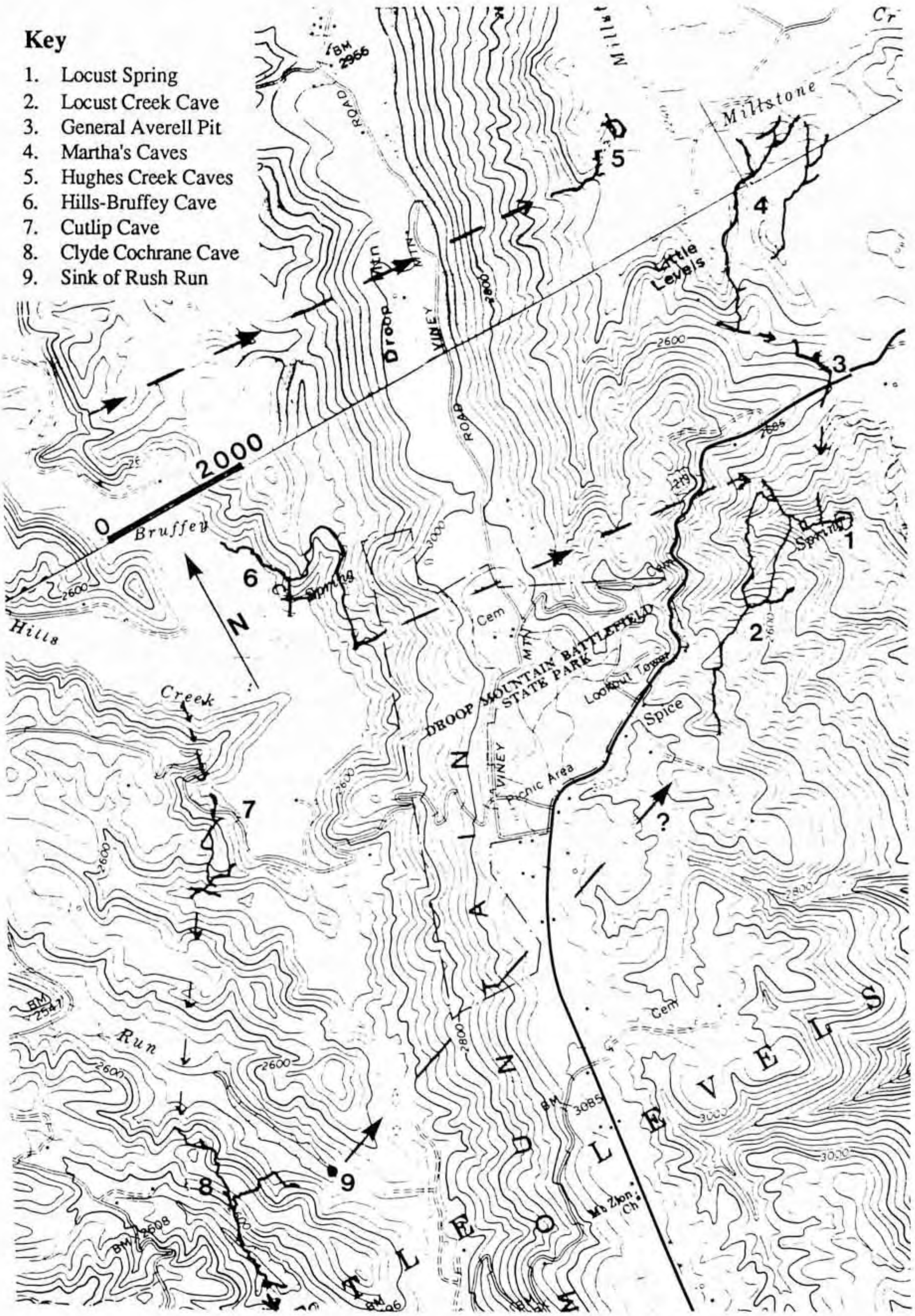

Figure 2: Northern Droop Mountain, showing caves and stream traces.

Structural Setting

The structural setting was summarized by Jameson as follows: "Droop Mountain is near the southern margin of the central Appalachian fold and thrust belt... In this region, first order folds, major thrust faults, and many smaller structures trend about N 30 E. The area is at the southern margin of the Webster Springs block; bounded on the east by the Browns Mountain anticline, on the south by the Modoc lineament, and on the west by the Webster Springs anticline... The Modoc lineament, considered to be somewhat illusory by Dean *et al.* (1979), is a 3 to 7 mile-wide zone which trends east across the regional structural trend of N 30 E" (Jameson, 1985, p. 106, 108).

Faults beneath the northern end of Droop Mountain that strike east-west and may be associated with the "illusory" Modoc lineament are described in this paper. Their influence on both cave-passage termination and on allowing flow to occur across the regional dip are noted. These faults are distinct from the thrusts that strike at N 25-30 E and are observed in many locations in the Friars Hole Cave System just to the south.

An inconsistency between the observed elevations of the top of the Union Limestone on the eastern and western flanks of Droop Mountain and measured dips should be noted; this inconsistency supports Worthington's hypothesis that an anticlinal axis exists beneath the northern end of the mountain (Worthington, 1984, p. 24-26). Measured dips to the northwest, on both sides of the mountain, are 4 to 5 degrees. On the eastern side of Droop Mountain, the bottom of the basal limestone member, the Hillsdale Limestone, is at 2100 feet at Locust Spring. Directly downdip and on the western side of Droop Mountain, the Union/Greenville contact is at an elevation of 2510 feet, directly above the Bruffey Creek Cave entrance. The relative elevations of these contacts are consistent with a uniform dip of 1.5 to 2.0 degrees rather than the 4 to 5 degree dips that are observed. For a consistent 4 degree dip, the top of the Union Limestone should be 200-250 feet lower than the floor of the Hills and Bruffey Creek valleys rather than 50 feet above these valley floors as is the case.

The inconsistency is illustrated in a profile beneath Droop Mountain looking along strike (Figure 3). Profiles of two major caves in the area - Hills/Bruffey Cave on the left and Locust Creek Cave on the right - are shown, along with the 600-foot thickness of the Greenbrier Group between the Union/Greenville contact and the base of the Hillsdale Limestone. A 4 degree dip is assumed and, on the eastern side of Droop Mountain, the location of the Taggard Formation, a prominent shale-limestone-shale marker bed, is shown. This bed is observed in Locust Creek Cave in the location shown on the figure.

Deep-seated thrust faulting and an associated anticline beneath Droop Mountain has been suggested by Worthington (1984) as an explanation for the observed elevations of the Union/Greenville contact. The hypothetical fault plane parallels regional strike with the downthrown side to the west. If this is the case, however, the top of the Union, across an anticlinal axis west of the fault, should be even lower in elevation than is the case for a uniform 4 to 5 degree northwest dip. Whereas westerly dipping, strike-oriented thrusts are observed in caves in the area just west of Droop Mountain, these may be back thrusts associated with the hypothetical fault beneath Droop Mountain. Although displacement along these is minimal, such faults, *en echelon* and beneath Droop Mountain, may account in part for the observed elevations of the Greenbrier Group in the Hills/Bruffey valleys.

Subsurface Drainage

Known flow paths beneath the northern end of Droop

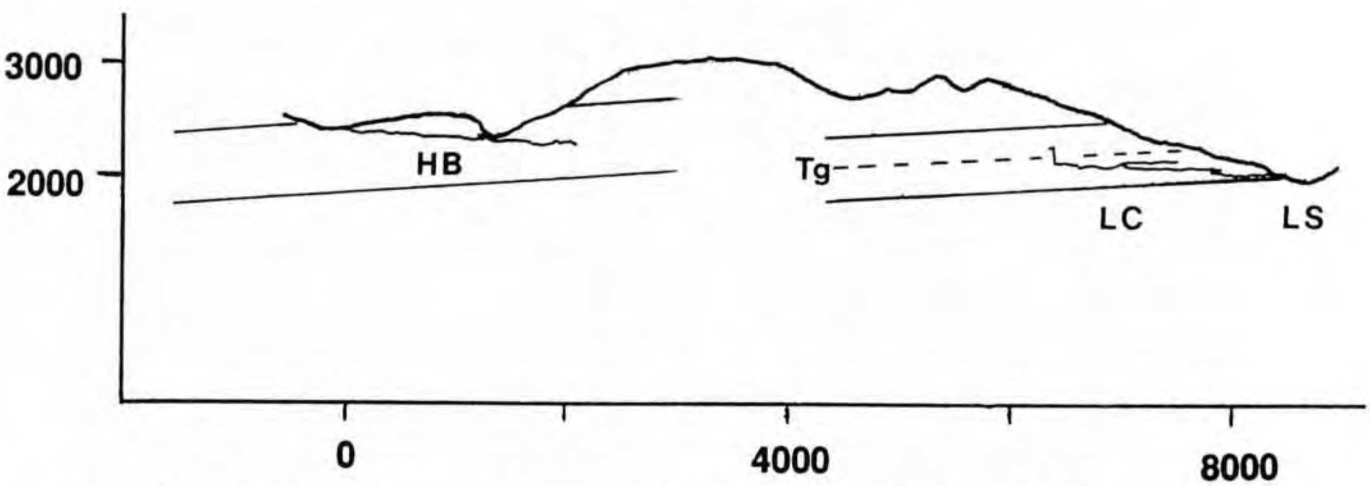

Figure 3: Profile of Droop Mountain looking north-northeast along strike. Key: HB - Hills-Bruffey Cave; LC - Locust Creek Cave; LS - Locust Spring; and Tg - Taggard Formation. Elevations and horizontal scale are in feet.

Mountain are illustrated in Figure 2. To date, three such paths have been identified by Jones and others. The northernmost of these begins with the sinking of Bruffey Creek in its bed at the top of the Union Limestone, about a half-mile north of the entrance to Bruffey Creek Cave. The water flows almost due east for 7400 feet, drops 220 feet, and is seen again in Upper Hughes Creek Cave on the eastern side of Droop Mountain. Although the observed gradient in Upper Hughes Creek Cave and the projected straight-line gradient between the sink of Bruffey Creek and this cave should permit a fairly rapid flowthrough time, Jones (1991) reports a flowthrough time of about a month. This suggests a circuitous route beneath Droop Mountain and/or a substantial amount of ponding between sink and rise. From Upper Hughes Creek Cave, water has been traced to Locust Spring, probably via Lower Hughes Creek Cave, Lower Marthas Cave, and General Averell Pit, although specific cave-to-cave traces have not been carried out.

Hills Creek is also a losing stream and under most flow conditions enters small caves at the base of a ridge on the western side of its valley. This water has been traced for 11.3 miles, more-or-less along strike to the southwest, and through a series of caves (Cutlip Cave, Clyde Cochrane Sinks, and the Friars Hole System) before rising at springs along Spring Creek.

In moderate- to high-flow conditions, however, some of water in both Bruffey Creek and Hills Creek overflows the sinkpoints noted above and flows into cave entrances on the western side of Droop Mountain These entrances are about 50 feet below the top of the Union Limestone and are about 1200 feet apart. The streams can be followed to a junction and the combined flow follows a circuitous route to the northeast and then south. The stream is finally lost in rockfall along an east-west trending fault that terminates the cave.

This overflow route for Hills and Bruffey Creeks (presumably, an older route) passes beneath Droop Mountain, crossing the dip of the limestones. The rising of the Hills-Bruffey overflow waters is at Locust Spring, 9000 feet to the southeast and 480 feet lower in elevation. The flowpath is direct; dye has not been detected at Upper Hughes Creek Cave.

The third path beneath Droop Mountain is that between the ultimate sink of Rush Run and Locust Spring. Rush Run, flowing from the northwest, normally sinks in its bed and is captured by the southwest-trending Cutlip-Clyde Cochrane-Friars Hole flow path noted above. Under high-flow conditions, some of Rush Run passes this sink point and flows into cave entrances at the end of its blind valley, 1/4 mile west of the western flank of Droop Mountain. While most of this water still flows southwest via Clyde Cochrane Sinks, some of it flows east, rising at Locust Spring (Williams and Jones, 1983). As with the Hills-Bruffey Cave water, dye was not detected at Upper Hughes Creek Cave.

Probable Flow Routes

Locust Creek Cave lies behind Locust Spring and contains about 2 miles of surveyed passage. The cave's entrance is flooded and thus the cave is accessible only to divers. Much of the cave consists of large, base-level passages in the lower Denmar Formation. The cave contains three streams representing the downstream ends of the flowpaths described above. All traces conducted to date have been to the cave's entrance; no traces have been completed to the streams inside the cave. Jones (1991) notes that the cave's high-flow drainage area, including the Hills Creek and Bruffey Creek basins, is about 28 square miles and that discharge from the cave has been measured as 2.5 to 350 cubic feet per second.

The cave contains three streams, all of which join upstream of the sumped entrance passage. The major source of water in the cave is a stream entering from the north. This may be the Hills-Bruffy water; the volume is consistent with that seen in Hills-Bruffey Cave and the hypothetical route is direct; a continuation of the easterly trending flow of the Hills-Bruffey water. The straight-line distance is 5400 feet and the vertical difference is 320 feet (gradient is 0.06).

A smaller stream in Locust Creek Cave flows from the southwest and probably represents water sinking in valleys on the eastern side of Droop Mountain. This stream, however, may also contain the water traced from Rush Run. Figure 4 is a profile looking N 25 W and shows the relationship between the sink of Rush Run and Locust Creek Cave. The viewing direction is perpendicular to a straight line between the sink of Rush Run and the upstream end of Locust Creek Cave's southwest stream. Assuming a fairly direct flow path to the northeast, the hypothetical gradient (0.024) is consistent with that seen in the cave stream.

Finally, the Bruffey Creek streambed water, rising at Upper Hughes Creek Cave, has also been traced to Locust Spring. Whereas the probable flow path is via Lower Marthas Cave and then General Averell Pit, this has not been confirmed by tracing. Also, the specific stream within Locust Creek Cave where this water rises has not been identified through stream tracing, although it is probably the stream entering from the north. Figure 5 is a profile of these caves, looking from the east. The relative elevations of the streams in the caves and the gradients of these streams (0.02) are consistent with a single flow path. The location of the large northern stream entering Locust Creek Cave is indicated by an "N" in this figure.

Origin of Flow Paths

The existence of three distinct and more-or-less parallel flow paths beneath Droop Mountain, each of which

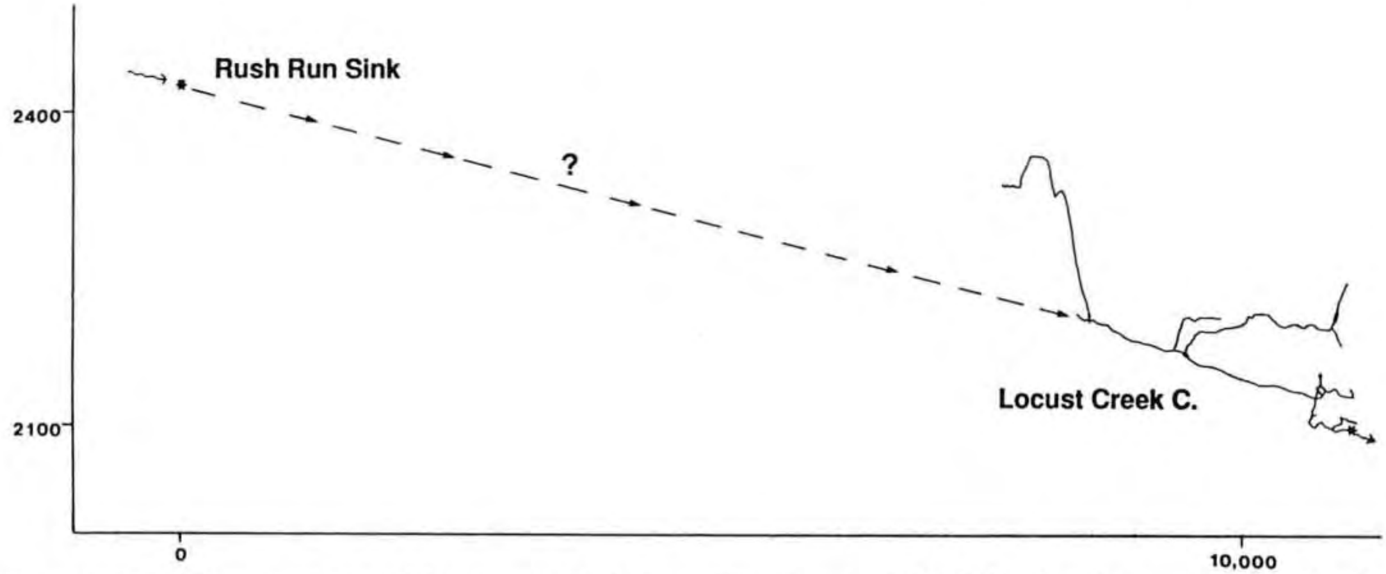

Figure 4: Profile from Rush Run Sink to Locust Creek Cave. View direction is N 25 W. Elevations and horizontal scale are in feet.

crosses the regional dip and drops through up to 600 feet of carbonates and interbedded clastics, is striking. The predominant directional orientation for solutional conduits in the area is along strike (NNE-SSW), following enlarged bedding-plane partings. Cave passages trending normal to the strike tend to be high-gradient inlets following joints with the flow direction downdip to the northwest. The occurrence of traversible conduits containing streams flowing across the dip is infrequent, although such conduits must exist, given the results of the stream traces carried out.

In some of the caves at the north end of Droop Mountain, several passage segments trending nearly east-west and containing easterly flowing streams, are accessible. In other caves, east-west passage segments, not necessarily containing streams, are also accessible. In both cases, these passage segments parallel high-angle thrust faults that strike east-west to east-southeast/west-northwest, *i.e.*, subparallel to the regional dip. It is hypothesized that these faults provide a structural mechanism for solution to be initiated and subsurface flow to occur beneath Droop Mountain. The dip of the faults is usually to the north or

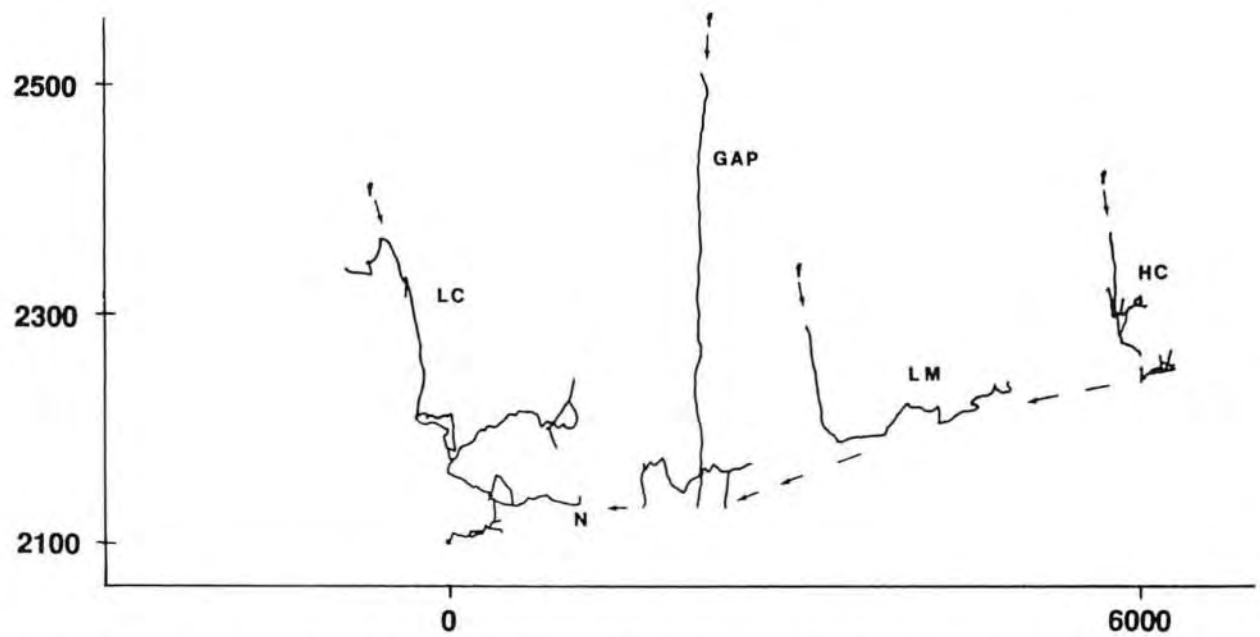

Figure 5: Profile showing caves draining to Locust Spring. View direction is looking west. Key: LC - Locust Creek Cave; GAP - General Averell Pit; LM - Lower Martha's Cave; HC - Hughes Creek Caves; and f - ESE-WNW trending thrust fault. Elevations and horizontal scale are in feet.

NNE, although in one case, the dip is to the SSW. The fault planes are almost perpendicular to those of the previously observed strike-oriented faulting in caves in the area. Examples of these faults are given below.

(a) *Hills-Bruffey Cave:* The stream in this cave meanders to the northeast and then south for over 6000 feet following an enlarged bedding-plane parting in the Union Limestone. It then turns abruptly to the east and is lost in breakdown along a fault. The cave passage, however, can be followed for another 425 feet to the east following this fault, which strikes N 80-85 W and dips to the north at 35 to 45 degrees.

(b) *Upper Hughes Creek Cave:* For over 800 linear feet, the cave's stream flows from west to east along a fault striking N 85 W and dipping to the north at 40 to 50 degrees. This is the longest traversible segment of such passage in the Droop Mountain caves. As previously noted, the cave stream has been traced from the bed of Bruffey Creek, on the west side of Droop Mountain.

(c) *Lower Marthas Cave*, trending SW along strike for 3000 linear feet, ends at a fault striking N 60 W and dipping northeast at 45 degrees. At this point, streamflow turns to the southeast along the fault although, as is the case with other passages along these faults, the stream is soon lost in rockfall.

(d) The entrance to *General Averell Pit*, on the northeastern flank of Droop Mountain, follows the plane of a thrust fault downward for 360 vertical feet to a stream segment. The fault strikes N 60 W and dips steeply (60 to 70 degrees). The dip direction, however, is to the SSW, in contrast to the fault terminating the south end of Lower Marthas Cave. A few hundred feet of traversible passage, containing a stream (almost certainly that seen in Lower Marthas) parallels the plane of the fault.

(e) *Cutlip Cave*, on the western side of Droop Mountain and containing that part of the underground Hills Creek that flows to the southwest, terminates in rockfall along a fault that strikes at N 60 W and dips to the NNE at 20 to 30 degrees. The cave stream is lost in rockfall and continues southwest toward Clyde Cochrane Sinks. In this case, east-west faulting only affects passage morphology and does not alter the flow direction.

(f) At the upstream termination of the stream flowing from SW to NE in *Locust Creek Cave*, an upper passage ascends very steeply over rockfall, climbing over 120 vertical feet and breaching the Taggard shales. While this may be an example of a passage ascending along a northerly dipping fault, additional evidence is needed.

The dips of the generally east-west trending faults described above are apparent when viewing the profiles of these caves from the east; *i.e.*, along the strike of the faults. This can be seen in Figure 5 where, for examples (b), (c), (d), and (f) above, the arrows labeled "f" indicate the locations of the faults. Whereas the faults exist and apparently provide routes for water to follow when passing beneath Droop Mountain, their origin and relationship to other regional structures is a problem for further study. It is worth noting that these faults parallel and perhaps coincide with the northern part of the postulated east-west trending Modoc lineament and indeed, could be used as evidence for the existence of this linear zone. Corroborating evidence such as surficial fault development or changes in axial trends, as suggested by Dean and others (1979), or changes in the alignments of surface streams has been neither observed nor sought in the area.

Summary

Cave development beneath northern Droop Mountain is a fairly complex process (Jones, 1983) involving stream piracy from one subsurface route to another, flow both along strike and across dip, and a degree of control over the direction of flow by thrust faults that both parallel the strike and are subparallel to the regional dip. Whereas the overall subsurface drainage pattern has been established, much work remains to be done concerning cave-to-cave flow routes and time-discharge relationships. Also, until such time as traversible conduits beneath Droop Mountain are found, the extent and nature of those structures beneath the mountain that influence cave development and flow direction can be hypothesized and discussed, but not observed.

References

Coward, J.M.H., 1975, *Paleohydrology and Streamflow Simulation of Three Karst Basins in Southeastern West Virginia, U.S.A.*: Ph.D. thesis (unpublished), McMaster University, Hamilton, Ontario, 394 p.

Dean, S.L.; Kulander, B.R.; and Williams, R.E., 1979, Regional Tectonics, Systematic Fractures, and Photolinears in Southeastern West Virginia: *Proceedings of the Second International Conference on Basement Tectonics, Newark, Delaware, 1976*, p. 10-53.

Jameson, R. A., 1985, *Structural Segments and the Analysis of Flow Paths in the North Canyon of Snedegars Cave, Friars Hole Cave System, West Virginia*: M.S. thesis (unpublished), West Virginia University, 421 p.

Jones, W.K., 1983, Evolution of the Hills Creek basin, West Virginia (abstract): *Program of the National Speleological Society Annual Convention, Elkins, West Virginia*, p. 69-70.

Jones, W.K., 1991, Locust Creek Cave hydrology, *NSS News*, v. 49, p. 139.

Kulander B.R. and Dean, S.L., 1978, Gravity, Magnetics, and Structure of the Allegheny Plateau/Western Valley

and Ridge in West Virginia and Adjacent States: *West Virginia Geological and Economic Survey, Report of Investigations*, no. 27, 91 p.

Leonard, A.D., 1965, *The Petrology and Stratigraphy of Upper Mississippian Greenbrier Limestones of Eastern West Virginia*: Ph.D. dissertation (unpublished), West Virginia University, 245 p.

White, W.B. and Schmidt, V.A., 1966, Hydrology of a karst area in east-central West Virginia: *Water Resources Research*, v. 2, no. 3, p. 549-560.

Williams, C.F. and Jones, W.K., 1983, Karst drainage systems of the northern Spring Creek basin, West Virginia (abstract): *Program of the National Speleological Society Annual Convention, Elkins, West Virginia,* p. 92-93.

Worthington, S.R.H., 1984, *The Paleo Drainage of an Appalachian Fluviokarst: Friars Hole, West Virginia:* M.S. thesis (unpublished), McMaster University, 218 p.

Zotter, H., 1963, Stream tracing techniques and results: Pocahontas and Greenbrier Counties, West Virginia: *NSS News*, v. 21, no. 10, p 136-142.

Zotter, H., 1965, Stream tracing techniques and results: Pocahontas and Greenbrier Counties, West Virginia, Part 2: *NSS News*, v. 23, no. 12, p. 169-177.

Karst and Caves of Mercer and Summers Counties, West Virginia

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ABSTRACT

Caves and karst drainage occur at four general stratigraphic levels in Mercer County: The Cambro-Ordovician dolomites, mid-Ordovician limestones, Mississippian Greenbrier limestones, and upper Mississippian Avis Limestone. In Summers County only the latter two are exposed. The dolomites run the length of East River Mountain, dipping into the mountain a thousand feet or more below the lengthy mountain crest. Most mountainside drainage above the lower exposures of the dolomites is captured and drains to springs in those lower exposures; several caves less than 600 feet long are known there. The overlying mid-Ordovician limestones host plentiful mountainside dolines and ponors. The longest known cave by far is three-mile-long, strike-aligned Beacon Cave; there is a swing and branching in the St. Clair Fault and mountainside drainage discharges from limestone. The Greenbrier limestones are exposed as part of an overturned syncline along the length of the foot of East River Mountain and also in the Abbs Valley Anticline. Here, two major features are a water-supply cave draining Interstate 77, and a 4000-foot, straightline piracy of Big Spring Branch. The Avis Limestone is up to 40 feet thick, mostly relatively low dip, doline-scarce, and exposed in steeply dissected terrain in eastern Mercer County and most of Summers County. Cave passages in the Avis are highly joint-controlled, often mazy, and associated with piracy of nearby streams. Jones Cave (5655 feet long), where a landslide was instrumental in the cave's development, has two insurgences, a flood-outlet entrance and a perennial-spring entrance. Brickyard Ridge Cave (2450 feet long) is a complex maze developed in the plunge of the Abbs Valley Anticline, and has a high-water outlet. There are few perennial springs in the Avis, but Adair Run valley has three of them along its length, where the limestone drops 700 feet in two miles. At this point, below the valleyside exposures of the Avis, several segments of almost half of the two-mile course are often dry, and subsurface flow occurs in the Payne Branch Sandstone. An 820-foot maze in the Avis Limestone is the longest cave known in Summers County.

Introduction

Mercer and adjacent Summers County lie in southern West Virginia (Figure 1) and are two of the state's lesser known karst counties. Neither county has been involved in any of the publications of the West Virginia Speleological Survey. Davies (1958) did not cover Summers County at all in his compendium on the state's caves, and some of the information listed for Mercer County was in error. Even today parts of both counties containing caves are unknown to karst researchers and cavers.

Structure

Mercer County contains the southernmost extension of the Valley and Ridge Province in West Virginia. The southern edge of the county comprises the outcrop slope of the 23-mile long East River Mountain that continues into Virginia, one of the Appalachian long mountains whose karst was described by Saunders and others (1977). The longest major fault in the county, the St. Clair, parallels the flank of the mountainside for the entire length of East River Mountain (Reger, 1926). Just north of the fault line lies the Hurricane Ridge Syncline, which is overturned on its southern limb. Further north, the structure grades into the gentle folds of the Appalachian Plateau Province which predominates in Summers County. Just northwest of Bluefield is the very pronounced Abbs Valley Anticline. This anticline rapidly plunges to the northeast and is the structural extension of the Boissevain Fault that dies out

Figure 1: Location of Mercer and Summers counties, West Virginia. Summers County is northeast of Mercer County.

as it crosses into Mercer County from Virginia. The topographic expression of this anticline is not apparent north of Princeton even though the fold continues at depth into Monroe County.

Stratigraphy

Caves and karst drainage occur at four different stratigraphic horizons in Mercer County. The Cambro-Ordovician dolomites, up to 3500 feet thick, crop out along the length of East River Mountain, dipping fairly uniformly into the mountain crest. The lower exposures of the Cambro-Ordovician dolomites are located along the St. Clair fault line, which limits the proportion of the dolomites involved. The overlying mid-Ordovician limestones are approximately 100 feet thick, with the pink and white, thin-bedded, sandy Moccasin Limestone in the upper parts. Above the limestones are calcareous shales as well as sandstones, with the Tuscarora Sandstone forming the crest of the mountain. The Mississippian Greenbrier Limestone crops out in a narrow band on the southern limb of the overturned Hurricane Ridge Syncline that parallels the St. Clair Fault, as well as in a narrow, six-mile long strip along the crest of the Abbs Valley Anticline in the western part of the county. The Greenbrier is thicker (up to 1700 feet) and more argillaceous in Mercer County than in several counties along its outcrop belt to the north. The Upper Mississippian Avis Limestone (Hinton Group) reaches a thickness of up to forty feet in parts of Mercer and Summers counties. Some parts of the Avis are highly argillaceous, and even the beds most conducive to cave development have characteristic irregular, thin shale partings apparent throughout. The Avis Limestone is bounded below by a sharp contact with fissile red or yellow shale, 20 to 30 feet thick, and above by the Upper Avis Shale that is up to 40 feet thick.

Surface Expression

The Cambro-Ordovician dolomites are found in doline-poor terrain. The mid-Ordovician limestones crop out in doline- and ponor-rich terrain, often with benches and small, strike-aligned valleys of gentler slope. In the Greenbrier Limestone, dolines are common and limestones form valleys. The Avis Limestone crops out very prominently in steep, dissected terrain, usually as "turrets" up to ten feet long. Sandstone rubble covers the outcrop in many places, especially on the gentler slopes. Dolines are rare in the Avis.

Caves

Two caves are known in the Cambro-Ordovician dolomites. Big Spring Cave (500 feet long) and Tank Wilson Cave (approximately 60 feet long) both end in sumps. Big Spring Cave has a high-gradient stream with waterfalls.

Beacon Cave (Figure 2), more than three miles long, is by far the longest cave in the mid-Ordovician limestones in Mercer County. It has a major, strike-aligned trunk over 300 feet long that carries a large stream that sumps several hundred feet short of its final resurgence in 200-foot long Beaver Pond Spring Cave. Large blocks at the entrance to the Beaver Pond Spring Cave appear to have dammed the water and caused several feet of depth in the sump. Removal of the blocks might permit another access to Beacon Cave. Cave Rat Cave (3200 feet long) is a complex, but shallow, cave farther east on the mountain, entered at a ponor. KFC Cave, located behind the Kentucky Fried Chicken restaurant in Bluefield, drains the parking lot and is a small maze (500 feet long) that follows the dip. A 37-foot drop with a waterfall is developed near the end of the cave, beyond which the stream sumps in a rubble-filled joint. Several pits with depths of 60-150 feet are also found in the mid-Ordovician limestones. Holiday Hole, located adjacent to the parking lot of the Bluefield Holiday Inn, is a 115-foot deep breakout dome with a 70-foot entrance shaft into a large breakdown-floored room. Whispering Falls Pit, near Oakvale, is an enlarged joint measuring 120 feet long by 30 feet wide and 60 feet deep. A stream flowing off East River Mountain cascades into the pit as a 50-foot waterfall and sinks in rubble at the bottom.

The longest known cave in the Avis Limestone is Jones Cave in Mercer County (Figure 3), with four entrances and 5655 feet of passage, over a straight line distance of 3500 feet. Complex passage areas are found primarily in the vicinity of entrances, where dripping along ceiling cracks has created joint-orientated loops and deadends. Palmer (1975) identified this type of diffuse infiltration as a mechanism of maze-cave formation. A landslide composed of sandstone rubble at the upstream end of the cave appears to have blocked a valley and also been instrumental in the cave's development, as the pirated waters flow generally

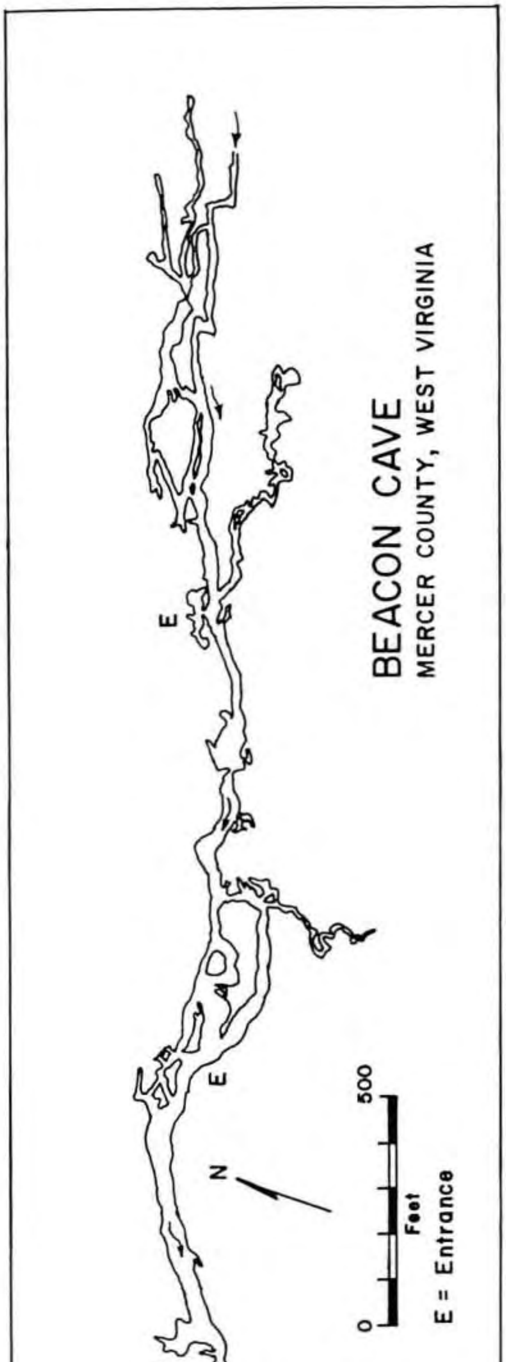

Figure 2: Beacon Cave, Mercer County. This cave is aligned along bedrock strike. Original Survey by V.P.I. Grotto, 1971-1974.

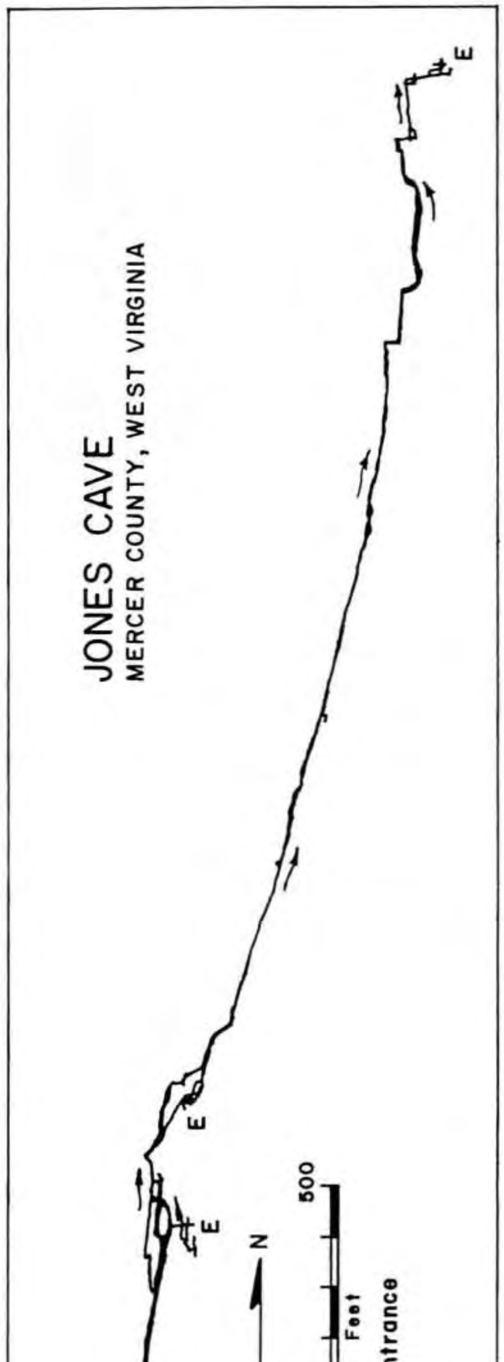

Figure 3: Jones Cave, Mercer County. The cave is the longest in the Avis Limestone. Map from Saunders and Koerschner, 1976.

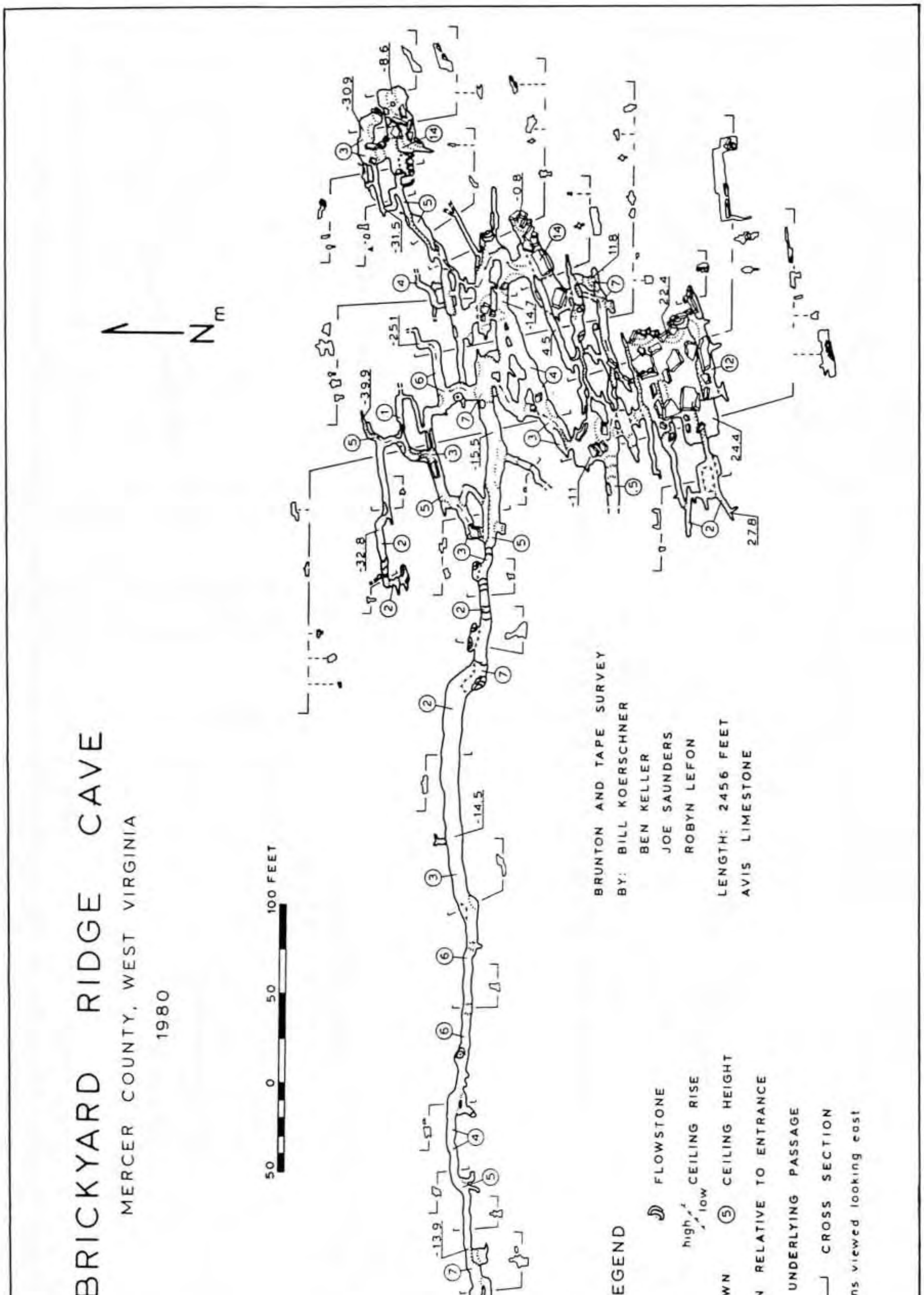

Figure 4: Brickyard Ridge Cave, Mercer County. This is a typical maze cave from the Avis Limestone.

downdip through the hillside to emerge in a parallel valley (Saunders and Koerschner, 1976). Brickyard Ridge Cave (2500 feet long) has a complex maze connected by a straight conduit to the only entrance known in the Avis that neither takes nor discharges flow (Figure 4). Brickyard Ridge Cave is formed along the plunging axis of the Abbs Valley Anticline, just before the Avis dips below drainage. Of all the Avis caves, only Jones and Brickyard Ridge have significant fossil passages (*i.e.* not presently carrying flow). All caves in the Avis are strongly joint controlled and all except Brickyard Ridge have a stereotypical vadose-spring entrance.

In Summers County, the longest known cave is 820-foot long Neely's Nose Cave, a maze developed in the Avis Limestone. At Barger Springs, the along Greenbrier River, there is a cave by the same name that is developed in the Alderson Limestone, the uppermost member of the Greenbrier series. This cave is approximately 600 feet long. Two topographic features in the county refer to caves: Cave Ridge may be named for 110-foot long Barker Hollow Cave in the Avis Limestone, and Cave Branch appears to be named after a large shelter cave in the Avis Sandstone.

To date there are 50 recorded caves in Mercer County, 20 in the Avis Limestone, 7 in the Greenbrier Limestone, and 23 in Ordovician carbonates. In Summers County there are 13 recorded caves, two large shelter caves in sandstone, 10 in the Avis Limestone, and one in the Greenbrier.

Subsurface Drainage Patterns

Most, if not all, streams flowing off the upper slopes of East River Mountain sink into the mid-Ordovician limestones during all but high-flow conditions. Most discrete sink points do not permit human entry; many are in sinkholes choked with sandstone float from the overlying caprock. Many stream sinks are located stratigraphically in the highest limestone unit, the Moccasin. The sinking streams do not occupy all of the upper-slope ravines on East River Mountain and preliminary examination has found that this higher drainage is confluent, exiting primarily from solitary springs in as yet unidentified stratigraphic units somewhat above the Moccasin.

By number and greater proportion of total discharge, most significant springs in the carbonates along East River Mountain are found in the Cambro-Ordovician dolomites. Most examined so far do not have associated caves. Some, like Big Spring Cave spring and Tank Wilson Cave spring are located immediately above the St. Clair Fault, at the lowest available exposure of the dolomite. Many other springs in the dolomite are located somewhat above the lowest possible exposures. As noted earlier from more preliminary generalizations (Saunders and others, 1977), failure of many of the discharge points to be found at the lowest carbonate level is mostly a reflection of the dolomite lithology in the long mountain setting.

The only possible overflow outlet known along East River Mountain may be another example of the low permeability of the dolomites. Pigeon Creek, south of Oakvale, rises at a spring in dolomite that is well above the St. Clair Fault. A high-water outlet in dolomite exists a thousand feet up the valley. Yet another thousand feet farther up the valley is an outlet in the limestone with intermittent flow, suggesting that it is partially related to flow in the intermittent outlet in dolomite.

The largest limestone spring along East River Mountain is the Beaver Pond Spring, the outlet for the Beacon Cave stream. Nearby is the most significant irregularity in the St. Clair Fault along East River Mountain in West Virginia. The fault branches and swings closer to the mountain crest. It may not be coincidence that the longest and most accessible limestone cave along the mountain is located here.

No water tracing has been performed in the East River Mountain karst. Based on models of long, underground-drainage basins proposed by Saunders and others (1977), two quite different configurations of underground flow can be visualized and may co-exist in some of the underground-drainage basins in East River Mountain. One scenario, seen in Beacon Cave, involves essentially strike-orientated flow along and beneath the mountainside in subsurface basins of elongate shape. A contrasting situation involves main flow generally diagonal or perpendicular to the strike, against the dip, away from the mountain crest. Tributary flow in both types is often down components of the dip and into the mountain. Limestones usually host strike-aligned main flow in the long mountains, as exemplified by Beacon Cave. In contrast, main flow against the dip appears to occur primarily in the dolomite, creating shorter, but wider, basins. Unfortunately, the actual conduit orientations, if conduits exist, are largely unknown due to limited access to passages. However, Big Spring Cave is like several other dolomite caves elsewhere in the long mountain setting; it has a steep gradient with waterfalls and cascades, in contrast to the typical, nearly flat, gradient seen in limestone caves in the long mountains.

The Greenbrier Limestone is exposed in two linear bends in Mercer County, one of which runs the length of the county along the flank of the Hurricane Ridge Syncline. Significant hydrological features are evident at two locations in the latter. The eastern Big Spring Branch sinks entirely for part of the year into its bed southwest of Kellysville and appears to resurge 4000 feet down strike to the east along the East River. Secondly, Ingleside Cave carries a domestic water supply stream that would appear to drain nearby Spangler Valley, a doline-rich area through which Interstate Highway 77 runs.

Most known caves in the Avis Limestone are associated with sinking streams and most cave entrances in the

Avis are springs. Most of the sinking streams and springs are seasonal in nature. All but the two longest caves in the Avis can be described as simple dendritic-flow systems with a single discharge point at the solitary cave entrance. The second longest Avis cave, Brickyard Ridge Cave, has both a perennial and a high-water outlet. The longest cave, Jones, also appears to be the most hydrologically complex. There are two streams sinking into the cave, as well as a high-water outlet behind a passable collapse zone, well upstream of the perennial spring. All four locales provide entrance to the cave. It appears that increasing cave length is associated with increasing hydrologic complexity.

Other than the association of Brickyard Ridge Cave with a steeply plunging anticline, no predictive relationships of Avis Limestone cave location with structure or topography have been formulated. Given the low-dip characteristic of most of the Avis outcrop area, it can not be concluded that dip is a factor in development of the commonly short (less than 900 feet long) and complex Avis caves. The ability of the Avis Limestone to conduct flow over long distances would entail capture of drainage on updip sides of mountains and downdip discharge on the other side of the same mountains. Prospective flow under Bent or Tallery Mountains, for example, would occur along flow paths of at least three miles in length.

Perhaps the most interesting, as well as unusual, area of subsurface drainage in the two-county area is in Adair Run valley, north of Kellysville in Mercer County. Situated in a structurally transitional area between the overturned syncline at the base of East River Mountain and the gentle dip to the north, the valley follows the dip, losing 700 feet in two miles. Just south of Elgood, the headwaters of Adair Run cross the Avis Limestone, sinking and rising within a hundred feet. Shortly downstream, flow in the dry month, April 1988, sank into the Payne Branch Sandstone underlying the lower Avis Shale. Flow finally appeared amidst cobbles above the confluence with Woodall Fork (approximately two miles distant), which itself flows a short distance in the Payne Branch Sandstone before discharging from cracks in the sandstone bedrock floor of Adair Run. For the entire length of the valley, the Avis limestone is 20 to 50 feet above the valley floor. Three small caves are known in the Avis Limestone on the northeast side of Adair Run, plus a thousand-foot piracy of a tributary, an unexplored cave, and two water-supply springs on the southwest side near the lower end of the 2.5-mile long valley.

Summary

In summary, karst exploration in Mercer and Summers counties has begun to locate and define caves and karst drainage in the long mountain setting as well as in an isolated limestone confined primarily to these two counties. Four stratigraphic levels have been identified where caves and karst are found. Much information is yet to be learned because these two counties have long been neglected in favor of the more traditional karst areas of West Virginia.

References

Davies, W.E., 1958, *Caverns of West Virginia*: West Virginia Geological Survey, v. 19, 330 p.

Palmer, A.N., 1975, The origin of maze caves: *National Speleological Society Bulletin*, v. 37, p. 57-76.

Reger, D.B., 1926, Mercer, Monroe and Summers Counties: *West Virginia Geological Survey County Report*, 963 p.

Saunders, J.W. and Koerschner, W.F., III, 1976, West Virginia's Jones Cave: *NSS News*, v. 34, p. 6-10.

Saunders, J.W.; Medville, D.M.; and Koerschner, W.F., III, 1977, Karst drainage patterns in the long mountains of the eastern United States, *in* Ford, T.D. (editor), *Proceedings of the 7th International Speleological Congress, Sheffield, 1977*: British Cave Research Association, Somerset, England, p. 375-376.

Hydrochemical and Structural Controls on Speleogenesis in the Appalachian Foldbelt*

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ABSTRACT

The formation of caves is affected by composition and porosity of rock, structure of bedrock, and the relation of caves to groundwater flow and surrounding topography. Caves in Cave Hill, Augusta County, Virginia were investigated in order to determine the relative effects of each of these factors in the folded Appalachian Mountains.

Part of this study was a determination of porosity of samples taken from Grand Caverns, a cave that cuts through many vertically dipping beds in Cave Hill. Previously, these samples were analyzed for content of calcite and insolubles. Results indicate that beds of rock that are purest in calcite and microcrystalline are the least porous, and those that are low in calcite content and contain mostly insoluble materials are the most porous. That the most soluble rocks are the least porous or permeable implies that the groundwater flow that dissolved them may have been carried through the rock sequence along adjacent highly porous and permeable beds of insoluble material, such as interbedded sandstones. This mechanism for speleogenesis in rocks of mixed lithologies is proposed as a new hypothesis that may have application in many other cave regions.

A second part of the study was to produce a geologic map of Cave Hill. The resulting map, incorporating measurements of bedrock attitude and position and elevations of caves, explains much of the variation seen among caves in the hill.

Introduction

The origin of caves has been of interest for some time and many of the details have been studied in various geographic and geologic settings (White, 1988; Ford and Williams, 1989; Palmer, 1990, 1991). However, at some localities, additional information can come to light with precise fieldwork. Recent visits to Grand Caverns, located in Cave Hill in northeastern Augusta County, Virginia, indicated that this would be an excellent site to investigate the roles of bedrock composition, porosity, and structure in cave development in the Appalachian Mountain region. Because Cave Hill is a miniature version of the ridges of the Appalachian Mountains that run northeast and southwest through western Virginia, processes of cave development here should be similar to those that have occurred throughout the mountain range. Therefore this study should add to the knowledge about cave development in folded rocks of this region.

The first phase of this study, completed in 1990, analyzed selected samples of the Conococheague Formation of late Cambrian age that is host to at least eight caves on Cave Hill. A very strong correlation between calcite content of the rock and passage enlargement was established (Kastning, 1990a,b).

Several caves of various sizes have formed in Cave Hill (Figure 1). They differ significantly in shape and size even though they are located very close to each other along a 915-meter stretch of the hill. There must be basic geologic reasons for these differences. This study is an investigation of factors that have given each cave its geometric characteristics, including (1) the porosity and permeability of the bedrock, (2) the structure of the bedrock (folds, faults, joints), and (3) the relationships of caves to the surrounding land (topography) and how the origin of the caves may have been affected by changes in the topography during the geologic past.

*High school science fair project. Author was a finalist and award winner at the 1990 and 1991 International Science and Engineering Fairs, and recipient of a 1991 Naval National Science Award and the 1990 James G. Mitchell Award of the National Speleological Society.

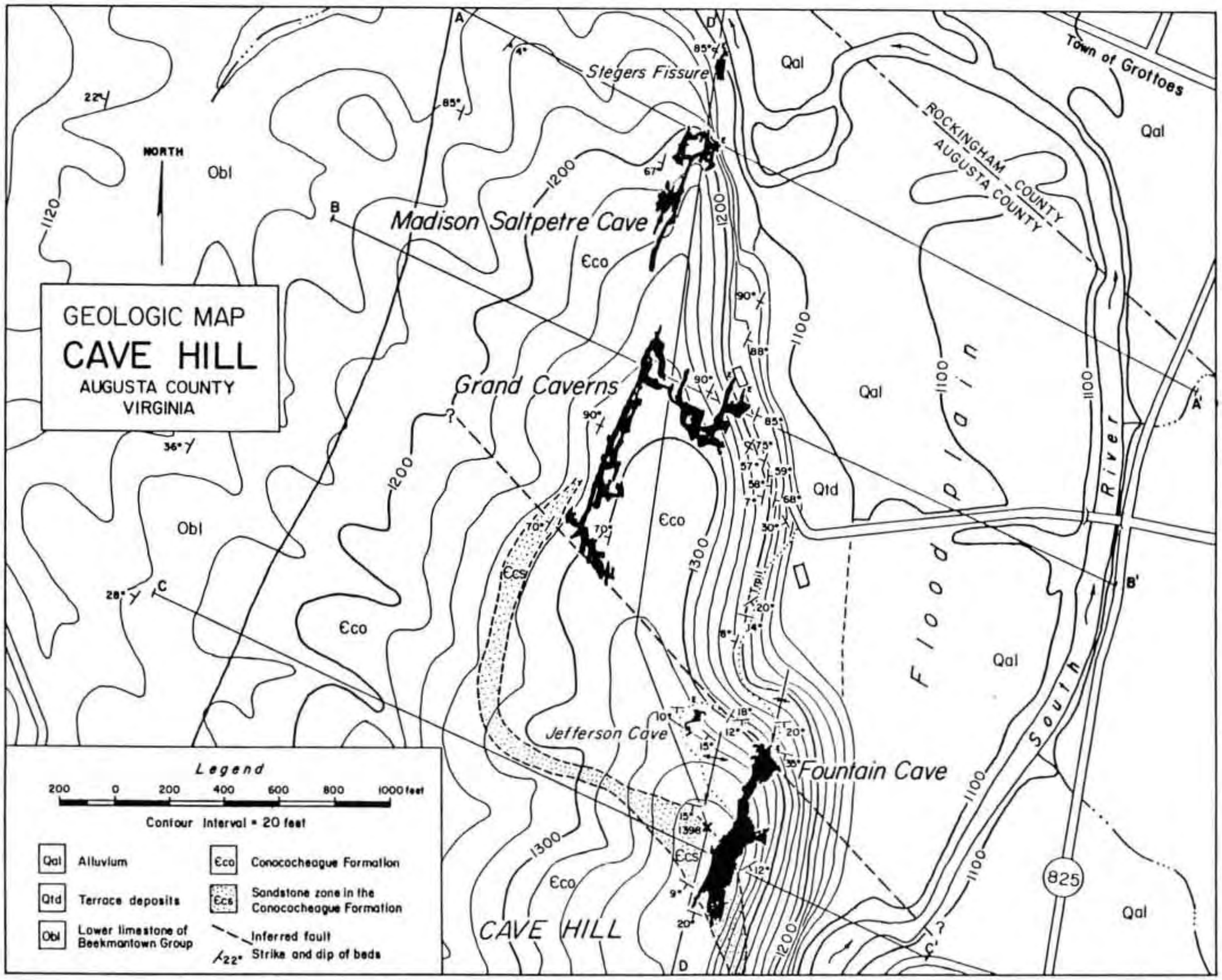

Figure 1: Geologic map of Cave Hill showing caves, bedrock, and structure. Cross-sections are shown in Figures 4 and 9.

Three hypotheses are developed in this study: (1) Beds of the Conococheague Limestone that are purest in calcite and microcrystalline are the least porous and those that are high in insoluble minerals have generally higher porosity, (2) Distinct changes in the structure of the bedrock from one place on Cave Hill to another are responsible for differences in the morphology of cave passages at these locations, and (3) Caves have formed at elevations that correspond to former levels of the streams in the area (namely the South River, a tributary to the Shenandoah River). As streams eroded and lowered the surrounding land surface, water tables dropped, and lower and lower cave levels developed in sequence.

The caves are developed in the Conococheague Formation that is easily recognizable throughout the mountainous region of western Virginia because it consists of a dense crystalline gray-to-black limestone that contains numerous thin wavy beds of light brown-to-tan clay and sand. Sedimentary rocks in western Virginia have been severely folded during the Paleozoic Era when continental plate collisions formed the Appalachian Mountain range. The beds of Cave Hill exhibit considerable structural deformation.

Methods

Laboratory Analysis of Rock Samples

Samples collected from Grand Caverns were analyzed in a laboratory to determine the approximate specific gravity, porosity, and permeability of various types of beds. The following procedure was followed in acquiring data: (1) weigh the samples while completely dry using an analytical balance, (2) soak the samples in a beaker of water for about ten minutes to allow the sample to take in all the water possible, (3) weigh the wet samples, (4) determine volumes of the samples by measuring the volume of water displaced as the sample is submersed, (5) weigh soaked samples while suspended in a beaker, that is kept above the pan of the analytical balance, and (6) calculate

porosity and specific gravity. Porosities and specific gravities of the samples are listed in Table 1.

Mapping

It was necessary to determine the precise locations of important features on Cave Hill so that they could be placed on maps of the surface or of caves. This was done with accurate "compass-and-tape" surveying. The survey progresses from station to station and the data is then processed by computer and plotted.

Table 1

Porosity and Specific Gravity

Sample Number	Percent Content Calcite*	Percent Porosity	Specific Gravity
1	89.8	3.3	2.71
2	0.0	27.1	2.14
3	32.6	2.2	2.62
5A	9.9	72.0	1.83
5B	6.25	13.0	2.40
6	6.4	28.4	2.19
7	76.3	1.5	2.71
8	55.3	1.6	2.68
9	64.6	5.2	2.57
10A	1.1	21.1	2.22
10B	2.85	28.6	2.21
11	0.0	68.0	1.81
12	0.0	40.4	1.70
13	89.6	7.0	2.71
14	10.9	25.9	2.17
15	0.4	47.4	1.73
16	88.1	3.8	2.62
17	92.2	28.4	2.39
18	5.9	33.1	2.34
19	73.5	3.3	2.67
20	76.05	2.9	2.76
21	59.2	4.4	2.70
22	36.4	5.2	2.72
23	0.0	5.5	1.99
24	0.0	20.9	2.24
25	68.5	4.8	2.67
26	61.5	6.4	2.60
27A	86.1	5.1	2.65
27B	3.8	38.9	2.05
28	0.0	6.5	2.43

*From Table 5, Kastning, 1990a.

It was also important to precisely determine the elevations of the caves in Cave Hill and of many of the surface features that are in the vicinity. This was accomplished using the U-tube method described by Palmer (1970). The U-tube is a length of transparent plastic tubing that is nearly filled with water. Two stadia rods are fashioned from wood on which a datum can be marked. When the tube is open to the atmosphere, water in the tube will be at the same level (elevation) at each end of the tube. During surveying, a datum is maintained and is moved up or down with rising or lowering terrain. In this way precise elevations of features can be determined.

Sedimentary rock strata that began as horizontal deposits are steeply inclined at Cave Hill. Some of the beds in the caves are even vertical. The orientations, or attitudes, of beds of rock were measured as strikes and dips (Suppe, 1985; Marshak and Mitra, 1988). This was done with a Brunton Compass.

During this study, forty-six strike-and-dip measurements were made on Cave Hill and in the caves. Strike-and-dip symbols were plotted on the geologic map (Figure 1) as it was constructed.

Once a number of measurements of attitude of beds were taken, it was possible to statistically plot these on a stereographic equal-area (Schmidt) net (Figure 2). This graphical representation clearly shows clustering of data (beds that have nearly identical attitudes) and can be used to work out many geometrical solutions to folds, such as the attitude of a fold axis or the plunge of folds (*i.e.* how the entire fold may slope into the ground). These analyses were conducted using standard stereo-net techniques (Marshak and Mitra, 1988).

It is obvious from looking at maps of caves on Cave Hill (Figure 1) that passages typically appear to be composed of straight segments (lineaments). Many of these are parallel to one another; others may be perpendicular to these. This suggests that these passages are controlled by the geometry of the strata or by fractures in the rocks.

Lineaments may be plotted on polar graph paper to indicate how these directions cluster. Statistical plots, known as rose diagrams, were constructed for the cave-passage data using the method of Smith (1968). Rose diagrams for eight caves on Cave Hill and the nearby South River are shown in Figure 3.

One of the main objectives of this study was to compile a geologic map of Cave Hill in the vicinity of the largest caves. Because a map shows only a two dimensional view of the area, it was also desirable to construct cross sections through the hill that would show the subsurface geology and caves in relation to each other. Both the map and cross sections together provide a three-dimensional perspective of the geology of Cave Hill.

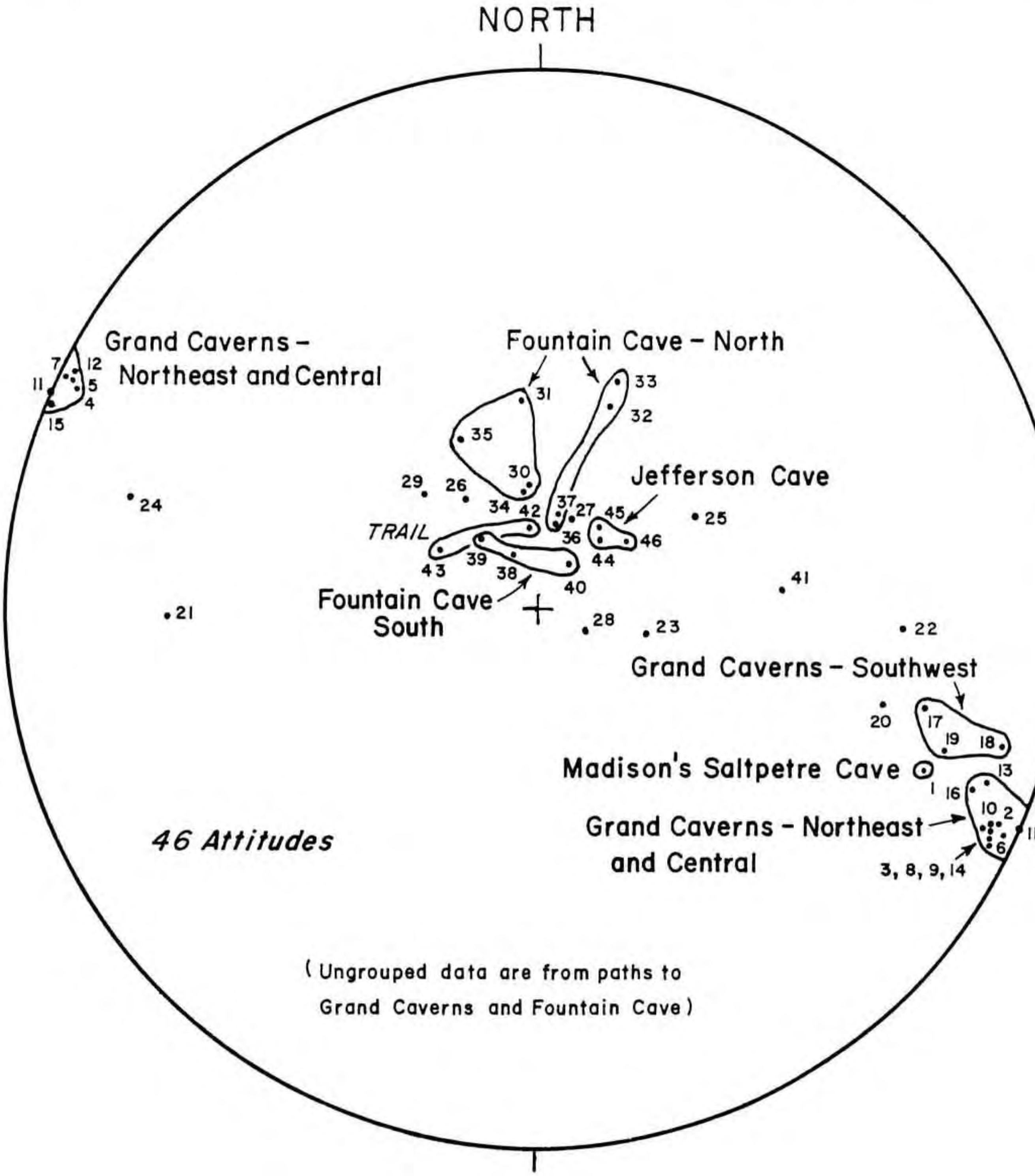

Figure 2: Schmidt equal-area net showing poles of dipping beds. Note differences in orientation of beds in caves north of the fault in Figure 1 and those south of the fault.

The geologic map (Figure 1) and four cross sections (Figures 4 and 9) were constructed using a photographically enlarged topographic map, existing geologic maps (Gathright, Henika, and Sullivan, 1978a,b), horizontal-survey data, leveling-survey data, structural field data (Figure 2), and cave maps. The geologic cross sections of Figure 4 were constructed roughly perpendicular to the trend of Cave Hill and crossing through key geologic features including major caves. These sections show the South River and its flood plain. Cross sections were

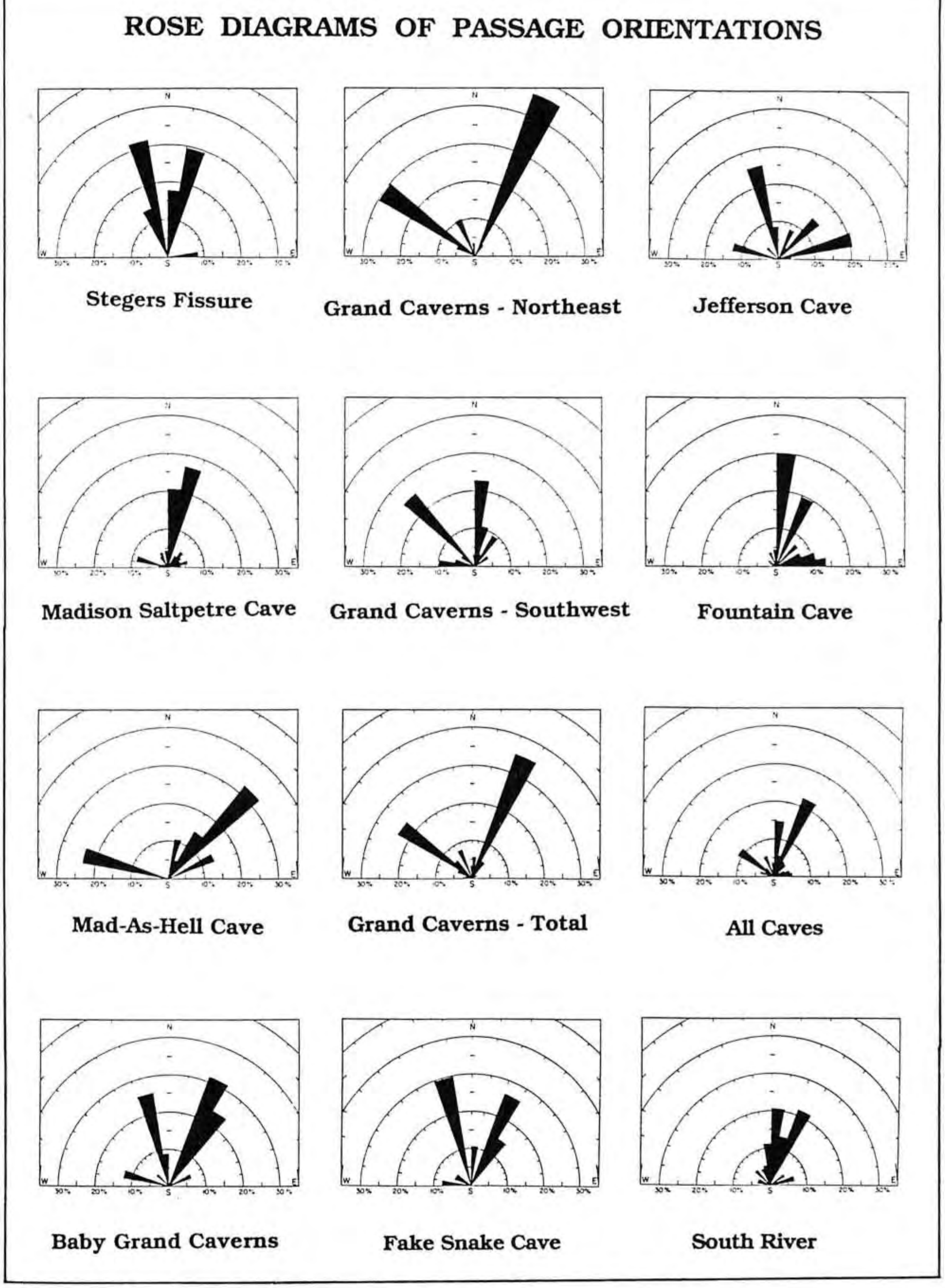

Figure 3: Rose diagrams for eight caves in Cave Hill and the South River.

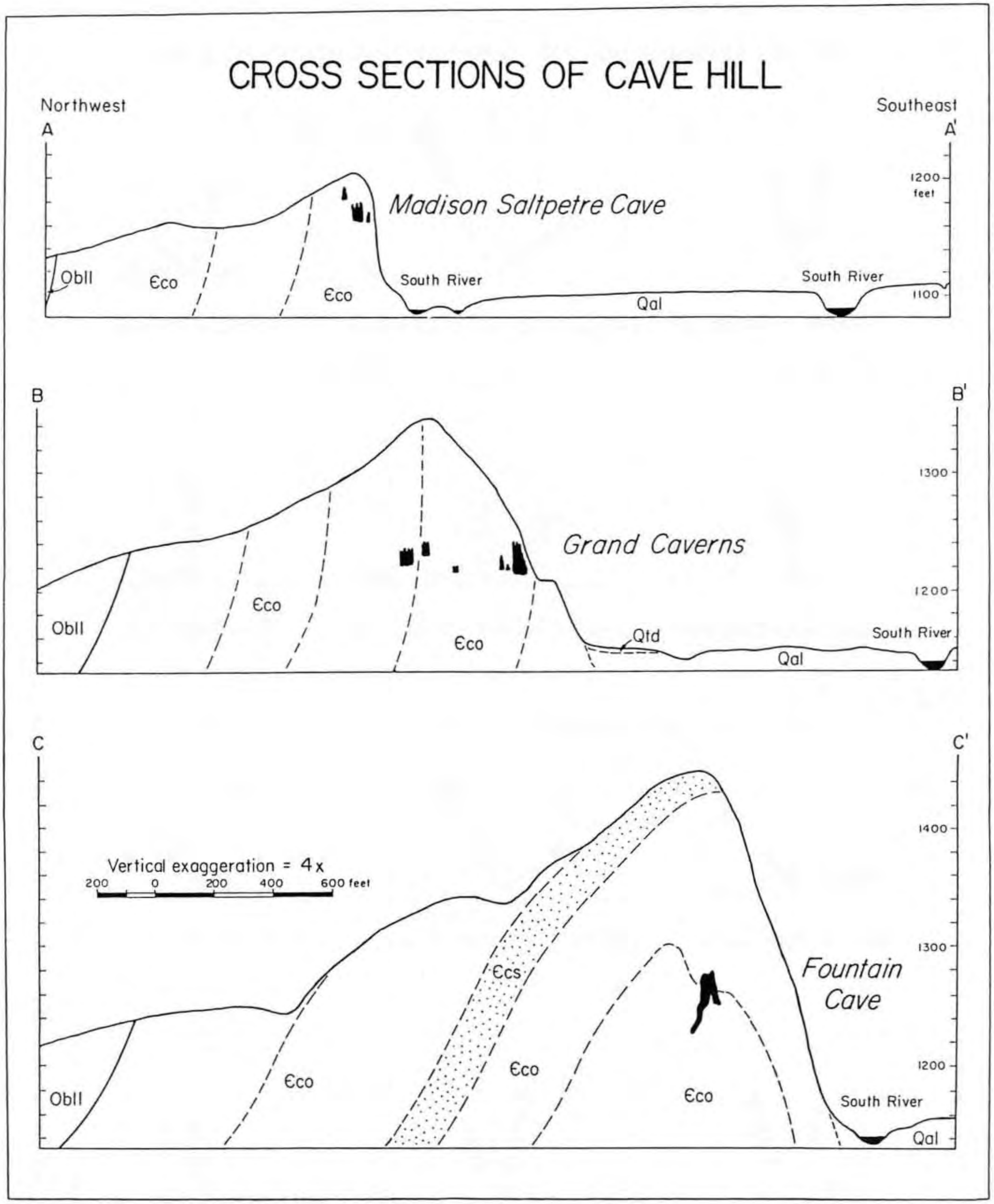

Figure 4: East-west cross sections through Cave Hill showing attitude of rocks and vertical position of caves. *See* Figure 1 for locations of cross sections.

drawn with an expanded vertical scale in order to best illustrate vertical relationships, including slopes. The map and cross sections put the topographic features, caves, and bedrock together in one package so that all of these things can be compared and relationships determined.

Measurements of bedding attitude were used to construct beds at their correct angles on cross sections. Using data from the U-tube leveling surveys, elevations of major caves were placed in their correct vertical position (Figure 4). This was done to show how caves relate to the surrounding beds, bedrock structure, topographic features, and to each other.

Results

Characteristics of Rocks

Porosity and specific gravity for samples collected in Grand Caverns are listed in Table 1. The percent calcite compositions determined in an earlier study (Kastning, 1990a,b) are given in the second column of Table 1. Porosity and specific gravity were each plotted against composition of the beds (expressed as percent calcite) in order to determine any relationship between these variables (Figure 5).

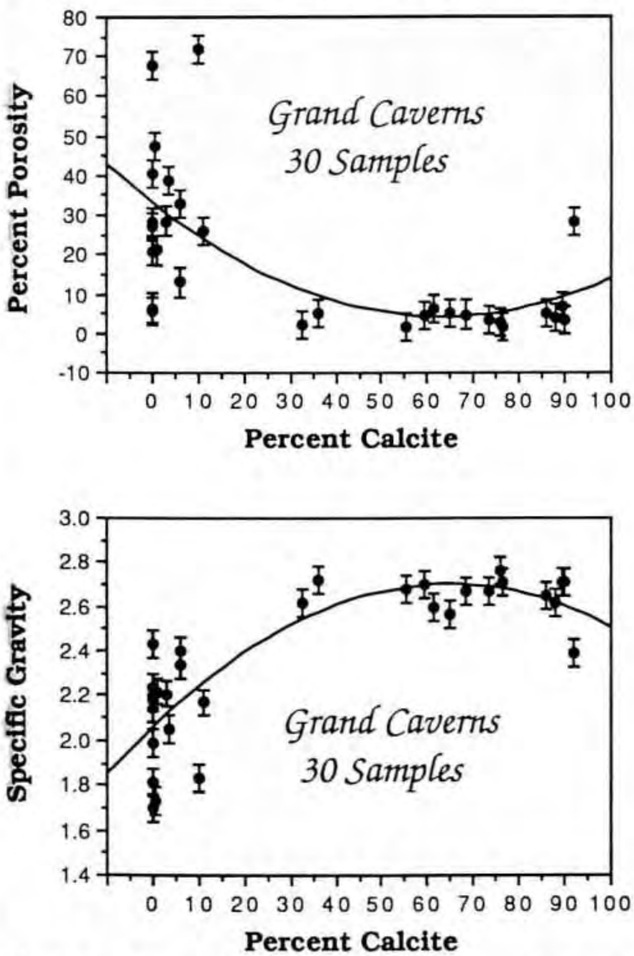

Figure 5: Graphs of porosity and specific gravity versus percent calcite for 30 samples from Grand Caverns.

The graph of porosity versus limestone purity shows distinct differences in porosity for samples of vastly different composition. There is a strong correlation between porosity and calcite content. It is easily seen from Figure 5 that high-purity limestone samples (30 percent and higher) have very low porosity and that samples with a high level of insoluble material show considerable porosity. In fact, porosity is generally less than 10 percent for relatively pure limestone and greater than 20 percent for rocks high in insoluble content.

In an earlier study, Kastning (1990a,b) determined that high-purity limestone was dense and microcrystalline. The data of Figure 5 supports the hypothesis that dense materials would have considerably fewer pores and will thereby hold less water.

The graph of specific gravity (relative density) versus limestone purity also indicates a distinct relationship. Rock samples of high-calcite content have a high specific gravity. The values of specific gravity for these samples lies very close to the value of 2.71, the known specific gravity for calcite (Klein and Hurlbut, 1985). On the other hand, the specific gravity for samples of mostly insoluble material is significantly lower, around 2.0 to 2.4.

The significance of these determinations is that the most soluble materials (highest purity limestone with least insoluble material) are the least porous and that the least soluble materials (lowest purity limestone with most insoluble material) are the most porous.

Structural Control of Caves

In general, the rose diagrams of Figure 3 show that orientations of passages in each cave agree well with the known structure in the immediate vicinity of the cave. This is particularly true for two of the large caves, Madison's Saltpetre Cave and Grand Caverns. In Grand Caverns there is a difference between average passage orientation in the southwest end of the cave and that in the main part of the cave. In this case the rose diagrams agree well with the strike-and-dip readings taken in the cave or just outside the entrances (*compare* Figures 1 and 3).

The notable exception to good agreement between the rose diagram and structural data is Fountain Cave (Figures 1 and 3). The structural data indicates that the strikes of rocks taken in the cave lie between N77°W and N55°W and between N50°E and N86°E. But, the rose diagram shows that most of the cave is oriented between true north and N30°E. Instead of following the strike, Fountain Cave is roughly following the plunge of a small syncline. The rose diagrams for smaller caves and for all of the caves combined (Figure 3) show a trend that is consistent with the overall structural trend of Cave Hill (Figure 1).

The combined rose diagram for all caves was constructed by weighing each cave according to its length. Consequently, the trends of the three larger caves dominate this diagram. Figure 3 also gives a rose diagram for linear segments of the South River in the vicinity of the caves. The trend of the river is generally parallel to the overall structural trend of the area. This is as expected since the river lies between mountain ranges that follow the regional structural pattern.

The 46 strike-and-dip measurements are plotted on an equal area (Schmidt) net (Figure 2). Poles clustering at or near the circumference of the net represent vertical beds, such as those in most of Grand Caverns (poles number 2-16 on Figure 2). Poles 17-19 are the attitudes of rocks in the southern part of Grand Caverns. These have a slightly different position and represent a different structural trend. Beds in Jefferson Cave (poles 44-46) are close to being horizontal (the poles are near the center of the Schmidt net) but they have about a 13° to 17° dip to the southwest. Beds in Fountain Cave (poles 30 through 41) dip between 10 to 30 degrees to the south-southwest or south-southeast.

Discussion and Conclusions

An Explanation for Speleogenesis in Mixed Lithologies

In some limestones, the porous rock is not necessarily the one that is the most soluble. Crystalline limestone, although high in calcite content, may be impermeable to water flow. In the Conococheague Limestone of Cave Hill, numerous thin beds of insoluble material, generally sand and clay, easily soak up water and transmit it. However, where acidic water comes in contact with limestone, the rock will dissolve, and the insoluble material will not.

An explanation for cave development is proposed here that is generally different from other explanations normally found in books and articles on cave origin. This is a direct result of the studies performed at Cave Hill and in the laboratory with samples that came from Grand Caverns.

Because water has a difficult time getting through dense microcrystalline rock, it will be unable to travel within the limestone and dissolve openings into the rock. However, if adjacent beds are highly porous and permeable, they can bring water to the limestone (Figure 6, step 1). The water must flow through openings to get to the rock and to leave the rock carrying dissolved material.

As water flows through a bed of porous rock, it comes in contact with the adjacent limestone (Figure 6, step 1). An opening begins to form next to the insoluble (but permeable) rock. The opening is in the limestone (Figure 6, step 2). When the opening becomes large enough, flow within it will increase, allowing still larger flows to occur. The flows become rapid enough to efficiently excavate a cave (Figure 6, step 3). The insoluble beds are left behind, perhaps jutting from the walls or ceilings until they break off (Figure 7).

Structural Controls

Cave passages north of an inferred fault, namely Steger's Fissure, Madison's Saltpetre Cave, and Grand

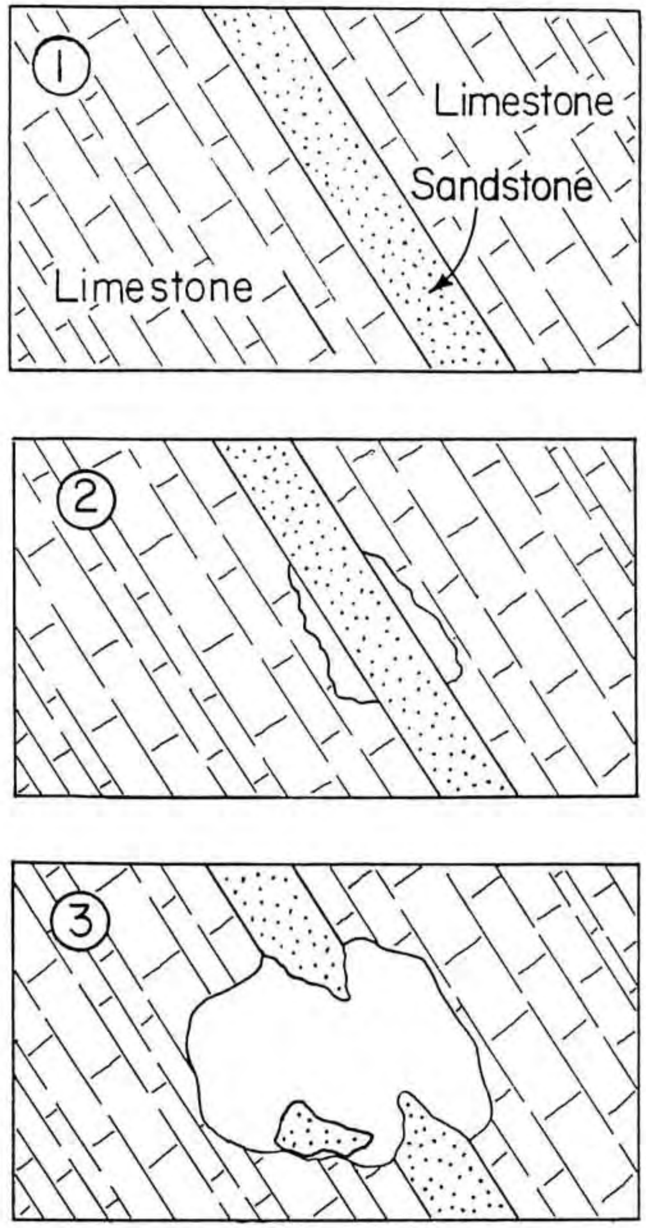

Figure 6: Sequence of development of cave passages in interbedded porous sandstone and non-porous, microcrystalline limestone. Initially, groundwater flows only in the sandstone (step 1). Once adjacent beds of limestone begin to dissolve (step 2) both flow and solutional excavation accelerate within the limestone (step 3).

Figure 7: Beds in the ceiling of a large room in Grand Caverns. Note blades of insoluble sandstone and clay (dark colored) protrude into the passage, whereas beds of microcrystalline limestone (light colored) have been dissolved and are recessed. Width of view is about 3 feet.

Caverns (Figure 1), have developed along strike within nearly vertical beds that are high in calcite content and very soluble. In contrast, cave passages south of the inferred fault, namely Jefferson Cave and Fountain Cave, are not strike oriented. Passages in Jefferson Cave and parts of Fountain Cave are aligned along fractures. Additionally, other passages in Fountain Cave trend parallel to the axis of a small plunging syncline (Figures 1 and 4). Statistical support for these relationships are graphically shown in rose diagrams (Figure 3) and on a Schmidt equal-area net (Figure 2).

Passages in the caves may be classified as following the strike, oriented along fractures, or positioned in other ways (such as following the dip). Distribution of passage lengths within each cave with respect to these three categories are shown in the ternary diagram of Figure 8. This plot concisely summarizes the differences in structure and its control on speleogenesis for caves on either side of the fault.

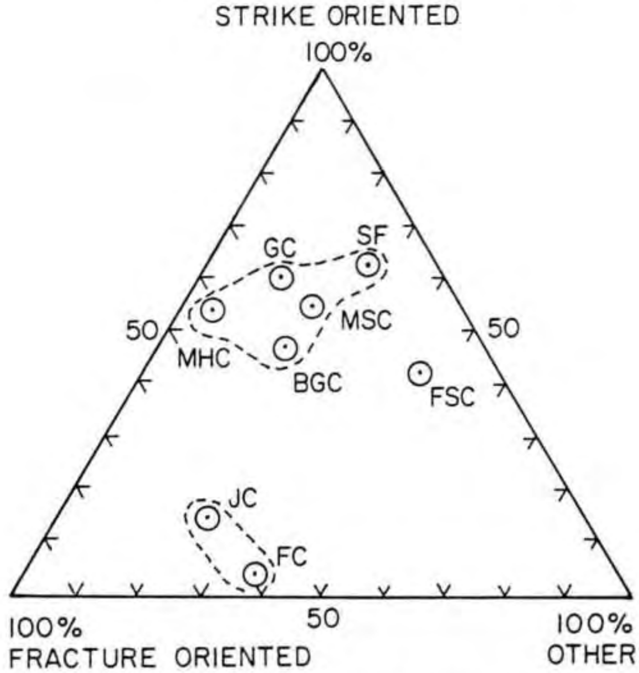

Figure 8: Ternary graph showing structural controls on cave development in Cave Hill. Caves north of the inferred fault are Steger's Fissure (SF), Madison's Saltpetre Cave (MSC), Baby Grand Caverns (BGC), Mad-As-Hell Cave (MHC), and Grand Caverns (GC). Those south of the fault are Jefferson Cave (JC) and Fountain Cave (FC). Fake Snake Cave (FSC) is a small cave on the western flank of Cave Hill.

Topographic Relations

Cave levels in Cave Hill are lower toward the north. The older passages, toward the south, are well drained and flow now occurs in the flooded lowest levels (Figure 9). The geologic cross sections suggest a correlation between elevations of the drained caves and former floodplains, marked by river terraces on the edges of the modern-day floodplain (Figures 4 and 9).

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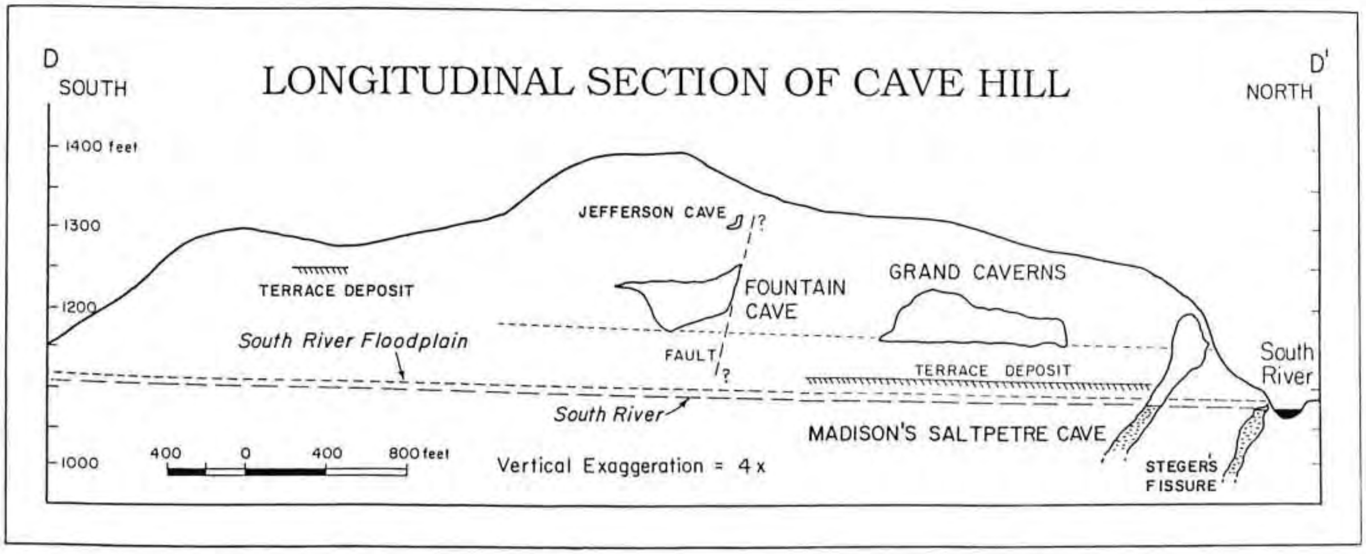

Figure 9: Longitudinal section of Cave Hill showing major caves, topographic profile, South River, and mapped terrace and floodplain deposits. *See* Figure 1 for location of the section.

Morton, Mr. Tom Spina, Mr. Phil Lucas, and Dr. John Holsinger for providing information on caves located on Cave Hill.

References

Ford, D.C. and Williams, P., 1989, *Karst Geomorphology and Hydrology*: Unwin Hyman, Winchester, Massachusetts, 320 p.

Gathright II, T.M.; Henika, W.S.; and Sullivan III, J.L., 1978a, Geology of the Crimora quadrangle, Virginia: *Virginia Division of Mineral Resources Publication 13* text and 1:24,000-scale map.

Gathright II, T.M.; Henika, W.S.; and Sullivan III, J.L., 1978b, Geology of the Grottoes quadrangle, Virginia: *Virginia Division of Mineral Resources Publication 10,* text and 1:24,000-scale map.

Kastning, Kass, 1990a, *Speleogenesis in the Appalachian Foldbelt: Phase I: Effects of Limestone Composition on Cave Formation:* Science Fair Project and Report, Radford High School, Radford, Virginia, 88 p.

Kastning, Kass, 1990b, Effects of limestone composition on cave formation (abstract): *Abstracts, 41st International Science and Engineering Fair, Tulsa Oklahoma,* Science Service, Washington, D.C., p. 125.

Klein, C. and Hurlbut, C.S., Jr., 1985, *Manual of Mineralogy (After James D. Dana),* 20th edition: John Wiley and Sons, New York, 596 p.

Marshak, S. and Mitra, G., 1988, *Basic Methods of Structural Geology*: Prentice Hall, Englewood Cliffs, New Jersey, 446 p.

Palmer, A.N., 1970, Some recent innovations in cave surveying: *National Speleological Society News,* v. 28, no. 4, p. 44-46.

Palmer, A.N., 1990, Groundwater processes in karst terranes, *in* Higgins, C.G. and Coates, D.R. (editors), Groundwater geomorphology: The role of subsurface water in Earth-surface processes and landforms: *Geological Society of America Special Paper 252,* p. 177-209.

Palmer, A.N., 1991, Origin and morphology of limestone caves: *Geological Society of America Bulletin,* v. 103, no. 1, p. 1-21.

Smith, A.R., 1968, A rose is a rose is not a flower: *The Texas Caver*, v. 13, no. 10, p. 113-115.

Suppe, J., 1985, *Principles of Structural Geology*: Prentice Hall, Englewood Cliffs, New Jersey, 537 p.

White, W.B., 1988, *Geomorphology and Hydrology of Karst Terrains,* Oxford University Press, New York, 464 p.

Concept and Classification of Cave Breakdown: An Analysis of Patterns of Collapse in Friars Hole Cave System, West Virginia

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ABSTRACT

Cave breakdown consists of locally derived, predominantly bedrock fragments. Features of breakdown include individual fragments, their accumulations, and the morphologic modifications they impose upon passages. The surfaces of breakdown range from dissolutional surfaces to fracture surfaces as endmember forms. Individual breakdown is classified into fracture forms, dissolutional forms, and mixed forms, according to surface type. Breakdown is described by loose shape terms such as curved sheet, flake, beam, wedge, blade, chip, slab, and block. Breakdown with a predominant origin is classified by genesis; examples include exfoliation and crystal-wedging fragments. Breakdown with a distinctive combination of shape and origin is classified morphogenetically; examples include pendant cluster fragments and canyon trench blocks. Breakdown is classified according to its mode of occurrence within a collapse continuum. The continuum describes breakdown within the context of the modifications that fragmentation and particle accumulation impose upon passage morphology. The continuum is arranged in order of roughly increasing size or complexity. The collapse continuum includes *in situ* fragments, isolated collapse fragments, local feature collapse, passage-junction collapse, major passage collapse, and large chamber collapse. Features of the continuum in Friars Hole Cave System include paleo-mud-crack fragments, exfoliation fragments, gypsum-crystal wedging fragments, cross-bed blades, fault wedges, pendant clusters, floodwater-maze spans and blades, pothole fragments, flute fragments, shaft debris, trench blocks, disintegrating blocks, junction-room trench blocks, breakout domes and associated debris, terminal breakdowns, and large chambers.

Introduction

Many accounts of breakdown in limestone caves (Davies, 1949, 1951; White and White, 1969; Sweeting, 1972; Jennings, 1971, 1985; Bögli, 1980; Ford, 1988; White, 1988) identify collapse as the predominant, or at least proximate (Ford and Williams, 1989) origin of breakdown. Often, breakdown is equated with jagged or angular bedrock fragments that have fractured surfaces. Such a conception, however, can easily misdescribe breakdown. Breakdown need not be angular or have a significant percentage of area as fractured surface at the time of (or immediately following) fragmentation or collapse. Breakdown need not later retain more than a few percent of surface area as fractured surfaces and still be considered breakdown. Moreover, breakdown need not originate by collapse: much material considered breakdown, such as gypsum-crystal wedging and exfoliation fragments, has fragmented *in situ*, and then been displaced only a few centimeters, slowly by sliding and rotating, even upward against the force of gravity (Figure 1). Such considerations suggest that a conception of breakdown as collapsed angular fragments is inadequate: it is simply too narrow to characterize the bedrock fragments that appear in limestone caves and both scientists and explorers habitually identify as breakdown.

This paper therefore offers a reinterpretation of the concept of cavern breakdown. The interpretation is based on study of Friars Hole Cave System, West Virginia, and other eastern caves. It is supported by observations of breakdown in caves in Texas, South Dakota, Montana, Minnesota, and Mexico. Not suprisingly, altered concepts of breakdown force an altered perspective on breakdown classification. Thus, an additional purpose is to offer a new classification of cave breakdown and collapse features.

Figure 1: Lifted gypsum-crystal wedging fragments. Framboidal pyrite in argillaceous micrites and dolomicrites is altered, leaving hematitic pseudomorphs. Released aqueous sulfur in the form of sulfate reacts with calcite to form gypsum. The Highway, Rubber Chicken Cave.

Definition of Breakdown

Breakdown is produced by the fragmentation and local displacement of bedrock from cave walls, ceilings or floors. Breakdown is also produced by local fragmentation of pre-existing breakdown. Individual breakdown consists almost exclusively of bedrock fragments. Minor amounts of breakdown consists of bedrock with speleothem encrustations or clastic sediments that became cemented before or even after fragmentation. Chert nodules, fossil fragments, quartz silt and sand, or other material weathers from bedrock, but these are best considered weathering detritus (White and White, 1969). Although speleothems collapse, it is not useful to designate their fragments as breakdown. In caves with paleokarst fills, or caves intersecting intrastratal karst (Palmer and Palmer, 1989), bedrock fragments may appear in breccias of varying age and origin. It may then be difficult to distinguish between fragments formed within the present cave and fragments formed in earlier caves, or in intrastratal karst zones preceeding cave development. However, cave breakdown can usefully be characterized as locally derived, predominantly bedrock fragments that owe their origin to processes operating in caves.

Processes of Breakdown

A number of processes affect the form, size, and shape of breakdown during its residence time within a cave. Too many studies lose sight of the fact that the processes operate before, during, and after initial fragmentation. Processes contributing to the formation of breakdown fall into at least three groups: (1) dissolution, (2) chemical growth and alteration, and (3) mechanical overloading. The processes have been discussed by Davies (1949, 1951), White and White (1969), Renault (1968), White (1988), Jennings (1971, 1985), Powell (1977), Jagnow (1978, 1979), Bögli (1980), and Jameson (1983, 1985). The following paragraphs emphasize processes that operate before or during fragmentation, and identify a few processes that do not appear in the literature.

Dissolutional processes include (1) enlargement of fractures to liberate *in situ* fragments, (2) enlargement of fractures to weaken bedrock by diminishing the surface area of cohesive bonds across fractures, (3) undercutting of bedrock by free surface streams, (4) mining or coring by drips, films, sprays, or waterfalls, to reduce the area of attachment of speleogens such as flutes, blades, and projections on shaft walls, or to cut through bedrock spans, blades, and ledges, and (5) dissolution and transport of cements or other soluble material out of porous beds, to produce weakened, often friable bedrock.

Processes of chemical growth and alteration include (1) weathering of clay minerals by leaching of ions and uptake of water to produce swelling, mechanical weakening of bonds, and opening of cracks and partings and (2) crystal wedging by growth of ice or salts, including gypsum and halite.

Processes directly triggering mechanical overloading include: (1) growth of the conduit to a critical unsupported size, under the prevailing overburden stresses, residual tectonic stresses, and bedrock characteristics, to cause mechanical rupturing and often collapse, (2) expansion toward a free surface to produce stress-release rupturing and fragmentation as a type of exfoliation, (3) loss of buoyant support as phreatic conduits are drained or floodwaters rise and fall, (4) loss of support by removal of clastic sediments, (5) impact as collapsing bedrock shatters, or preexisting fragments are swept over vertical shafts and then impact, and (6) hydraulic toppling (the breakage of coherent bedrock or tipping of *in situ* fragments, by water flowing against weakly attached bedrock, or against bedrock with a large surface area but a small area of attachment).

Source of Bedrock Fragments

Bedrock fragments in caves may have a wholly internal (endogenetic) or external (exogenetic) origin. **Endogenetic** fragments are produced in caves, or at surface/subsurface intersections such as cave entrances, as a result of locally operating processes. **Exogenetic** fragments originate outside caves, or originate as fragments derived from solution, chemical alteration, and collapse, in intrastratal karst preceeding cave development. Exogenetic fragments are transported by fluvial processes to depositional sites in caves. Subsidence, collapse, or subglacial processes are also possible transport mechanisms. Common exogenetic fragments include surface weathering debris and fragments of surface karren. In alpine caves, glacial or periglacial debris forms exogenetic fragments. Also included are pinnacle and cutter fragments produced at the soil/bedrock interface or deeper within the subcutaneous zone. In caves of the eastern United States, pinnacle and cutter fragments are found near entrances, in terminal

breakdowns, and in debris associated with shafts and dolines.

Recognition of fragment provenance as endogenetic or exogenetic, and the identification of fragments according to genesis, is not merely academic. It is important in the evaluation of breakdown features and other deposits, such as paleontological and archeological deposits, in which bedrock fragments are included.

Transport of Breakdown

In situ fragmentation of bedrock produces trapped fragments that may resist relocation over long periods of time. Many fragments originate *in situ*, but rotate or slide as a result of foreign crystal growth (or by fluvial removal of underlying support) to become breakdown. However, most breakdown has been transported farther; the fragments are displaced a few meters to tens of meters from their points of origin, accumulating as **local deposits**. The upper limit on the distance breakdown can move and still be identified as breakdown is poorly defined. It depends on size and shape of the fragments, processes of transport, and gradient of the cave passage. Most importantly, it depends on the extent to which fragments are modified and incorporated into fluvial or other deposits.

An example that tests the limits of distinction between breakdown and fluvial sediments, and the concept of a local origin for breakdown, is the following. In high relief karsts such as those of Huautla, Mexico, where caves have long sequences of vertical shafts separated by short, horizontal passages, bedrock fragments detach in large number from shaft walls, particularly in thinly bedded units of high dip. The fragments fall down shafts, or slide down steep chimneys, often aided by stream action, especially during floods. Many of the fragments probably are detached during the floods. Shaft floors and ledges exhibit extensive piles of angular debris, or dissolutionally modified angular debris. It is not possible to distinguish the distance of transport of the debris. Much of it likely is transported down a number of shafts, often to distances of three or four hundred meters vertically below the location of original fragmentation and a few hundred meters horizontally. Even assuming that the debris has fragmented several times by shattering during the movement, and given that fluvial action is involved in transport, it is not clear how to interpret the origin of the sediment. Is it a fluvial sediment? Is it breakdown? I favor the latter interpretation, and suspect most cavers exploring these systems would too.

Collapse

Collapse is the relatively rapid downslope movement of fragments under the influence of gravity. Collapse movements include falling, sliding, rotating, and bouncing. Collapse can be into open cavities filled with air. Or collapse may involve settling in water, assuming that much breakdown forms when passages are first drained of water and buoyant support is lost (White and White, 1969). Collapse most often results as massive failure of bedrock, in which rock strength is exceeded and initial fragmentation is followed by displacement and additional fragmentation (for example, by shattering as fragments impact against one another, thus producing **shattering fragments**). However, collapse need not involve fragmentation. Collapse can occur when pre-existing fragments are undermined by the erosional removal of support provided by underlying fluvial sediments.

Breakdown Features

Breakdown features include individual fragments, their accumulations, and the morphologic modifications they impose upon passages. Where breakdown features are distinctive and common, it is useful to describe and name them. Thus, **breakdown types** are defined on the basis of (1) predominant modes of origin (exfoliation, crystal wedging fragments) or (2) the characteristics and origin of individual fragments (pendant clusters, spongework debris, bedrock spans from floodwater mazes). **Breakdown associations** are defined on the basis of: (1) the characteristics of breakdown accumulations (entrance talus cones, breakout-dome talus), (2) the characteristics of modified passage morphology (breakout domes, rock stumps remaining from collapse of exfoliation fragments), (3) the passage context (terminal breakdowns, passage-junction breakdown), or (4) some combination of the above.

Individual Breakdown: Previous Classifications

Davies (1949) classified breakdown on a morphogenetic basis. He distinguished ceiling and wall blocks, ceiling and wall slabs, scaling plates, and scaling chips. A restatement of the classification (Davies, 1951) emphasized breakdown as collapse debris for the purpose of mathematical analysis of collapse mechanics, but retained the descriptive terms block, slab, plate, and chip for individual forms.

White and White (1969) modified Davies' classification, explicitly for the description of breakdown, to recognize blocks, slabs, and chips, based on the number of included beds. Division into block, slab, and chip forms was presumed to offer a strict morphologic basis for classification, it being judged that a morphogenetic classification was difficult to apply. White and White (1969) and later White (1988) nonetheless recognized that the classes allowed some leeway in actual shape, because (1) fragments are rarely end-member forms, and (2) shapes are somewhat variable, particularly for chips. They also recognized a scale dependence based upon bed thickness, because a given size fragment could be a block or slab depending on the thickness of beds. Despite these difficulties, White and White (1969) considered the classification generally useful because it was judged to allow classification without speculation on origin.

The White and White (1969) classification has been widely but uncritically used. It adequately distinguishes blocks and slabs *if* beds are homogeneous depositional units bounded by prominent, subparallel bedding-plane partings (no cross beds), and if much fragmentation follows prominent bed partings and bed-perpendicular joints. These conditions are locally common in the eastern United States, particularly in Kentucky and Tennessee, but even in these areas, bedding can be lithologically variable and structures complex. In complex structural settings (with faults, inclined joints, stylolites, fault breccias) more characteristic of the folded Appalachians, or in cases of breakdown produced by exfoliation (Jameson, 1983) and gypsum-crystal wedging, the classification falters. Attempts to rigorously apply it lead to serious difficulties.

The problems begin with the failure to operationally define the term bed. A bed can be either (1) a depositional unit or (2) a unit bounded by field-recognizable bedding-plane partings. Recognition of bed partings depends on their prominence, which itself depends on chemical weathering of the beds and on the presence of associated dissolutional features. If definition (1) is used, then one has to look carefully at fragments to make the distinction, which may not be easy. Textural variations within beds, surface irregularities of the fragment, weathering rinds, and clay or other clastic-sediment coatings easily complicate the analysis. Also, it is necessary to define depositional unit for this context, which literally no one has done. If definition (2) is used, the task is somewhat easier, because surface irregularities and coatings do not so easily hide partings. There still may be disagreement on parting recognition, particularly for thinly bedded or laminated limestones with easily weathered clay aligned along laminae. Such layers often weather in a distinct fashion (so that they readily form tight but recognizible partings) but fail to have associated dissolution features, or promote significant fragmentation. Usage in the literature appears to favor definition (2), which appears most useful, but that usage is hardly explicit.

The choice of definition makes considerable difference, and highlights one of the five difficulties of the block/slab/chip classification. The first difficulty is that *the classification fails to unambiguously classify fragments.* For example, fragments classified blocks by one definition are chips by the other. Figure 2 shows a curved exfoliation sheet. If a bed is a depositional unit, then the fragment is a block. If a bed is a unit bounded by prominent bed partings, then the fragment is a chip.

Even if one settles on the definition of a bed as a unit bounded by prominent bed partings, the classification fails to *usefully* classify fragments. The curved exfoliation sheet of Figure 2 is not usefully described as a block or a chip, since there is at least one better description. Figure 1 shows lifted fragments that contain recognizable partings: Is it useful to classify these small fragments as blocks? Most workers would call these chips, but there

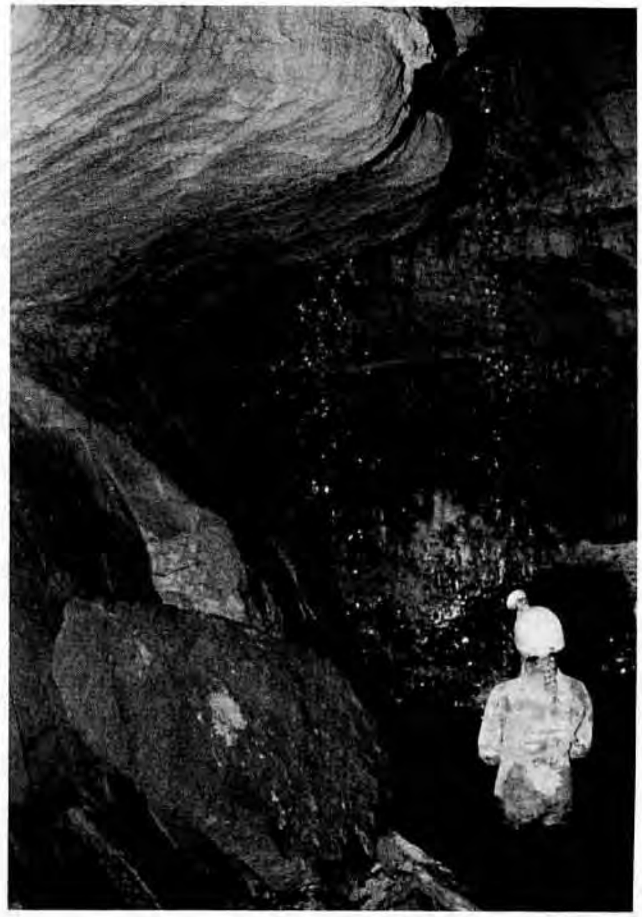

Figure 2: Curved exfoliation sheet. The contact in background separates overlying pure biosparites from argillaceous micrites that have exfoliation. Lithologic variations within the argillaceous unit suggest that more than one bed is present, but prominent bed partings are absent.

may be yet better descriptors.

A third difficulty is that *the classification misdescribes actual shapes.* The block/slab/chip classification is not really a shape classification, as is apparent from the White and White (1969) statement of it. Shapes are often variable, and do not closely approach end-member shapes that one might want to ascribe to the terms block and slab. However, if common shapes directly contradict the denotations or connotations of the terms, then it is legitimate to criticize the classification for misdescribing breakdown. Figure 3 shows breakdown that could be described as slabs and as wedges. The slabs would be classified by White and White (1969) as blocks, slabs, or chips; the wedges would be classified as blocks or chips, depending, as usual, on the choice of bed definition.

A fourth difficulty, recognized in part by White and White (1969) is the *problem of scale.* They minimize it, but the problem is serious. Far too many small fragments must be considered blocks. Far too many large fragments, weighing tens to hundreds of thousands of tons, must be

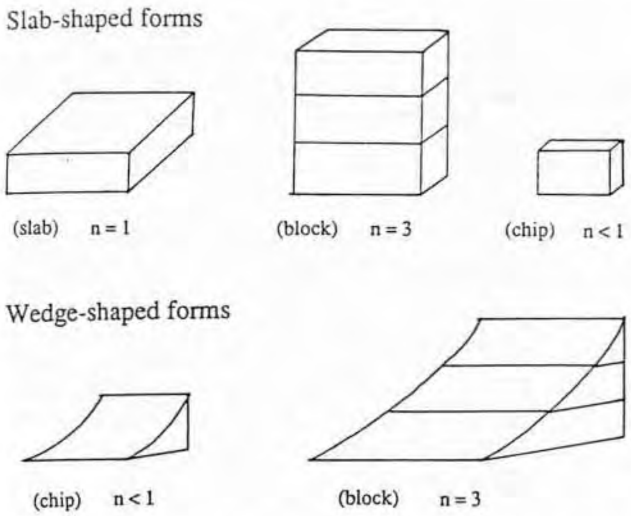

Figure 3: Slabs and wedges. n = number of beds. Top: slab-shaped forms have identical shape, but can be blocks (n = 3), slabs (n = 1), or chips (n < 1) with the White and White (1969) classification. Bottom: wedge-shaped forms, common in fault zones, have identical shape, but can be blocks (n = 3) or chips (n < 1). The geometry of wedges precludes their classification as slabs, by definition, unless the left and right ends were the bed partings.

considered chips, because they are in relatively homogeneous bedrock without prominent bed partings. One wonders whether the often cited breakdown block of Carlsbad Caverns, Iceberg Rock, estimated to weigh over 200,000 tons (Bullington,1968), is in fact a chip rather than a block.

A final problem arises for those cases in which speleogens such as pendants (Figure 4), bedrock blades, vertical shaft flutes, or pothole walls have collapsed, or fragmented in place and have been locally moved. These bedrock features are common, but are simply ignored by previous classifications of breakdown.

Individual Breakdown: Alternative Classifications

Possible Morphologic Classifications

A descriptive classification needs to less ambiguously classify breakdown, retain essential shape or size data, and be useful within the cave. It is dubious that a single, purely descriptive classification that is useful for all breakdown, today can be constructed: Breakdown develops in too many ways and exhibits too many poorly studied forms. Careful description and measurement of smaller breakdown of certain types (exfoliation, crystal-wedging fragments) may be feasible and lead to rigorous descriptive classifications. Such classifications may employ shape terms based on dimensions of long axes, such as that developed by Zingg (1935) for pebbles. Alternately, descriptive classifications might better define endmember geometric shapes as blocks, slabs, beams, rods, wedges, flakes, sheets, arrowheads, or the like. These could then be compared with breakdown and individual fragments could be more objectively classified. The author, in fact, has begun this process for the analysis of exfoliation fragments, and expects thereby to gain some insight into the mechanics of exfoliation jointing on walls and at bends of passages.

Figure 4: Pendant breakdown. Near the Otter Slide, Friars Hole Cave.

Morphogenetic Classification

This study will instead develop a morphogenetic classification based in part upon the characteristics of fragment surfaces. In doing so, it is convenient to retain the loosely descriptive terms block, slab, and chip. The terms are allowed only minimal shape connotations and are not defined on the basis of the number of included beds. The terms are supplemented by other loosely descriptive terms (sheet, flake, blade, wedge, arrowhead-shaped, etc.) as needed. The shape terms are combined with genetic or locational modifiers that help identify a particular morphogenetic class of breakdown (fault wedge, exfoliation sheet, cupola span, floodwater-maze span).

Classification of Breakdown Surfaces

Initially, the surfaces of breakdown consist of fracture surfaces or dissolutional surfaces. In analogy with the crystal faces of the mineralogist, it is sometimes useful to describe the surfaces as **fracture faces** or **dissolutional faces**. Over time, fracture faces are dissolutionally modified, and have dissolutional forms such as scallops superimposed upon them. In many caves, weathering rinds up to a few centimeters thick also modify breakdown and obscure fragmentational origin. The degree of modification is a function of time, the lithology of the fragment, and the hydrologic setting. In some settings breakdown lasts up to millions of years with little modification; in others, breakdown surfaces may be modified significantly within decades. Collapses that choke passages and promote floodwater-maze development favor rapid dissolutional modification of breakdown. Argillaceous units with

high clay contents, such as the illite- and kaolinite-bearing micrites and dolomicrites of the Pickaway Limestone in Friars Hole Cave System, develop a thick weathering rind that modifies fracture surfaces beyond recognition and makes breakdown extremely slippery.

Breakdown Forms

Breakdown is classified into dissolutional forms, fracture forms, and mixed forms, according to the predominant surface character of the fragment. **Dissolutional forms** have surfaces formed primarily by dissolution. Examples include speleogen fragments such as pendants and pendant clusters, spongework spans, and pothole wall fragments.

Fracture forms have surfaces formed by fractures. These may be tectonic/sedimentation fractures (bed partings, joints, faults, stylolite surfaces) or fractures formed during cave development. The latter type includes induced tension or shear fractures (new fractures formed by fragmentation across cohesive rock, caused by mechanical overloading), shattering fractures, crystal-wedging fractures, exfoliation fractures, and fractures associated with chemical weathering of bedrock, particularly of clay minerals to promote minor surface flaking.

Mixed forms have surfaces formed by relatively unmodified fracture surfaces in addition to dissolutional surfaces. There is a gradation from fracture forms to mixed forms and dissolutional forms; fracture and dissolutional forms are thus endmembers of a continuum. The continuum between dissolutional and fracture forms is further complicated by the fact that, over a sufficiently long time scale, fracture forms should become dissolutional forms as topographic surface lowering destroys caves.

Collapse Continuum

Breakdown features are usefully classified within the context of a collapse continuum (Figure 5), arranged in rough order of increasing size and complexity of individual fragments, their accumulations, or the modifications they impose upon pre-existing passage morphologies. The continuum includes *in situ* fragments, isolated collapse fragments, local feature collapse, junction collapse, major passage collapse, and large chamber collapse. In part, the collapse continuum can be thought of as the **mode of occurrence** of breakdown features. This interpretation works well for *in situ* fragments, isolated collapse fragments, and local feature collapse. The interpretation seems less satisfactory for junction collapse, major passage collapse, and large chamber collapse, because of a tendancy to think of these as breakdown features in their own right. However, it is more appropriate to consider them as classes of breakdown features. The classes subsume particular features and breakdown forms (depending on lithology, structure, and dissolutional patterns), many of which are as yet only poorly defined or` documented.

In Situ Fragments

At the lower end of the spectrum are ***in situ* fragments**. These are common where closely spaced fissures and partings intersect, forming three-dimensional solids whose surfaces can be solutionally widened sufficiently to loosen bedrock. Such fragments are often trapped on walls or ceilings until undercutting or lateral enlargement by vadose streams is sufficient to allow their collapse or displacement by stream transport. *In situ* fragments are not collapse features *per se*, but over time they will be moved and become indistinguishable from breakdown proper, or be modified and incorporated into other deposits.

Isolated Collapse Fragments

Where fragments collapse individually and fail to form multiple accumulations, it is appropriate to speak of **isolated collapse fragments.** Isolated collapse fragments need not, but often do, appear as distinctive forms.

Local Feature Collapse

Local feature collapse is distinguished for sites with a multiple accumulation of fragments that form a distinctive, well-characterizable accumulation. It is also defined for sites with a modification of existing passage morphology brought about by multiple fragmentation. In many cases, only fractured walls and ceilings testify to former collapse, due to fluvial removal of debris. Local feature collapse may occur as relatively isolated features, or as separate accumulations scattered over extensive lengths of passage, often merging into one another. Small breakout domes and associated debris can appear as local features, such as in Mystery Cave, Minnesota , and in many South Wales caves (Davies, 1977). Circular cross sections and associated exfoliation features in Indiana caves are local breakdown features (Powell, 1977). Ewers (1969) describes local feature rooms produced by lateral enlargement and collapse in Kentucky caves.

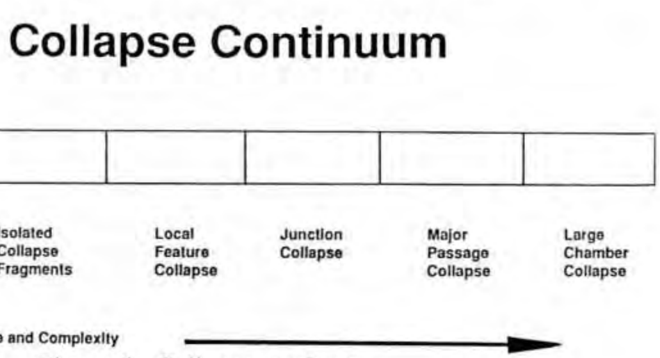

Figure 5: Collapse continuum.

Junction Collapse

In many caves, junctions are the preferred sites for collapse to form rooms. The rooms attain widths several to many times greater than mean passage widths away from the rooms. The mechanics of room collapse are well known (White and White, 1969). Details of junction-collapse features are not as well documented. However, passage junctions do exhibit distinctive patterns of collapse and fragment accumulation. Collapse patterns are a function of lithology, structure, hydrologic setting, passage type, and geometry and location of the intersection relative to other passage features. An example from Friars Hole Cave System is described in a later section.

Major Passage Collapse

Collapse along long segments of passage is characteristic of passages that form in zones of densely fractured bedrock, or that underlie valleys or other locations, such as sinkholes, where aggresive waters can be diverted underground. In many cases, major passage collapse terminates passages, forming terminal breakdowns beneath isolated dolines or, more commonly, valley walls (Brucker, 1966).

Large Chamber Collapse

Large chamber collapses typically occur at passage junctions, in zones of extensive fracturing, and in locations where passages have developed at several levels and one crosses beneath another. Large chamber collapses are also associated with near proximity to the surface. Fish (1977) describes large-chamber-collapse features for some of the deep Mexican surface pits. Many of these are huge phreatic chambers intersected by surface lowering. In some, conduits to lower levels apparently provided loci for removal of the collapse debris, by subsidence and dissolutional removal where streams were available. Large chamber collapse has been well documented for anthropogenically-induced collapse by draining of chambers through groundwater mining (Beck, 1984).

Breakdown in Friars Hole Cave System

The classification scheme is selectively illustrated below for Friars Hole Cave System. This branchwork cave exhibits the following morphologies: tubes, canyons, fissures, shafts, and large collapse chambers. Most of the surveyed passage length (63 km) is in canyons. Additional structural data and stratigraphic sections are in Jameson (1985). Worthington (1984) gives a broad account of the development of the system; Jameson (1985) analyzes the growth of the North Canyon of Snedegar's Cave, and describes some breakdown in more detail.

Classification By Surface Form

Fracture Forms: Fracture faces are formed by bed partings, N 60-75 E set joints, inclined joints, thrust faults, fault-subparallel joints, exfoliation joints, and fractures induced by gypsum-crystal wedging or shattering. Polygonal joints bound paleo-mud-crack fragments in some argillaceous units of the Union and Pickaway Formations. In Monster Cavern, the Crows Nest Room, and other large chambers estimated by Worthington (1984) to account for about 30% of the known volume of the system, most of the fragments are true collapse debris. The debris has induced tension and shear fractures except where breakage is along bed partings, faults, or shattering fractures. Fault zones contain an abundance of fragments. The most distinctive are wedge shaped, with faces formed by joints, a fault, and a bed parting. Inclined extension joints appear in many fault zones, or in crests of overlying anticlines. Much of the associated debris is slab-shaped, with inclined joint-fracture faces (Figure 6).

Dissolutional Forms: The most common forms are pendant clusters. Blades modified from N 60-75 E set *en echelon* joint-bounded bedrock are typical of floodwater settings. Cross-bed blades are locally abundant in oolitic cross-bedded units of the Union Limestone, but if dissolutional modification is minimal, the blades are more properly classified as fracture forms. Flutes, blades, and irregularly shaped (but typically sharp) bedrock fragments may be found detached on walls of shafts or as debris below shafts, either as debris piles or isolated fragments littering shallow bedrock pans filled with water. Pothole wall fragments and undercut ledges, often containing chert lenses, are found in many canyons in the Pickaway Limestone.

Mixed Forms: The most common distinctive mixed forms are canyon-trench blocks and fault wedges. Figures 7 and 8 illustrate a single type for each form. Variations

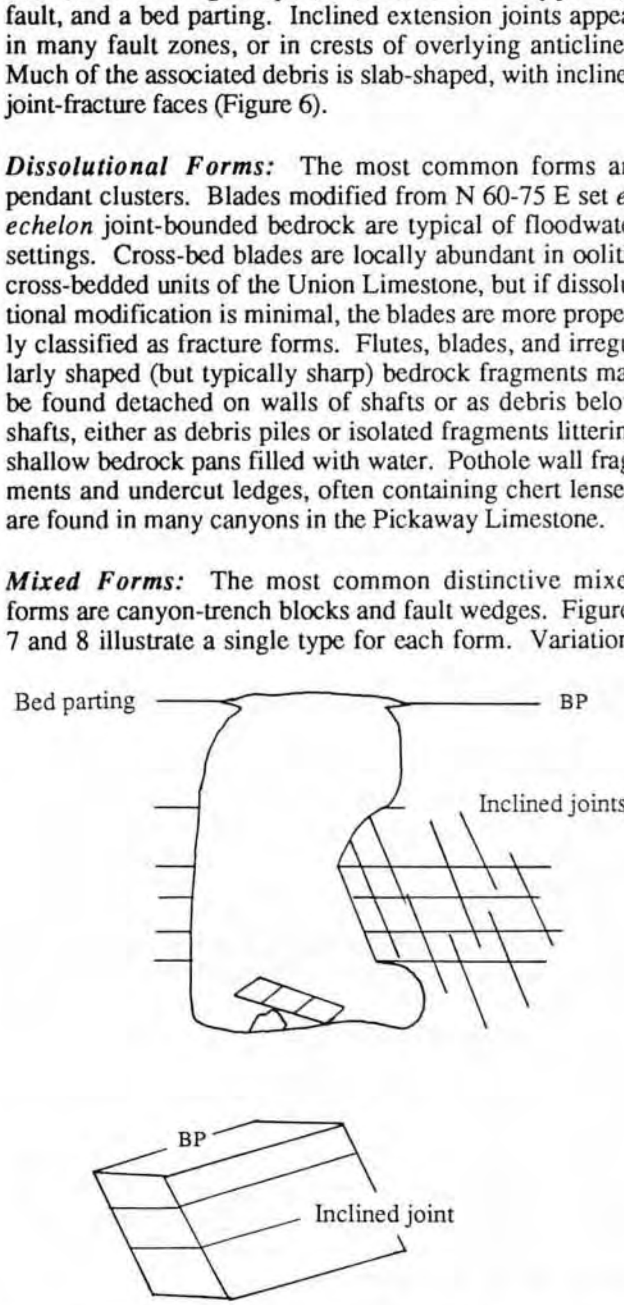

Figure 6: Origin of inclined joint slabs. A fracture form is shown, but mixed forms also develop.

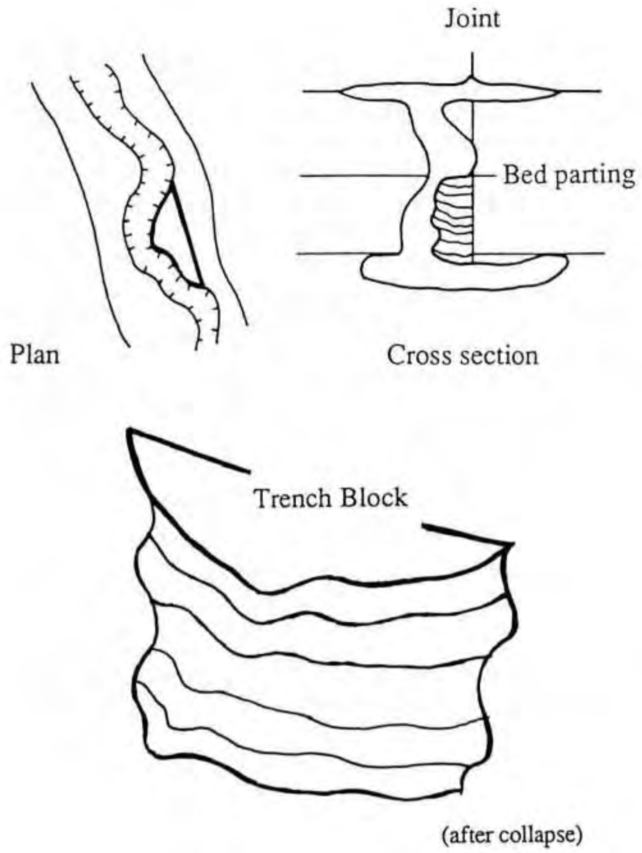

Figure 7: Origin of trench blocks in canyons. The block has surfaces formed by a bed parting (top); a trench wall and a joint (sides); and an undercut (bottom).

replace one or more fragment faces with undercut surfaces, bed partings, N 60-75 E set joints, or inclined joints.

Classification by Predominant Origin

Exfoliation: The exfoliation (Figure 2) occurs as chips, flakes, plates, curved sheets, and wedges, ranging from cm-sized to 10 m-sized fragments, with many about 1 m long and high and 10-30 cm thick, but tapering toward ends. The fragments have surfaces curved convexly out into the cave, imparting a tendancy toward hyperbolic parts of cross sections of canyons, rounded passage bends, and a variety of other features described by Jameson (1983). This type of exfoliation is confined to argillaceous units and is most abundant in canyons in the Pickaway Formation.

Classification by the Collapse Continuum

In Situ Fragments: The most common are gypsum-crystal wedging and exfoliation fragments. Others are: cross-bed blades, *en echelon* joint blades, inclined extension joint slabs, fault wedges, fault-subparallel joint fragments, and pendant clusters.

Isolated Collapse Fragments: These are found in many canyons and tubes that otherwise have seen little

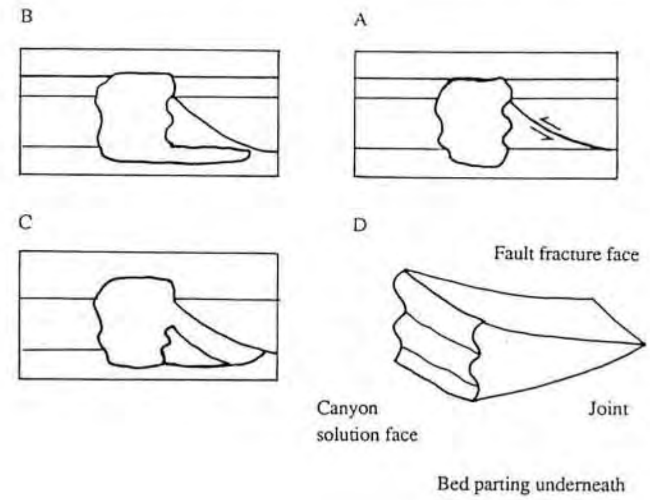

Figure 8: Origin of fault wedges in canyons. Cross sections: (A) Canyon with bed partings and fault. (B) Canyon after lateral undercutting. (C) Collapse of fault wedge. Fault wedge: (D).

fragmentational degradation. The fragments litter bedrock floors or floors covered with fluvial deposits. Canyon trench blocks or slabs are the most common distinctive forms. Blades, spans, and other speleogen breakdown in floodwater mazes nearly always occur as isolated fragments. Pendant clusters that have collapsed usually occur as isolated fragments.

Local Feature Collapse: Multiple accumulations of fragments to form local feature collapses occurs with nearly all of the fragment types just listed in the section on *in situ* fragments. In many cases, particularly with gypsum-crystal wedging and exfoliation fragments, *in situ* fragments and collapsed fragments occur together. Thus they can be described collectively as a local feature collapse forming a single accumulation over a given passage reach.

Two other local feature collapse types at Friars Hole were described by Jameson (1979) from Snedegar's Cave as disintegrating blocks and retreating walls. **Disintegrating blocks** separate along closely spaced fractures (Figure 9). **Retreating walls** are walls with closely-spaced fractures, *in situ* fragments, nearby collapse fragments, often of disintegrating blocks, and wall rock stumps (Figure 10). Retreating walls have a recessed or re-entrant geometry. Some retreating walls are developed in clay-rich homogeneous or laminated micrites and dolomicrites overlain by purer biosparites and oosparites. Thrust faults locally splay up into purer beds from contacts between these units. Jameson (1979) suggested that the fractures were associated with thrust faulting, but later (1985) noted that at many locations the fractured walls are formed in purer units with or without associated faults. Most fractured walls are at entrances or in near-surface passages within about 500 m of present or suspected former entrances. In the purer units, major fracture surfaces often parallel existing passage walls as nearly straight planes; this con-

tributes to an unusual tunnel-like cross section in some reaches of Snedegar Trunk. The major fractures also curve around passage bends. The geometry of the major fractures is unlike the convex-out into-the-passage geometry of the exfoliation fractures of the argillaceous units. These observations nonetheless suggest an unloading type of origin by exfoliation. However, the issue is further complicated by abundant minor fractures. These have elliptical traces in sections cut parallel to bedding. Many parting fragments from these zones have arrowhead and wedge shapes. Others can be described as slivers or fragments with shapes derived from these forms. The fragments are vaguely reminiscent of the lenticular masses of bedrock between normal faults in tensional structural regimes. It is at least plausible that minor fractures and parting fragments represent the response of these particular lithologies to an expansional regime in bedrock adjacent to the cave. The proximity of these features to cave entrances, and the presence of surficial periglacial features in this part of the Appalachians, suggests the possibility that their unusual prominence has been aided by frost wedging. Frost wedging is active each winter on nearby surface cliffs and in cave entrance rooms (*e.g.*, at Crookshank and Toothpick caves).

Junction Collapse: A distinctive example occurs at the junctions of canyons in the Union Limestone. Canyons are often tall and narrow, but widen locally by lateral enlargement. Lateral enlargement occurs where clastic sediments armour the floor, or an argillaceous unit impedes downcutting. In either case, solution is more efficient on lower walls, particularly at passage bends where flow from one wall is directed to the opposite wall. At junctions, the undermining may be directed at the same wall from opposite directions, one from each passage. Large canyon-trench blocks form at these locations. Typical fragment faces are (1) an upper bed parting, (2) a lower undercut surface, and (3) two walls consisting of solutional forms characteristic of entrenchment (cusps, ledges, and undercut

Figure 9: Disintegrating blocks. Nearby walls are heavily fractured. Blocks collapse from walls and disintegrate along fractures. Fragmentation is today aided by summer condensation corrosion. Trunk passage of Snedegar's Cave.

Figure 10: Retreating wall. Surface above the recessed wall has quartz crystals set on calcite-crystal fibre slickensides along a thrust fault. The fault is parallel to bedding at the contact between the lower argillaceous unit (recessed wall with spalls) and the purer unit above. Quartz Room of Snedegar's Cave.

surfaces; *see* Jameson, 1985). On the downstream ends of junctions, trench blocks also form; they usually have at least one fracture face aligned on N 60-75 E set joints.

Conclusion

Breakdown consists of locally derived, predominantly bedrock fragments. Breakdown can be described and classified in a number of ways:

Fragment surface (or face): fracture surface, dissolutional surface

Fragment type: dissolutional forms, fracture forms, mixed forms

Fragment shape: flake, sheet, chip, block, slab, wedge, beam, arrowhead-shape, etc.

Predominant mode of origin: exfoliation, gypsum-crystal wedging fragment,. etc.

Morphogenetic class: trench block, fault wedge, pendant cluster, etc.

By mode of occurrence in a collapse continuum: In situ fragments, isolated collapse fragments, local feature collapse, junction collapse, major passage collapse, large chamber collapse

Breakdown fragments range over at least four orders of magnitude in length, twelve orders in volume, and ten orders in mass. The large range in size and mass, combined with an abundance of large fragments, effectively limits the kinds of quantitative studies to those that are feasible within the cave with a minimum of displacive sampling. Obviously, such standard techniques as sieving are impractical. For this reason, among others, the study of cave breakdown has remained largely qualitative, with the exception of the theoretical analysis of collapse mechanics (Davies, 1951; White and White, 1969).

This study is no exception to the trend; the preceeding is based on qualitative observations. To advance the analysis of breakdown, it is necessary to undertake more detailed, quantitative analyses of breakdown form and genesis. The accurate assessment of breakdown characteristics (size and shape) may be a difficult task, but for certain types of breakdown - such as that of exfoliation and crystal wedging - it may hold the key to a better understanding of fragmentational origin, as well as understanding of the mechanical stress history surrounding cave passages.

References

Beck, B.F. (editor), 1984, *Sinkholes: Their Geology, Engineering, and Environmental Impact: Proceedings of the First Multidisciplinary Conference on Sinkholes, Orlando, Florida, 15-17 October 1984*: Rotterdam and A.A. Balkema, Boston, 429 p.

Bögli, A., 1980, *Karst Hydrology and Physical Speleology*, Springer-Verlag, Berlin and New York, 284 p.

Bullington, N.R., 1968, Geology of the Carlsbad Caverns, *in Delaware Basin Exploration*: West Texas Geological Society, Guidebook 68-55, p 20-23.

Brucker, R., 1966, Truncated cave passages and terminal breakdown in the central Kentucky karst, *National Speleological Society Bulletin*, v. 28, no. 4, p. 171-178.

Davies, C.W., 1977, Breakout domes in South Wales caves, *in* Ford, T.D. (editor), *Proceedings of the Seventh International Congress of Speleology, Sheffield, England, September, 1977*: British Cave Research Association, Somerset, England, p. 136-139.

Davies, W.E., 1949, Features of cavern breakdown: *National Speleological Society Bulletin*, v. 11, p. 34-35.

Davies, W.E., 1951, Mechanics of cavern breakdown: *National Speleological Society Bulletin*, v. 13, p. 36-43.

Ewers, R.O., 1969, *A Model for the Development of Subsurface Drainage Routes Along Bedding Planes*, M.S. thesis (unpublished), University of Cincinnati, 84 p.

Fish, J.E., 1977, *Karst Hydrogeology and Geomorphology of the Sierra de el Abra and the Valles-San Luis Potosi Region, Mexico*, PhD. dissertation (unpublished), McMaster University, Hamilton, Ontario. 469 p..

Ford, D.C., 1988, Characteristics of dissolutional systems in carbonate rocks, *in* James, N. and Choquette, P. (editors), *Paleokarst*, Springer-Verlag, New York, p. 25-57.

Ford, D.C. and Williams, P.W., 1989, *Karst Geomorphology and Hydrology*, Unwin Hyman, London, Boston, Sydney, and Wellington, 601 p.

Jagnow, D.H., 1978, Geology and speleogenesis of Ogle Cave: *National Speleological Society Bulletin*, v. 40, no. 1, p. 7-16.

Jagnow, D.H., 1979, *Cavern Development in the Guadalupe Mountains*: Cave Research Foundation, 55 p.

Jameson, R.A., 1979, Breakdown in fault zones: Snedegar's Cave, Friar's Hole Cave System, West Virginia (abstract): *Program of the National Speleological Society Annual Convention, Pittsfield*: National Speleological Society, Huntsville, p. 27.

Jameson, R.A., 1983, Exfoliation in Friars Hole Cave system, West Virginia, *in* Dougherty, P.H. (editor), *Environmental Karst*, GeoSpeleo Publications, Cincinnati, p. 119-129.

Jameson, R.A., 1985, *Structural Segments and the Analysis of Flow Paths in the North Canyon of Snedegars Cave, Friars Hole Cave System, West Virginia*, M.S. thesis (unpublished), West Virginia University, Morgantown, 421 p.

Jennings, J.N., 1971, *Karst*: The MIT Press, Cambridge, Massachussets, 252 p.

Jennings, J.N., 1985, *Karst Geomorphology*, Basil Blackwell Ltd, Oxford and New York, 293 p.

Palmer, M.V. and Palmer, A.N., 1989, Paleokarst of the United States, *in* Bosak, P.; Ford, D.C.; and Horacek, I., (editors), *Paleokarst: A Systematic and Regional Review*, Elsevier and Academia, Amsterdam and Prague, p. 337-363..

Powell, R., 1977, Lateral unloading of isotropic rock as a process of solution channel enlargement, *in* Tolson, J.S. and Doyle, F.L. (editors), *Karst Hydrogeology*, International Association of Hydrogeologists Memoir 12, p. 433-441.

Renault, P., 1968, Contribution a l'etude des actions mechaniques et sedimentologiques dans le speleogenese, duxieme partie, Les effets mechaniques a l'echelle de la cavite: *Annales de Speleologie*, v. 23, p. 259-307.

Sweeting, M.M., 1972, *Karst Landforms*, Macmillan, London, 362 p.

White, E.L. and White, W.B., 1969, Processes of cavern breakdown: *National Speleological Society Bulletin*, v. 22, no. 1, p. 43-53.

White, W., 1988, *Geomorphology and Hydrology of Karst Terrains*, Oxford University Press, New York, 464 p.

Worthington, S.R.H., 1984, *The Paleodrainage of an Appalachian Fluviokarst: Friars Hole, West Virginia*: M.S. thesis (unpublished), McMaster University, 218 p.

Zingg, T., 1935, Beitraege zur Schotteranalyse: *Schweizerische Mineralogische und Petrographische Mitteilungen*, v. 15, p. 39-140.

Mud Flow in a Karst Setting

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ABSTRACT

Mud banks are present in most caves, and therefore are appropriate for investigation as part of a typical karst environment. In 1987, a peculiar mud bank in Shine Cave (Alabama No. 210) caught the attention of the authors. Rather than drip pits, the mud has linear drip channels, with each channel terminated up-slope under an actively dripping stalactite. The mud is also impressed with prominent down-slope striations caused by irregularities at a roof offset against which the bank is squeezed as it enters the cave room. The source of the mud is Fault Sink which lies above and behind the bank, although no open passage connects the sink on the surface and the cave. The slope of the bank is 30 to 40 degrees. Its general appearance is that of a "mud glacier". The authors reported a qualitative assessment of the site at the 1988 Friends of the Karst meeting, but quantitative data were lacking then. Therefore, in November 1987 and additionally in February 1988 and in February 1989, we placed groups of markers on the slope. A marker consists of a plumb-bob strung from the cave roof and a vertical wire inserted into the mud directly below the bob. As the mud moves, the bob tip and the wire separate. Measurements of these separations on 28 dates now span more than three years. Analyses of the motions give average rates of a few millimeters per year. A correlation with rainfall was particularly dramatic after a record flood in December 1990, when the motion was 4 mm between measurements separated by 43 days. With regard to rainfall dependence, the phenomenon resembles surface mud flows.

Introduction

Shine Cave, (Alabama Cave Survey No. 210) is a short, isolated section of an abandoned trunk passage that can be traced through much of the four-mile length of Newsome Sinks (Jones and Varnedoe, 1980; Varnedoe and Lundquist, 1986). At both ends, Shine Cave is terminated by collapsed segments of the trunk (Figure 1). These collapsed sites form Fault Sink to the south and Shine Sink to the North. The cave entrance is through Shine Sink. Just north of Shine Sink, another intact trunk section is found in Chapel Cave, (Alabama No. 208).

The southern end of Shine Cave is a typical mud and rubble slope, below and to the northwest of Fault Sink. Figure 2 illustrates a profile of the slope along the line A-A in Figure 1. Most of the slope appears chaotic, but one area, 10 m horizontally and 4 m downslope is a rather uniform mud bank with a slope of 30° to 40° from horizontal. This is identified as "study area" on Figures 1 and 2. On its upper side, the slope reaches the roof, where the mud seems to have been squeezed under a vertical roof offset as the mud enters the room. Irregularities in the roof at that point impress striations in the mud surface that continue down slope (Figure 3). Its general appearance is that of a mud glacier. This situation was brought to the attention of the authors by Michael Martin during a trip on September 5, 1987.

Further examination of the slope found parallel, linear drip channels rather than drip pits, under each active stalactite (Figure 4). At its up-slope end, each channel begins directly under an active drip. Qualitative evidence said that the channels were caused by the mud moving slowly under the drip (Varnedoe and Lundquist, 1988).

Quantitative Measurements

The rate of mud motion down slope was an obvious next question to investigate. To obtain quantitative data, a series of plumb bobs and surface markers were installed: two on November 14, 1987, one on February 27, 1988

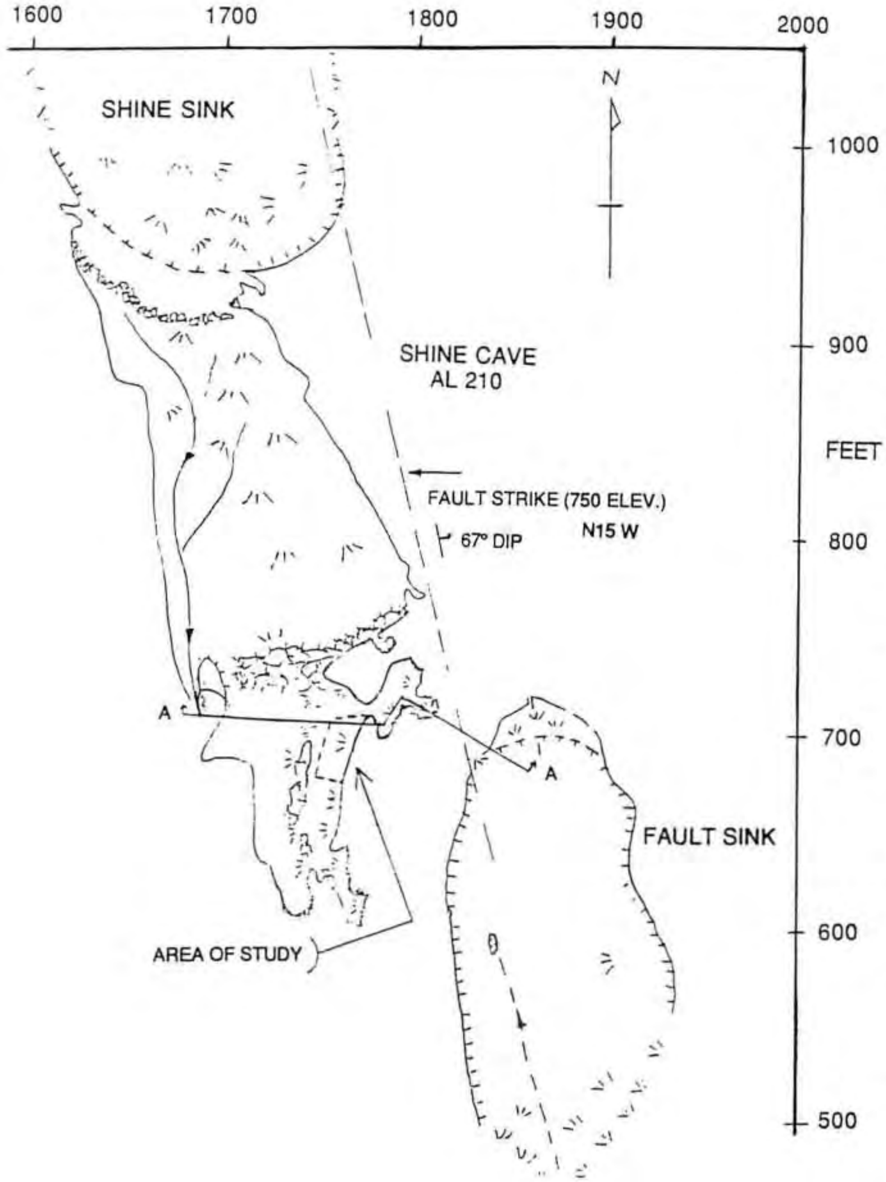

Figure 1: Shine Cave lies between Shine Sink and Fault Sink. Coordinates, in feet are relative to a common origin selected for the caves in Newsome Sinks.

and two on February 25, 1989. Four of these consist of a plumb-bob strung from the cave roof such that its tip nearly touches the mud. A stiff vertical wire was inserted into the mud directly below the bob. As the mud moves, the tip and the wire separate (Figure 5). One marker was placed at the very apex where the mud and roof meet. It again has a vertical wire in the mud, but has a scratch on the roof instead of a plumb-bob.

The use of rust resistant metal was found desirable. An inexpensive set-up uses sections of bicycle spokes. The threaded end with its tightening head make an effective centered attachment for the string. The string is fed through the central hole of the head and knotted so that it will not pull out. A segment of spoke with the threaded end is screwed into the head. The other end of the spoke segment is sharpened to form the bob-tip. Another spoke section is the vertical wire in the mud. The upper end of the string must be attached to the roof in a way that excludes motion.

The earliest set-ups used screw eyes as plumb bobs, but these are not as satisfactory as the bicycle spoke construction. On a few occasions between observations, the screw eyes became configured so that the string loop through the eye was not at its top, opposite the stem. Perhaps formation of rust generated enough force to move the string position, or perhaps bats hit the string. In these few instances, the string and eye were readjusted to the proper configuration before making measurements. On one occasion, two of the mud markers seemed to have been slightly disturbed between visits. These were reinitialized and measurements for that date discarded.

Measurements of marker separations on 28 dates now span more than three years. Each reading was made independently by each author and the mean recorded. These measurements never disagreed by more than one mm or occasionally two. Thus the precision of the measurement is judged to be plus or minus one mm.

Tilting of the wire in the mud was another concern. It was unclear initially whether a short wire segment in the surface of the mud was desirable or whether a longer segment should be pushed into the mud to observe deeper bulk flow. The shallow wire could be more susceptible to surfaces changes from seasonal drying and wetting. The deeper wire could tilt from differential flow rates with depth. At the position in Figure 5, a shallow wire (few cm) and a deeper wire (more than 10 cm) were emplaced side by side. The bob was over the shorter wire. No appreciable differential tilt was observed in this case. In general, while minor tilts did occur, their effects were significantly smaller than the observed displacements. Thus tilt was a systematic effect influencing accuracy, but not seriously so.

Whereas individual measurements have the uncertainties described (plus or minus a mm or two), these are small relative to the long-term motion. Also, consistency among the results of the five markers supports the reliability of the data.

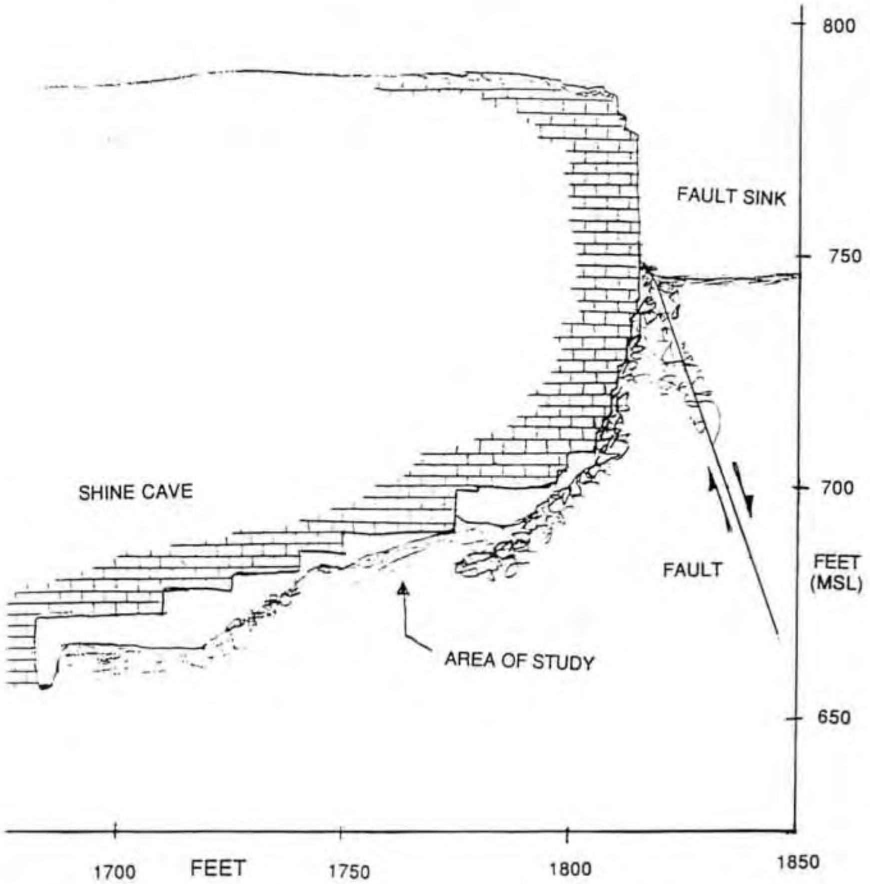

Figure 2: The area of study is on a mud and rubble slope fed by Fault Sink. This profile follows the line A-A in Figure 1. Altitude is in feet above mean sea level.

As a less accurate confirmation, lines were scribed in the mud tangent to the up-slope end of a few drip channels. This was done midway in the measurement period. By the end of the period, the channels had indeed extended beyond the lines by an amount approximately consistent with the observed marker motions.

Figure 3: The study area is shown looking south from a point on the profile in Figure 2.

Results

Figure 6 is a graph for each marker of the displacements versus Julian date. Table 1 shows the mean rate for each marker over the interval it was in place. These data show that a flow rate of a few mm per year is a consistent characteristic of the top (outer) several cm of the bank. This confirms the qualitative explanation offered above for the appearance of the bank and its drip channels.

One of the three years spanned by the measurements, 1988, was slightly drier than normal. The National Weather Service at Huntsville-Decatur Airport recorded 46.66 inches of precipitation relative to the normal 54.72. The Airport is approximately 15 miles from Shine Cave. Years 1989 and 1990 were wetter than normal, with 73.58 and 72.26 inches respectively (NOAA).

An interesting result is illustrated in Table 2 where first the mean rate per year is compared with the displacement between Julian dates 2,448, 218 and 2,448, 261 (1990 Nov. 23 and 1991 Jan 5). During this interval, from December 21 through 23, north Alabama experienced a very heavy rain; Huntsville reported 12.0 inches. In Huntsville the month of December had 18.68 inches of precipitation, which was a record. The previous maximum for a December was 11.74 inches and the previous maximum for any month was 17.00 inches. Flooding was extensive, including the Newsome Sinks caves. Debris from high water indicated that Shine Cave was flooded at least to the base elevation of the study area (*see* Figure 2), and water may even have covered some of the measurement sites. This is very much higher than at any previous time since the measurement began. The last column of Table 2 gives the mean rates if the final measurement interval that spans the flood is not included. Thus it is evident that an episode of heavy rain can cause as much displacement as a year of more usual conditions. On the other hand, even an extreme flood did not produce catastrophic sliding of the mud bank.

Dependence of mud-slope motion on water conditions is to be expected, based on conventional theories for such circumstances. In these theories, soil movement on a slope depends on the balance between the shear strength of the slope materials and the down-slope component of the

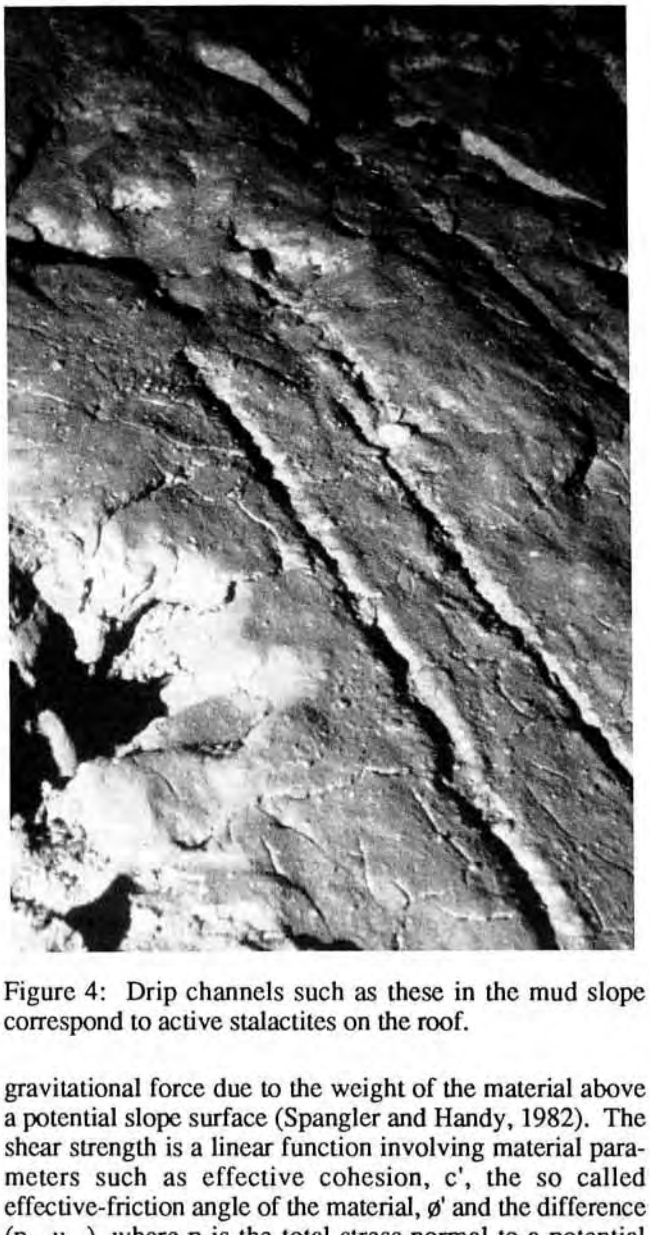

Figure 4: Drip channels such as these in the mud slope correspond to active stalactites on the roof.

gravitational force due to the weight of the material above a potential slope surface (Spangler and Handy, 1982). The shear strength is a linear function involving material parameters such as effective cohesion, c', the so called effective-friction angle of the material, ϕ' and the difference $(p - u_W)$, where p is the total stress normal to a potential slip surface and u_W is the pore-water pressure:

$$s = c' + (p - u_W) \tan \phi'.$$

Here, the important feature of this equation is the negative sign on the pore-water pressure. As this pressure increases, the shear strength at points in the bank decreases. When the shear strength becomes smaller than the down-slope component of the weight of the material above a slip surface, down-slope movement occurs.

When heavy precipitation falls in the collection area of Fault Sink, water drains into the pores of the mud bank below the sink, thus initiating motion. This same scenario and theory is the basis for real-time surface landslide warnings during heavy rainfall in California (Keefer and others, 1987).

Figure 5: This hanging bob and short wire in the mud are marker No. 4. The longer wire goes deeper into the mud to detect any tilt due to differential movement with depth.

That pore-water pressure is indeed present in the Shine Cave slope during wet seasons is indicated by another observed phenomenon. Early in the study, while listening and looking for other drip channels, drip-like noises were detected near the very base of the entire slope, well below the study area. No roof drips could be found. Instead sites on the slope were discovered from which small jets of water rose periodically, making a noise resembling a drip. Pore-water pressure seems to be the likely cause of this phenomenon. Knowing no accepted name for these features, the authors have adopted the name *pird* (drip backwards). Pird activity was observed regularly during moderately wet conditions, but not during dry periods or in extremely wet times when small streams replaced the pirds.

Another aspect of Table 1 may support the simple theoretical picture sketched above. Markers No. 1, 2, and 3 are nearly in a line, down-slope on the northward end of the study area. Markers No. 4 and 5 are near each other at the bottom of the study area toward its southern end. The slope of the bank increases slightly from north to south,

Table 1: Mean Rates

Marker No.	Interval, Julian dates (Last four digits)	Lapsed Time (days)	Displacement (mm)	Mean Rate mm/yr
1	7113 to 8261	1148	10.5	3.5
2	7113 to 8261	1148	14.5	4.6
3	7218 to 8261	1043	15.0	5.3
4	7582 to 8261	679	14.0	7.5
5	7582 to 8261	679	13.0	7.0
Total		4597	67.0	5.3

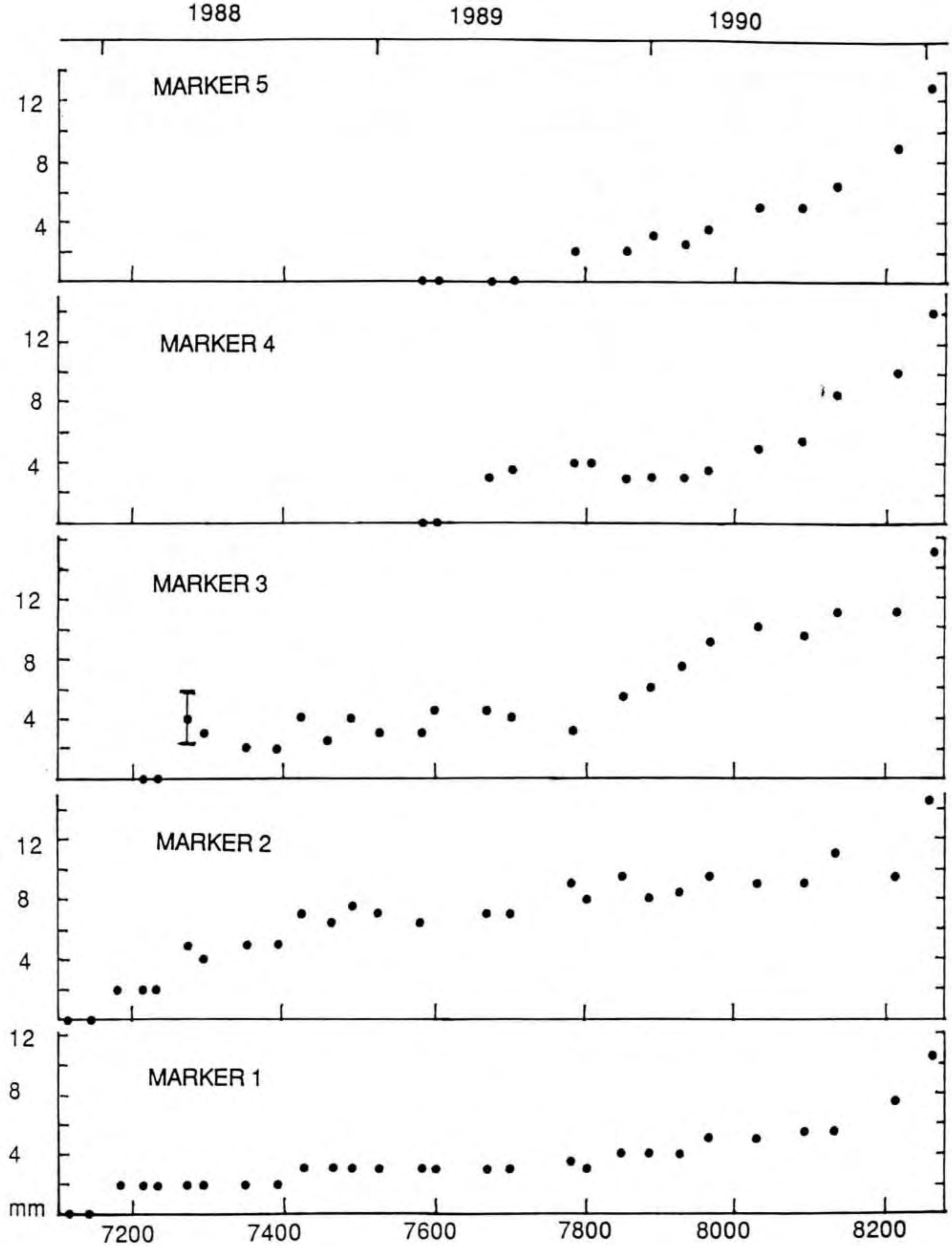

Figure 6: The movement of the five markers is shown in mm relative to the last four digits of Julian Date (2, 44X,XXX).

Table 2: Flood Conditions

Marker No.	Mean Rate mm/yr (Total Data)	Displacement 8218 to 8261 (mm) (43 days)	Mean Rate mm/yr (Excluding last 43 day interval)
1	3.5	3	2.5
2	4.6	5	3.1
3	5.3	4	4.0
4	7.5	4	5.7
5	7.0	4	5.3
Mean	5.3	4.0	3.8

from approximately 30° to 40°. Thus markers 3, 4 and 5 are along the bottom of the study area with the first near its north end and the latter two nearer the south. Other things equal, theory would predict that conditions for motion would occur sooner for the steeper slope condition at the south. The mean rates are higher for markers 4 and 5 than for marker 3, which is consistent with that concept. However, the difference is not great, and may not be statistically significant.

Marker 1 is at the apex where the slope reaches the roof. Its motion could be restrained by the surface friction between the bank and the roof. This effect is not in the theoretical model. The measured rate is indeed lower, but again the difference is not great and may not be significant.

Conclusion

In the aspects discussed above, movements of the mud bank studied in Shine Cave correspond well with the conventional theoretical model accepted for slopes on the surface. While the specific conditions in the study area may not reoccur often, the general situation illustrated in figures 1 and 2 is quite common in Appalachian karsts and elsewhere. Thus the results from Shine Cave may provide a guide to phenomena in similar situations.

References

Jones, W.B. and Varnedoe, W.W., 1980, Caves of Morgan County, Alabama: *Geological Survey of Alabama, Bulletin 112*, 205 p

Keefer, D.K and others, 1987, Real-time landslide warning during heavy rainfall, *Science*, 238, pp. 921-925, 13 Nov.

NOAA, National Climate Data Center, *Local Climatological Data, Annual Summary with Comparative Data, Huntsville, Alabama, Ashville, North Carolina, 1987, 1988, 1989, 1990.*

Spangler, M.G. and Handy, R.L., 1982, *Soil Engineering*: Harper and Row, Publishers, New York.

Varnedoe, W.W., Jr. and Lundquist, C.A., 1986, Speleogenesis in Newsome Sinks, Alabama, USA: *9th Congreso Internacional de Espeleologia, Barcelona, Spain*, v. 1 pp. 258-262.

Varnedoe, W.W., Jr. and Lundquist, C.A., 1988, Plastic flow of cave mud (abstract), *in* Mylroie, J.E. (coordinator), *Program, 10th Friends of Karst Meeting, San Slavador Island, Feb. 11-15*: Mississippi State University, p. 25-26.

Mud Pot: A New Thermal Water Cave in Alleghany County, Virginia

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ABSTRACT

As a part of the Warm River Cave Survey a new cave was discovered in Alleghany County, Virginia. The cave was mapped in February 1990, and was found to contain a thermal stream. This cave and nearby Warm River Cave are the only such thermal caves known in the eastern United States. Both caves are developed in limestones of Middle Ordovician age and are located near the southern end of the Warm Springs anticline. Falling Spring, from which the waters seen in Warm River Cave are known to surface, acts as a regional resurgence for mixed waters of both deep-circulating (hot) and shallow-karst (cold) origin (Hobba and others, 1979). Many proposals have been advanced as to the origin of the thermal waters, and these are summarized below.

In early 1991 dye-tracer tests were conducted in order to establish any hydrologic relationships among Mud Pot and nearby Warm River Cave and Falling Spring. It is concluded that the waters seen in Mud Pot resurge at Falling Spring, but do not pass through Warm River Cave.

Introduction

The Warm River Cave Survey was initiated in March 1989 by Dick Graham and Keith Goggin with the intentions of producing a high quality map of Warm River Cave, as well as to explore and survey other caves in the vicinity. This resulted in the location of several new caves and the rediscovery of one or two previously reported ones. It was during this exploration that a new cave containing thermal waters was found. This new cave and nearby Warm River Cave are the only known caves in the eastern United States containing thermal waters. The presence of hot water, combined with the extremely muddy nature of the cave's lower levels, resulted in the name Mud Pot.

The water temperature in Mud Pot on 3 February 1990 was 28.5 C. Same day readings show this to be 2 C warmer than the water temperature in the "hot" stream of Warm River Cave. Mud Pot's proximity to Warm River Cave and the similarity in water temperatures suggested a hydrologic connection, and a fluorescein dye trace was conducted in order to establish any relationship between the two caves as well as with nearby Falling Spring, a large karst spring which acts as a regional resurgence for waters of both deep-circulating (hot) and shallow-karstic (cold) origin.

The entrance to Mud Pot is a 23-meter vertical shaft, located in a small, brush-filled sinkhole approximately 550 m north of Warm River Cave. This shaft leads to a short segment of canyon passage in which the thermal stream is found. This horizontal passage terminates both upstream and downstream in siphons, with two separate infeeders near the upstream terminus. Figures 1 and 2 are plan and profile maps of Mud Pot.

Location and Geologic Setting

The karst features mentioned in this paper are located in the southern end of the Warm Springs Valley in Alleghany County, Virginia. Figure 3 is a map of the area, showing locations of some caves and karst features found in the area, as well as the implied flow paths of the waters seen in Mud Pot and Warm River Cave, en route to Falling Spring.

To date, little detailed geologic mapping has been done in the southern end of the Warm Springs anticline. The rocks exposed here range from Lower Ordovician to Devonian in age. The karst features described in this report are developed in limestones of Middle Ordovician age. Rader (1984) summarizes the stratigraphy of the western anticlines containing thermal springs, and Bick (1962,

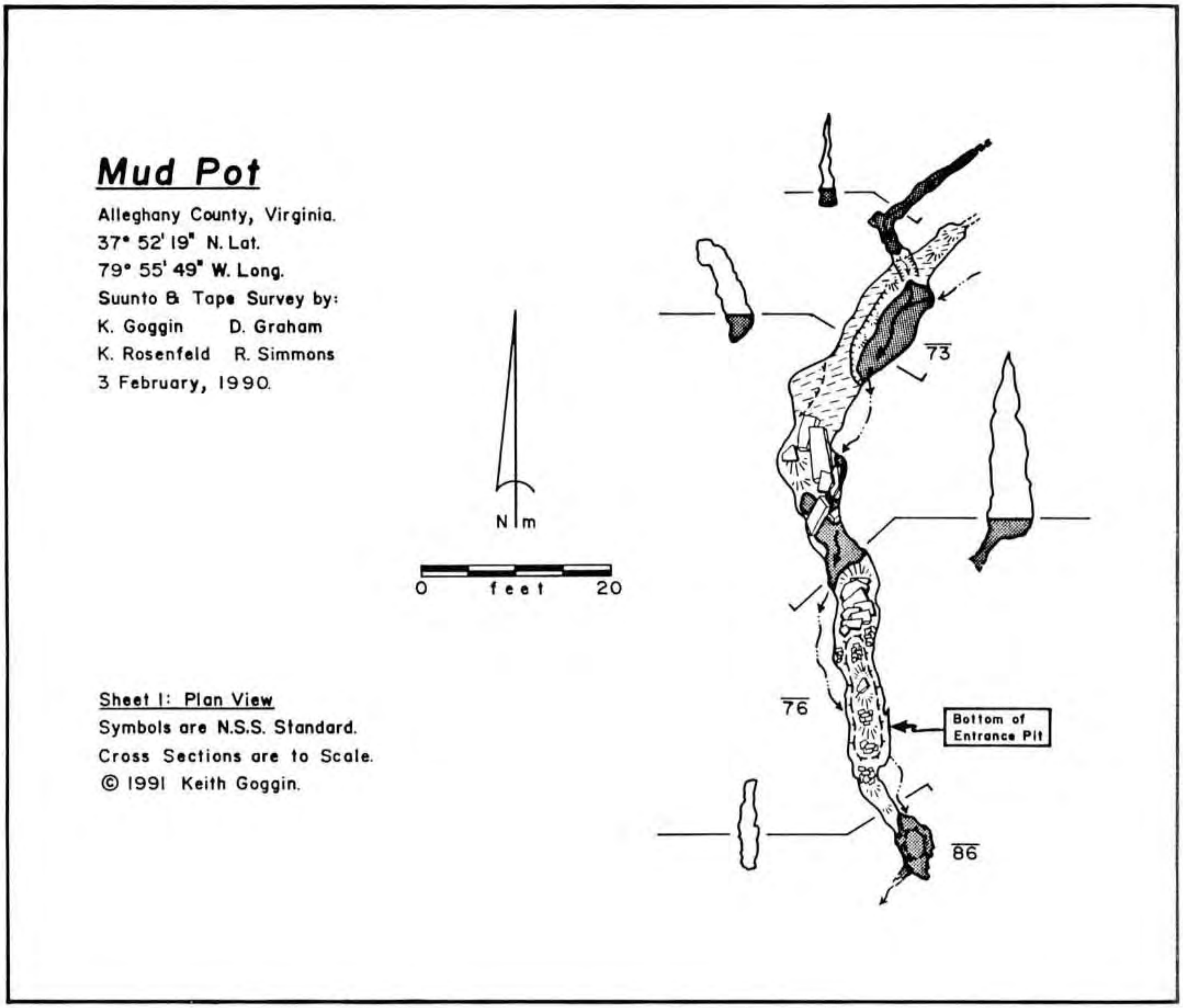

Figure 1: Plan view of Mud Pot.

p. 13) includes a stratigraphic section of the Big Valley Formation (a Middle Ordovician limestone) from about 1.5 km northeast of the entrance to Mud Pot.

Structural features observed both on the surface and within the caves in the area include joint "swarms" and many nearly vertical, normal faults of small displacement. The entrance section of Warm River Cave is developed along one of these normal faults, whereas much of the 'hot stream' section of the cave exhibits joint-controlled development along strike in steeply dipping limestones. Mud Pot appears to be developed along a vertical joint, in nearly horizontal limestone. A slight synclinal flexure is visible inside the cave.

Hobba and others (1979) observed the coincidence of nearly all the thermal springs found in western Virginia with lineaments or intersections of lineaments visible on LANDSAT imagery and high-altitude photography. These lineaments were interpreted to be surface expressions of extensive fracture zones. Figure 4 is a generalized geologic map of the Warm Springs anticline showing the above mentioned lineaments.

A number of papers discuss the geochemical nature of thermal waters emerging from Falling Spring and the geology of the massive travertine deposits located a short distance downstream from the spring. Primary references include Hobba and others (1979), Dennen and Diecchio (1984), Herman and Lorah (1986, 1987), Lorah (1987), Lorah and Herman (1990), and Dennen and others (1990).

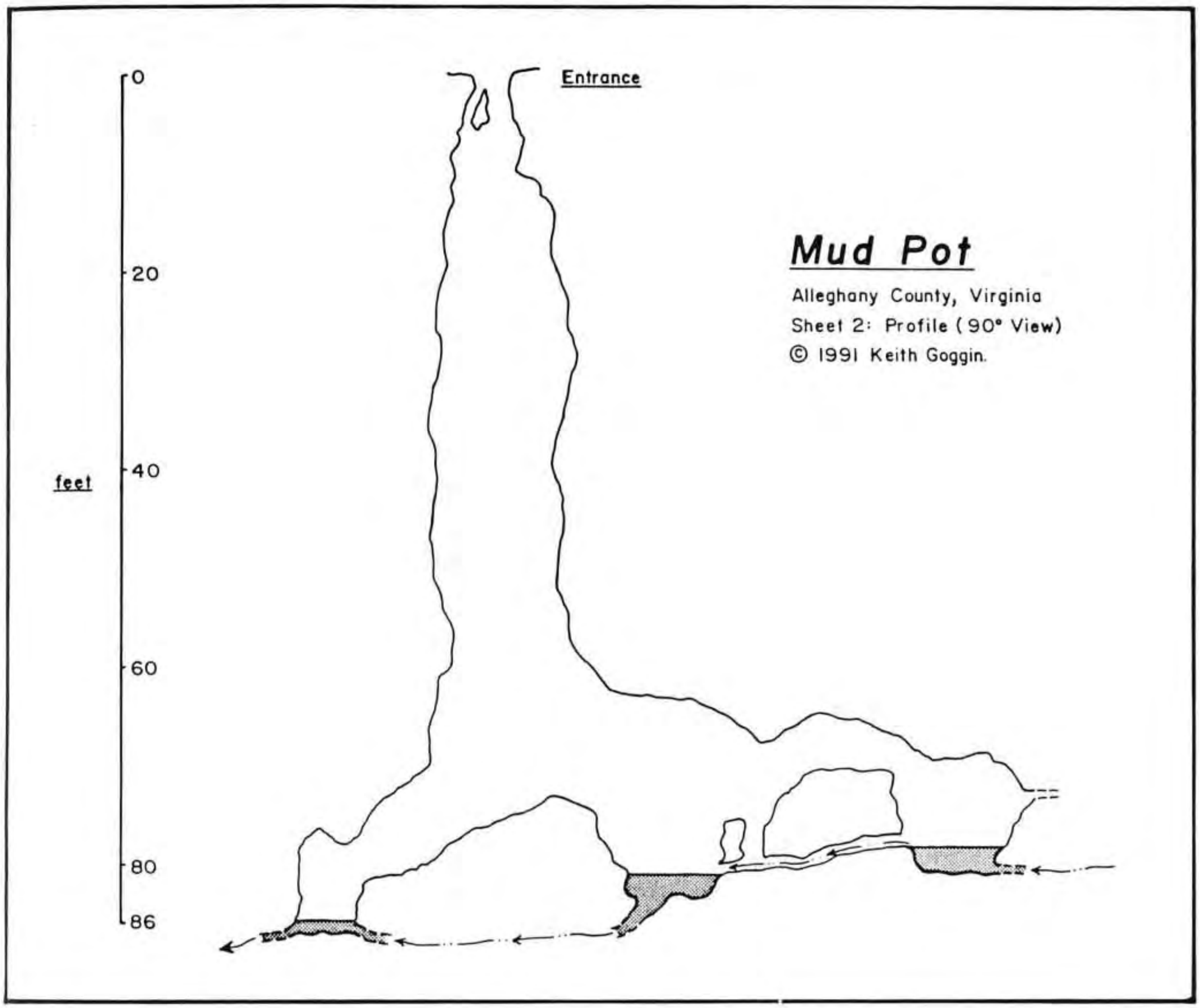

Figure 2: Profile of Mud Pot; view is from the east.

Origin of Thermal Waters in the Warm Springs Anticline

The origin of the thermal waters has been a subject of debate for almost 150 years. The following is a summary, in chronological order, of the major proposals concerning the origin of thermal waters in the Warm Springs anticline.

One of the first reports dealing with this problem was that of Rogers (1843), who believed the waters to be of normal meteoric origin based on their chemical characteristics. Rogers also observed the apparent connection between thermal springs and anticlinal axes and faults. Reeves (1932) expanded on Rogers' hypothesis, agreeing that the thermal waters were of meteoric origin, and suggesting that the waters enter permeable strata at high-altitude exposures on the crest or limb of one anticline and flow in an artesian manner to a similar anticlinal exposure at a lower elevation where the waters resurge as thermal springs.

Dennison and Johnson (1971) note that the structural setting of anticlines in the thermal spring area of Virginia (centered in Bath County) is not significantly different from elsewhere in the Valley and Ridge province extending from New York to Alabama. They also note the presence of middle Tertiary igneous intrusions just to the north of the thermal springs area of Virginia. They suggest that the heat source for the thermal springs may be residual heat in a deep, solidified pluton.

Helz and Sinex (1974) furthered Reeves' hypothesis by plotting elevations of thermal springs in the Warm Springs Valley, finding all to be below 700 m. They suggest that the recharge area for the thermal waters is probably the Browns Mountain anticline located to the west in Greenbrier and Pocahontas counties, West Virginia. They

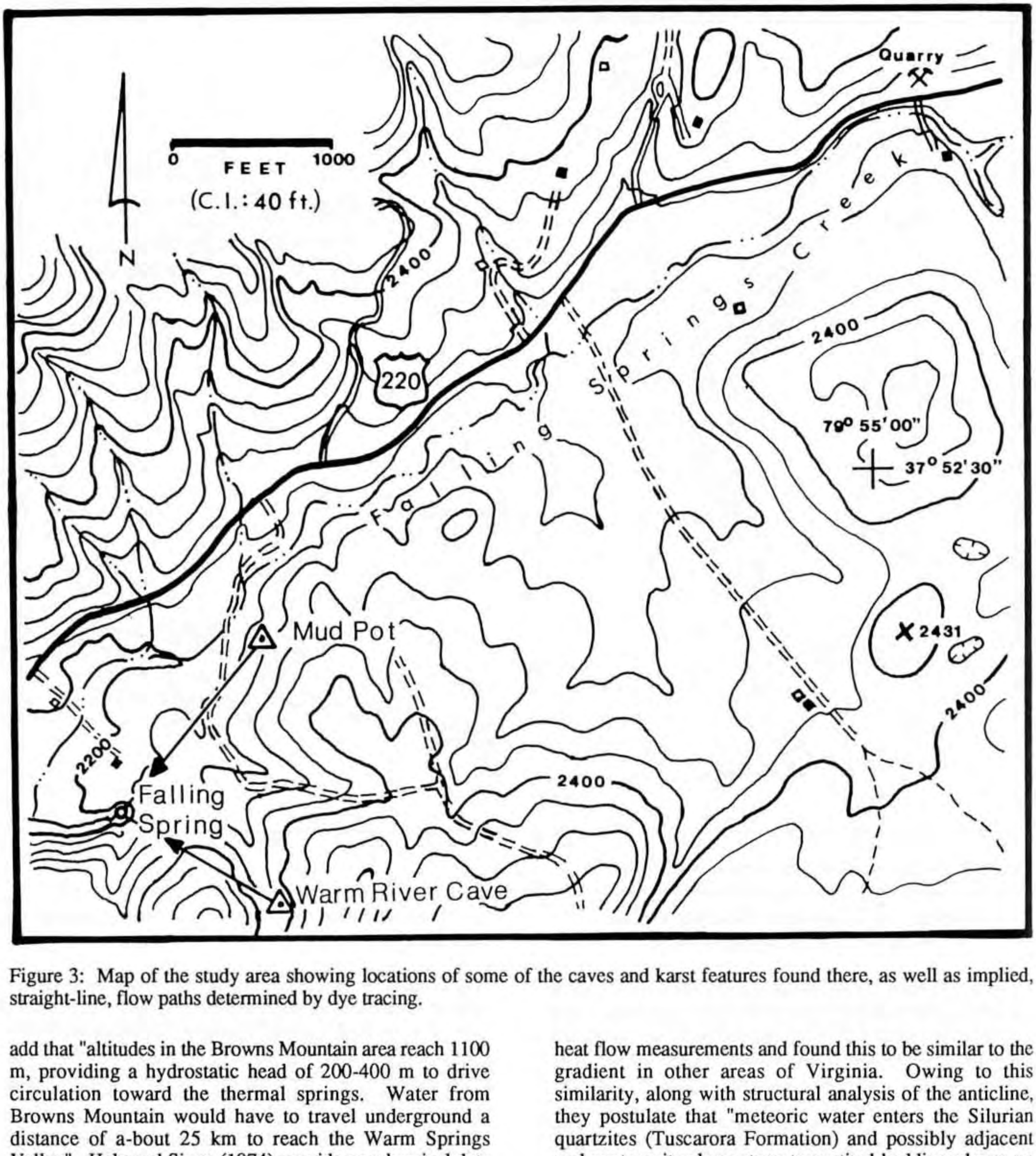

Figure 3: Map of the study area showing locations of some of the caves and karst features found there, as well as implied, straight-line, flow paths determined by dye tracing.

add that "altitudes in the Browns Mountain area reach 1100 m, providing a hydrostatic head of 200-400 m to drive circulation toward the thermal springs. Water from Browns Mountain would have to travel underground a distance of a-bout 25 km to reach the Warm Springs Valley". Helz and Sinex (1974) provide geochemical data that show that the thermal waters do not require a magmatic heat source as suggested by Dennison and Johnson (1971), but this data does not rule out the possibility of residual plutonic heat.

Perry and others (1979) calculated the geothermal gradient in the Warm Springs anticline using down-hole heat flow measurements and found this to be similar to the gradient in other areas of Virginia. Owing to this similarity, along with structural analysis of the anticline, they postulate that "meteoric water enters the Silurian quartzites (Tuscarora Formation) and possibly adjacent carbonate units along steep to vertical bedding planes on the northwest limb of the (Warm Springs) anticline, reaches depths sufficient to heat the water, and then rises rapidly along essentially vertical east-west fracture zones which intersect the bedding plane permeability at depth". Perry and others conclude that their findings do not support Dennison and Johnson's (1971) proposal of a residual magmatic heat source.

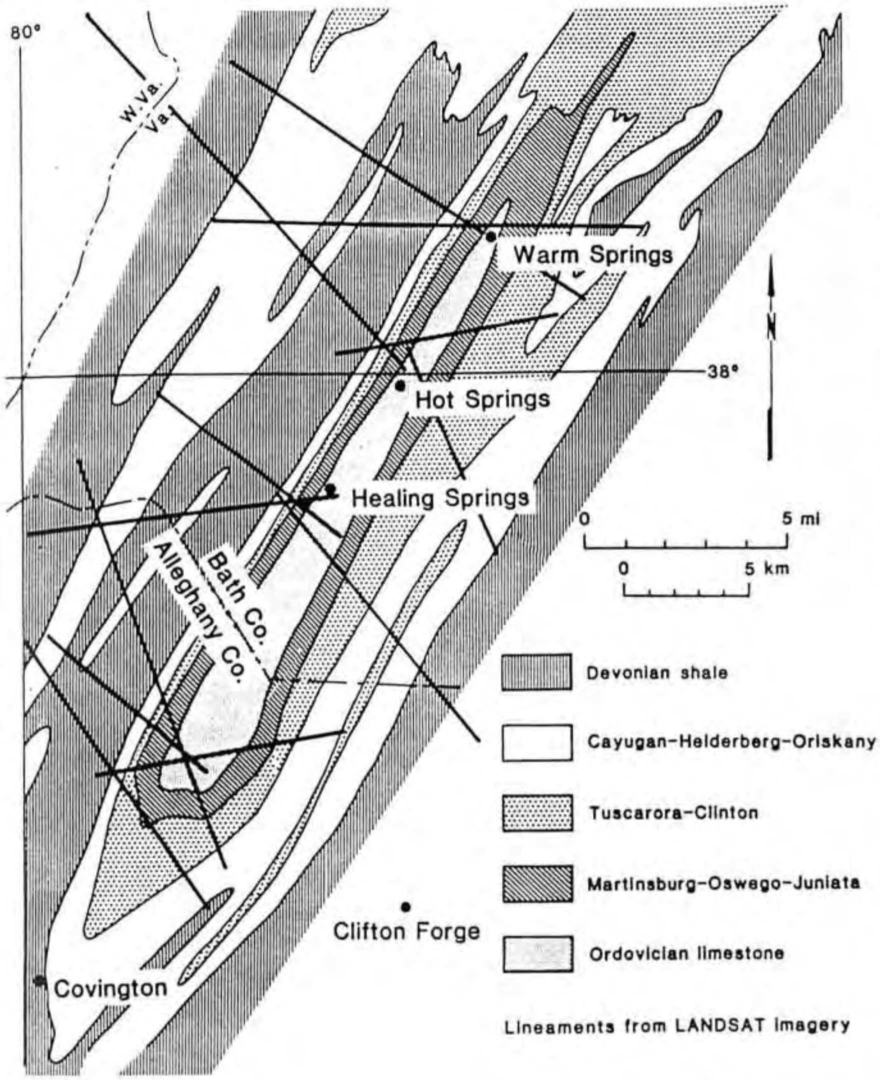

Figure 4: Generalized geologic map of the Warm Springs anticline (adapted from Dennen and Diecchio, 1984). The study area is located close to the intersection of three lineaments visible near the southern terminus of the Ordovician limestone exposure.

Receptor #	Location	Positive	Negative
1	Mouth of "cold-stream"		X
2	Base of climb-down to water, mixed stream		X
3	Thermal infeeder near base of rimstone dams		X
4	Thermal infeeder on west side of passage		X
5	"Hot" stream, in pool at mud slide room		X
6	Final duck-under, 30 m from terminus		X
7	Mouth of "hot" stream		X
8	Falling Spring	X	

Table 1: Description of receptor locations shown in Figure 5 and results (cumulative data from both tests).

Dye-Tracing Methods

Two dye traces were conducted to ascertain the subsurface flow path of water seen in Mud Pot. Receptors containing activated charcoal were constructed from fine nylon screening in the manner suggested by Aley and Fletcher (1976). In order to enable the receptors to hang freely in the water and ensure maximum dye recovery, modified "gumdrops" were constructed in the field using medium-gauge metal wire wrapped around suitable rocks found near each location. Before dye was injected into the groundwater, a receptor was placed at Falling Spring. This control receptor was used to establish background levels of any pre-existing substances (*i.e.* stock waste) in the groundwater that could interfere with visual observations used to determine test results.

In both tests, dye receptors were placed at strategic locations in Warm River Cave and at Falling Spring. Numbered receptor locations in Warm River Cave are shown in Figure 5 and described in Table 1. It should be noted that no receptors were placed in the extreme downstream reaches of Warm River Cave, but there exist no significant infeeders, thermal or otherwise, in this section of the cave.

Once the receptors were in place, approximately 0.9 kg of fluorescein dye (Acid Yellow 73) was placed in the stream flowing through Mud Pot. A 5% solution of potassium hydroxide (KOH) in 70% isopropyl alcohol was used as an elutriant for all tests.

First Trace Results

The first trace was executed on 5 January, 1991. Because water levels in Warm River Cave were very high, receptors were placed only at locations 1, 2, and 7, as well as at Falling Spring (location 8). After one week, receptors at the spring were recovered, but those inside Warm River Cave were inaccessible owing to severe flood conditions. The recovered receptors were elutriated in the KOH solution and were found to be "very strongly positive" using the criteria by Aley and Fletcher (1976). This confirmed the

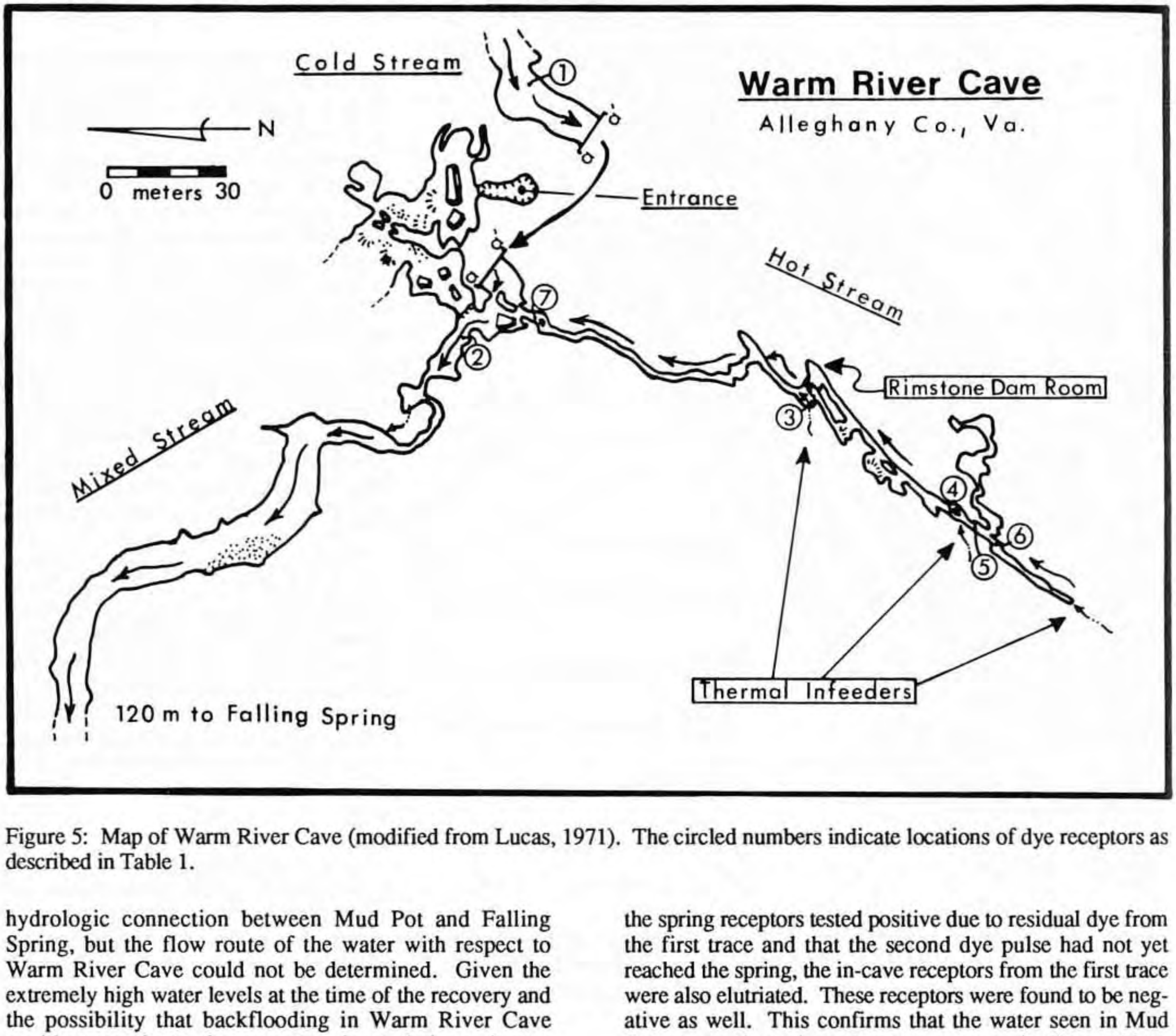

Figure 5: Map of Warm River Cave (modified from Lucas, 1971). The circled numbers indicate locations of dye receptors as described in Table 1.

hydrologic connection between Mud Pot and Falling Spring, but the flow route of the water with respect to Warm River Cave could not be determined. Given the extremely high water levels at the time of the recovery and the possibility that backflooding in Warm River Cave could contaminate the receptors located there, it was decided to repeat the trace.

Second Trace Results and Conclusions

The second trace was conducted on 24 February, 1991. Water levels, while still high, were low enough to allow receptors to be placed at all locations including Falling Spring. At this time receptors remaining in the cave from the first test were collected.

After one week, all receptors were successfully retrieved and elutriated. Once again the spring receptors tested strongly positive, but all the receptors from Warm River Cave tested negative, suggesting that the water from Mud Pot resurges at Falling Spring after following a flow path totally independent from known passages in Warm River Cave. To remove the very slight possibility that the spring receptors tested positive due to residual dye from the first trace and that the second dye pulse had not yet reached the spring, the in-cave receptors from the first trace were also elutriated. These receptors were found to be negative as well. This confirms that the water seen in Mud Pot passes through no known passages in Warm River Cave en route to Falling Spring.

Acknowledgments

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Special thanks are directed to Gustav Asboth and Teddy Dressler, the landowners on whose land the caves and Falling Spring are located. Without their kind permission none of this work would have been possible.

References

Aley, T. and Fletcher, M.W., 1976, Water Tracer's Cookbook: *Missouri Speleology*, v. 16, no. 3, 32 p.

Bick, K.F., 1962, Geology of the Williamsville Quadrangle, Virginia: *Virginia Division of Mineral Resources Report of Investigations.*, no. 2, 40 p.

Dennen, K.O. and Diecchio, R.J., 1984, The Falling Spring, Alleghany County, Virginia, *in* Rader, E.K. and Gathright, T.M., II (editors), Stratigraphy and Structure in the Thermal Springs Area of the Western Anticlines: *Virginia Division of Mineral Resources Sixteenth Annual Virginia Geologic Field Conference*, p. 17-22.

Dennen, K.O.; Diecchio, R.J.; and Stephenson, M.A., 1990, The geology of the Falling Spring travertine deposit, Alleghany County, Virginia, *in* Herman, J.S. and Hubbard, D.A., Jr. (editors), Travertine-marl: Stream deposits in Virginia: *Virginia Division of Mineral Resources Publication 101*, p. 79-91.

Dennison, J.M. and Johnson, R.W., 1971, Tertiary intrusions and associated phenomena near the Thirty-Eighth Parallel Fracture Zone in Virginia and West Virginia: *Geological Society of America Bulletin*, v. 82, p. 501-508.

Helz, G.R. and Sinex, S.A., 1974, Chemical equilibria in the thermal spring waters of Virginia: *Geochemica et Cosmochimica Acta*, v. 38, p. 1807-1820.

Herman, J.S. and Lorah, M.M., 1986, Groundwater geochemistry in Warm River Cave, Virginia: *National Speleological Society Bulletin*, v. 48, p. 54-61.

Herman, J.S. and Lorah, M.M., 1987, CO_2 Outgassing and calcite precipitation in Falling Spring Creek, Virginia, U.S.A.: *Chemical Geology*, v. 62, p. 251-262.

Hobba, W.A., Jr.; Fisher, D.W.; Pearson, F.J., Jr.; and Chemerys, J.C., 1979, Hydrology and geochemistry of thermal springs of the Appalachians: *U.S. Geological Survey Professional Paper 1044-E*, 36 p.

Lorah, M.M., 1987, *The Chemical Evolution of a Travertine-Depositing Stream*, M.S. thesis (unpublished): Charlottesville, University of Virginia, 175 p.

Lorah, M.M. and Herman, J.S., 1990, Geochemical evolution and calcite precipitation rates in Falling Spring Creek, Virginia, *in* Herman, J.S. and Hubbard, D.A., Jr. (editors), Travertine-marl: Stream deposits in Virginia: *Virginia Division of Mineral Resources Publication 101*, p. 17-32.

Lucas, P.C., 1971, Map of Warm River Cave, *in* Virginia Region of the National Speleological Society, 1971, National Speleological Society Convention Guidebook (Section 1): *The Region Record*, v. 1, no. 4, p. 79.

Perry, L.D.; Costain, J.K.; and Geiser, P.A., 1979, Heat flow in western Virginia and a model for the origin of thermal springs in the folded Appalachians: *Journal of Geophysical Research*, v. 84, p. 6875-6883.

Rader, G.K., 1984, Stratigraphy of the western anticlines - A summary, *in* Rader, G.K. and Gathright, T.M., II (editors), Stratigraphy and Structure in the Thermal Springs Area of the Western Anticlines: *Virginia Division of Mineral Resources Sixteenth Annual Virginia Geologic Field Conference*, p. 1-12.

Reeves, F., 1932, Thermal springs of Virginia: *Virginia Geological Survey Bulletin 36*, 56 p.

Rogers, W.B., 1843, On the connection of thermal springs in Virginia with anticlinal axes and faults: *Association of American Geologists and Naturalists Report of the 1st, 2nd, and 3rd Meeting, 1840-1842*, p. 323-347.

Plate A: Travertine-marl deposit at entrance to Wildwood Park, Norwood Street, City of Radford, Virginia. This deposit has formed on the wall of an artificial cut in the hillside where groundwater now seeps from bedding partings in the Elbrook limestone. Note the extensive accumulation of algal material and the steady dripping. This site was a stop on the geologic fieldtrip of the Appalachian Karst Symposium. For a discussion of travertine-marl deposits *see* Hubbard and Herman, this volume, p. 59. *Photograph by Karen M. Kastning.*

Travertine-Marl: The "Doughnut-Hole" of Karst©

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ABSTRACT

Karst is a negative relief topography formed by the dissolution of carbonate bedrock. Emergent karstic water, as springs or diffuse stream-bank or -bed seeps, has deposited Quaternary age travertine-marl buildups. Most of the travertine-marl deposits in Virginia are associated with known faults in folded and fractured carbonate rock. Some of these accretionary or positive karst features morphologically resemble spelean formations typical of depositional vadose cave environments. The ubiquitous presence of algae and moss in travertine-marl differentiate it from its spelean counterparts, but whether the role of the biota is only as a passive framework or is also as an active metabolic influence remains to be determined.

Locally valued for their aesthetic waterfalls and as a source of agricultural lime, travertine-marl deposits are an environmental barometer. Land-use practices resulting in increased runoff of precipitation and increased erosion have degraded these features: travertine-marl deposits have been smothered and partially buried by erosional debris; features have been destroyed by the corrasive action of large magnitude floods carrying large volumes of abrasive sediment. Polluted karstic water may inhibit calcite nucleation and the growth of framework flora, both of which influence the rate of travertine-marl deposition.

Introduction

Travertine-marl deposits of the Valley and Ridge physiographic province are composed of calcium carbonate precipitated from streams and springs. These accretionary fresh-water carbonate deposits are a common feature in karst areas of western Virginia (Figure 1; Sweet and Hubbard, 1990). Although karst is characterized as a negative relief topography formed by the dissolution of carbonate bedrock, chemical conditions may permit calcium carbonate to precipitate from karstic water. Travertine deposits in caves are well known to cave visitors and karst researchers. The term travertine-marl is used here to denote complex deposits of calcium carbonate, precipitated by surface streams and springs, comprised of travertine, tufa, and marl components. Some features of travertine-marl deposits morphologically resemble their spelean counterparts (Love and Chafetz, 1990). In surface streams, travertine-marl deposits commonly form impressive waterfalls. The buildups of travertine and tufa which create these waterfalls allow marl to accumulate upstream as the buildups simultaneously aggrade. In addition to their aesthetic waterfalls, these freshwater carbonate deposits have been utilized as a source of agricultural lime (Sweet and Hubbard, 1990) and for building stone (Austin and Barker, 1990).

Depositional Processes

Karstic groundwater, enriched in carbon dioxide, aggressively dissolves carbonate minerals until it reaches equilibrium with respect to these soluble minerals. Upon emergence at discrete and diffuse springs, karstic water often becomes supersaturated with respect to carbon dioxide as the gas exsolves from the stream water. Outgassing of carbon dioxide in response to the low atmospheric level of carbon dioxide drives the water to high degrees of supersaturation with respect to the carbonate minerals, especially calcite. Hydrological agitation and the dispersion of water over stream-bottom obstructions influence the rate of carbon dioxide exchange. Travertine buildups contribute to higher levels of agitation and dispersion and provide a source of positive feedback to the outgassing process and subsequent precipitation of travertine.

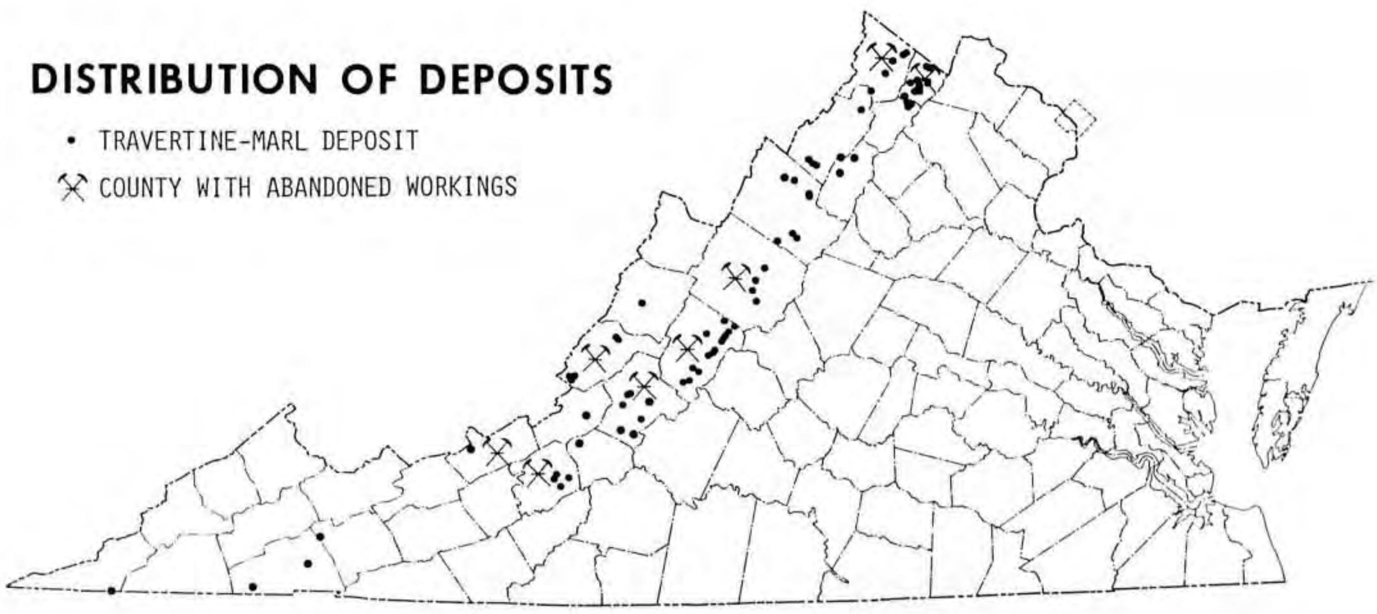

Figure 1: Travertine-marl deposits found in Alleghany, Augusta, Bath, Botetourt, Clarke, Craig, Frederick, Giles, Lee, Montgomery, Page, Roanoke, Rockbridge, Rockingham, Shenandoah, Smith, Warren, and Washington counties (Modified from Hubbard and others, 1985).

In Virginia, calcite precipitation is kinetically inhibited until stream water is five to six times supersaturated in Falling Spring Run (Hoffer-French and Herman, 1990) and five to nine times supersaturated in Falling Spring Creek (Lorah and Herman, 1990). Factors which influence calcite precipitation rates include the presence of organic material or strongly adsorbing ions. Algae and mosses seasonally increase the surface area available for nucleation and precipitation of calcite. Whether biological processes actively drive the deposition of travertine-marl has not been resolved. A 24-hour study of Falling Spring Creek failed to detect any evidence of photosynthetic activity affecting the partial pressure of CO_2 or the saturation index of calcite (Lorah and Herman, 1988). The physical proces-ses of hydrological agitation and CO_2 outgassing in re-sponse to a chemical gradient predominated over the biological effects on CO_2 uptake by stream biota in another diurnal and seasonal study of a travertine-depositing stream in Virginia (Hoffer-French and Herman, 1989). Studies at Plitvice National Park, Yugoslavia indicate that algal mucous excretions trap particles of calcite and serve to initiate precipitation of travertine (Emeis and others, 1987). Researchers, such as Emeis and others (1987) and Pentecost (1990) suggest that the metabolic involvement of algae may have a more significant role in the deposition of travertine. Chemical processes also influence rates of travertine formation. For example, calcite precipitation is inhibited by concentrations of strongly adsorbed ions that block nucleation sites (Morse, 1983). In this complex natural system influenced by chemical, hydrological, and biological processes, the rate of travertine deposition must exceed the rate of travertine erosion for these deposits to form and exist over time.

Travertine-Marl Deposits

Travertine, tufa, and marl have been deposited by stream and spring water in at least 18 of the 26 karst-bearing counties in the Valley and Ridge physiographic province of Virginia (Figure 1; Sweet and Hubbard, 1990). Of the approximately 70 deposits, most are associated with known faults in folded and fractured carbonate rock. A considerable range in morphology is exhibited by these travertine-marl deposits. Extensive low-relief deposits dominated by marl are typical of the sites in the northern part of Virginia. Travertine and tufa buildups (Figure 2), forming bluffs or falls in active stream channels, are notable components in most deposits throughout the State. Marl accumulations (Figure 3) typically are located upstream of buildups suggesting that carbonate sediments simultaneously accumulated behind aggrading buildups. Deposits are characterized by two morphological signatures that can be recognized in the field or remotely sensed. Buildups create waterfalls and cascades in depositing streams, whereas marl accumulations form broad flat fills or terraces. Additional keys to field recognition include fragments of travertine and tufa located downstream of incised deposits and the presence of travertine and tufa debris or an abundance of gastropod shells in stream banks; however, most marl deposits have been modified by soil development and may not be recognized in the absence of coarse carbonate debris without the aid of acid.

Figure 2: Beaverdam Falls, a buildup of travertine and tufa on Sweet Springs Creek, Alleghany County, Virginia (McDonald and Bird, 1986, cover).

Close examination of the travertine components of the deposits reveals that this material is highly porous. Organic detritus, including twigs and leaves trapped on buildups, greatly increases the available surface area for calcite nucleation and deposition in the late summer and fall. Moss and algae are commonly observed on many buildups and provide a framework for nucleation and deposition (Figure 4). Decay of the organic materials after entombment in calcite results in a porous travertine or tufa. Some travertine components are laminated. Bands of denser travertine are separated by porous partings. Each band and parting is thought to represent an annual cycle of travertine deposition from summer through spring. Steidtmann (1934) characterized a travertine deposit near Lexington with "winter layers being thin, compact, and relatively clastic, whereas the summer layers are "mossy" and highly calcitic."

Many of the features of travertine buildups morphologically resemble travertine features in caves. Both dripstone and flowstone cave features tend to be dense, lacking the porosity associated with the decay of organic substrates that are typical of surficial travertine and tufa.

Environmental Factors in Deposit Degradation

Most travertine-marl deposits, currently, appear to be undergoing net erosion. Accumulations of marl are entrenched, travertine buildups are incised, and large fragments of travertine and tufa are found downstream from broken buildups. Thornton (1953) suggested that travertine-marl deposits are relicts and were deposited during past climates. Steidtmann (1934), however, argues that the present destructive erosion of the deposits is a consequence of land cultivation. Muddy waters are less favorable for the growth of algae and moss (Steidtmann, 1935a), which serve as "both a framework and a protective cover for the growing travertine" (Steidtmann, 1935b, p. 334). Significant damage to travertine-marl deposits is caused by floods (Emig, 1917; Steidtmann, 1936; Chafetz and Folk, 1984). Hurricane Camille, in 1969, had a catastrophic effect on travertine buildups in Moores Creek and at Gibbs

Figure 3: Extensive accumulation of marl along Redbud Run, Frederick County, Virginia. Information on commercial mining of this deposit in Giannini, 1990. (Photograph by S.O. Bird, 1984.)

Falls in Rockbridge County, Virginia. One observer (Phillip Lucas, 1984, personal communication) estimated only 20 percent of the travertine still remains at Moores Creek Falls. Considerable flood damage to deposits in Rockbridge and Alleghany Counties has been observed by the authors.

Steidtmann's ideas that the destruction of travertine-marl deposits are a consequence of land cultivation and flooding are better termed a consequence of land-use modification. Siltation resulting from past land clearing and tillage practices have introduced tremendous quantities of clastic sediment into travertine-depositing streams. This sediment has buried some deposits (Thornton, 1953) and diluted the purity of the marl in other deposits during flooding (Sweet and Hubbard, 1990). Examination of the sediments in the streams in the vicinity of travertine buildups reveals that the carbonate content is significantly lower than in the adjacent entrenched marl. The noncarbonate fraction of this sediment also serves as an abrasive. Analyses of active and relict travertine from 10 deposits indicate that active travertine generally contains more noncarbonate material than relict travertive (Herman and Hubbard, 1990). This finding is the reverse case of the relative decrease in carbonate content that one would expect of exposed weathered relict travertine. The clearing and modification of land surfaces generally results in greater runoff of precipitation. In addition to carrying additional sediment, increased runoff generates higher magnitude flood pulses. The erosive potential of the higher magnitude floods of corrasive sediment-laden waters to the travertine-marl deposits is far greater than for the flood events predating the agricultural practices of the last two and one half centuries.

Degradation of travertine-marl deposits also results from pollution of groundwater. The presence of pollutants in emergent groundwater or in a calcite-precipitating stream may adversely affect the growth of algal and moss substrates or reduce the volume of calcite precipitated. Strongly adsorbing ionic species can inhibit calcite growth by bonding to calcite crystals and effectively poisoning growth sites. Magnesium, phosphate, and some organic compounds notably inhibit calcite precipitation (Morse, 1983). Pentecost (1990, p. 125) remarks that "high levels

Figure 4: Algal drapes cemented by calcite in Falling Spring Run, Augusta County, Virginia.

of dissolved phosphates may ... encourage plant growth but discourage travertine formation."

Summary

Most karst features are the result of solutional processes wherein carbonate bedrock is dissolved, forming a topography characterized by closed-contour depressions and internal drainage. Emergent karstic water containing high levels of dissolved calcium carbonate precipitate calcite as anomalous positive karst features. Travertine-marl deposits, comprised of travertine and tufa buildups and associated upstream accumulations of marl, are widespread in the Valley and Ridge physiographic province of Virginia and have been economically utilized as a source of agricultural lime. The commercial mining of these deposits has ceased in Virginia partially as a result of concerns for the environmental consequences of natural resource utilization. It is ironic that at a time of environmental awareness, when the aesthetics of these waterfall-generating deposits overshadows their mineral worth, the environmental consequences of land-use modifications are destroying these natural resources. Siltation from the clearing and tillage of land and higher magnitude floods fed by runoff from artificial and modified land surfaces have severely eroded many of these freshwater carbonate deposits. Polluted waters adversely affect the growth of algal and moss frameworks or inhibit travertine deposition by blocking crystal growth. The degeneration of travertine-marl deposits has been attributed to climatic change; more realistically, it heralds the decline of water resources by siltation and pollution.

References

Austin, G.S. and Barker, J.M., 1990, Commercial travertine in New Mexico: *New Mexico Geology*, v. 12, no. 3, p. 49-58.

Chafetz, H.S. and Folk, R.L., 1984, Travertines: depositional morphology and the bacterially constructed constituents: *Journal of Sedimentary Petrology*, v. 54, no. 1, p. 2289-2316.

Emeis, K.C.; Richnow, H.H.; and Kempe, S., 1987, Travertine formation in Plitvice National Park, Yugoslavia: Chemical versus biological control: *Sedimentology*, v. 34, p. 595-609.

Emig, W.H., 1917, Travertine deposits of Oklahoma: *Oklahoma Geological Survey Bulletin 29*, 76 p.

Giannini, W.F., 1990, A commercial marl deposit near Winchester, Virginia, *in* Herman, J.S. and Hubbard, D.A., Jr. (editors), Travertine-marl: Stream deposits in Virginia: *Virginia Division of Mineral Resources Publication 101*, p. 93-99.

Herman, J.S. and Hubbard, D.A., Jr., 1990, A comparative study of travertine-marl-depositing streams in Virginia, *in* Herman, J.S. and Hubbard, D.A., Jr. (editors), Travertine-marl: Stream deposits in Virginia: *Virginia Division of Mineral Resources Publication 101*, p. 43-64.

Hoffer-French, K.J. and Herman, J.S., 1989, Evaluation of hydrological and biological influences on CO_2 fluxes from a karst stream: *Journal of Hydrology*, v. 108, p. 189-212.

Hoffer-French, K.J. and Herman, J.S., 1990, A CO_2 outgassing model for Falling Spring Run, Augusta County, Virginia, *in* Herman, J.S. and Hubbard, D.A., Jr. (editors), Travertine-marl: Stream deposits in Virginia: *Virginia Division of Mineral Resources Publication 101*, p. 5-15.

Hubbard, D.A., Jr.; Giannini, W.F.; and Lorah, M.M., 1985, Travertine-marl deposits of the Valley and Ridge province of Virginia - A preliminary report: Virginia Division of Mineral Resources, *Virginia Minerals*, v. 31, no. 1, p. 1-8.

Lorah, M.M. and Herman, J.S., 1988, The geochemical evolution of a travertine-depositing stream: Geochemical processes and mass transfer reactions: *Water Resources Research*, v. 24, no. 9, p. 1541-1552.

Lorah, M.M. and Herman, J.S., 1990, Geochemical evolution and calcite precipitation rates in Falling Spring Creek, Virginia, *in* Herman, J.S. and Hubbard, D.A., Jr. (editors), Travertine-marl: Stream deposits in Virginia: *Virginia Division of Mineral Resources Publication 101*, p. 17-32.

Love, K.M. and Chafetz, H.S., 1990, Petrology of Quaternary travertine deposits, Arbuckle Mountains, Oklahoma, *in* Herman, J.S. and Hubbard, D.A., Jr. (editors), Travertine-marl: Stream deposits in Virginia: *Virginia Division of Mineral Resources Publication 101*, p. 65-78.

McDonald, J.N. and Bird, S.O. (editors), 1986, The Quaternary of Virginia - A symposium volume: *Virginia Division of Mineral Resources Publication 75*, 137 p.

Morse, J.W., 1983, The kinetics of calcium carbonate dissolution and precipitation, *in* Reeder, R.J. (editor), Carbonates: Mineralogy and chemistry: Washington, D.C., Mineralogical Society of America, *Reviews in Mineralogy*, v. 11, p. 227-264.

Pentecost, A., 1990, The algal flora of travertine: An overview, *in* Herman, J.S. and Hubbard, D.A., Jr. (editors), Travertine-marl: Stream deposits in Virginia: *Virginia Division of Mineral Resources Publication 101*, p. 117-127.

Steidtmann, E., 1934, Travertine deposits near Lexington, Virginia (abstract): *Proceedings of the.Virginia Academy of Sciences*, p. 56.

Steidtmann, E., 1935a, Travertine near Lexington, Virginia (abstract): *The American Mineralogist*, v. 20, no. 3, p. 206.

Steidtmann, E., 1935b, Travertine depositing waters near Lexington, Virginia: *Science*, v. 82, no. 2127, p. 333-334.

Steidtmann, E., 1936, Travertine-depositing waters near Lexington, Virginia: *Journal of Geology*, v. 44, no. 2, p. 193-200.

Sweet, P.C. and Hubbard, D.A., Jr., 1990, Economic legacy and distribution of Virginia's Valley and Ridge province travertine-marl deposits, *in* Herman, J.S. and Hubbard, D.A., Jr. (editors), Travertine-marl: Stream deposits in Virginia: *Virginia Division of Mineral Resources Publication 101*, p. 129-138.

Thornton, C.P., 1953, *The Geology of the Mount Jackson Quadrangle, Virginia*: Ph.D. dissertation (unpublished): Yale University, New Haven, Connecticut, 211 p.

Meteorology of the Butler Cave-Sinking Creek System

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ABSTRACT

The Butler Cave Conservation Society (BCCS), Inc. is conducting a meteorological study of the Butler Cave-Sinking Creek System in west-central Virginia. The study consists of an Entrance Project dealing with data from the passages forming the route from The Entrance to the Trunk Channel deep inside the cave, and a Trunk Channel Project dealing with data from this 9,000-foot-long passage. More than 550 pairs of temperature measurements (wet-bulb and dry-bulb) have been made since the study began in April 1984. The variations of the temperature, partial pressure of water vapor, and relative humidity are being studied as functions of time (season) and position (within the cave).

This paper reports preliminary results of the Trunk Channel Project. The sixteen measurement positions employed are all located more than 1,000 feet horizontally from any known air entry points of the cave and at least 150 feet vertically below the surface. Far inside the cave, meteorological conditions are remarkably stable. For example, seventy-one measurements made at one position (Sand Canyon) yield the following means and standard deviations: temperature 51.4 ± 0.4 F, partial pressure of water vapor 0.375 ± 0.005 inches of Mercury, and relative humidity $99.0 \pm 1.0\%$.

Three major trends are visible in the Trunk Channel data: gradually increasing temperature as one goes downstream (51.1 F increasing to 52.2 F); gradually increasing partial pressure of water vapor as one goes downstream (0.367 increasing to 0.385 in Hg); and relative humidities greater than 97% at most places and times.

Four minor occurrences in the data require further study: unusually low temperatures (~1 F lower) at one position; unusually low partial pressures of water vapor (~0.01 in Hg lower) at two positions; relative humidities as low as 92%; and higher variability in relative humidity at some positions (6% variability versus a more normal 3%). Explanations are suggested for the occurrences observed in the data.

Background

The Butler Cave Conservation Society (BCCS), Inc. is conducting a study of the meteorology of the Butler Cave-Sinking Creek System (*see* Figure 1). Owned and managed by the BCCS, this non-commercial cave is located in Burnsville Cove, Bath County, west-central, Virginia. It consists of 17+ miles (27+ kilometers) of passages with a single human-traversable entrance of less than ten square feet (one square meter). Wind speeds observed at the entrance and inside the cave are low, approximately 10 ft/s (3 m/s) or less. Other smaller openings are known to exist where water and/or air enters the cave system.

The system consists of a central main passage called the Trunk Channel lying along the axis of a syncline, and several, fairly distinct sections of passages feeding down dip from the slopes of the syncline. Because of its large size and relatively simple geometry, the Butler Cave-Sinking Creek System provides the opportunity to study cave meteorology both near the entrance and far inside of a large cave.

The BCCS acquired an aspirated psychrometer in the early 1980s at a time when the traditional exploration and mapping of the cave system were almost complete (*see* Wefer and Nicholson, 1982) and follow-on projects were needed.

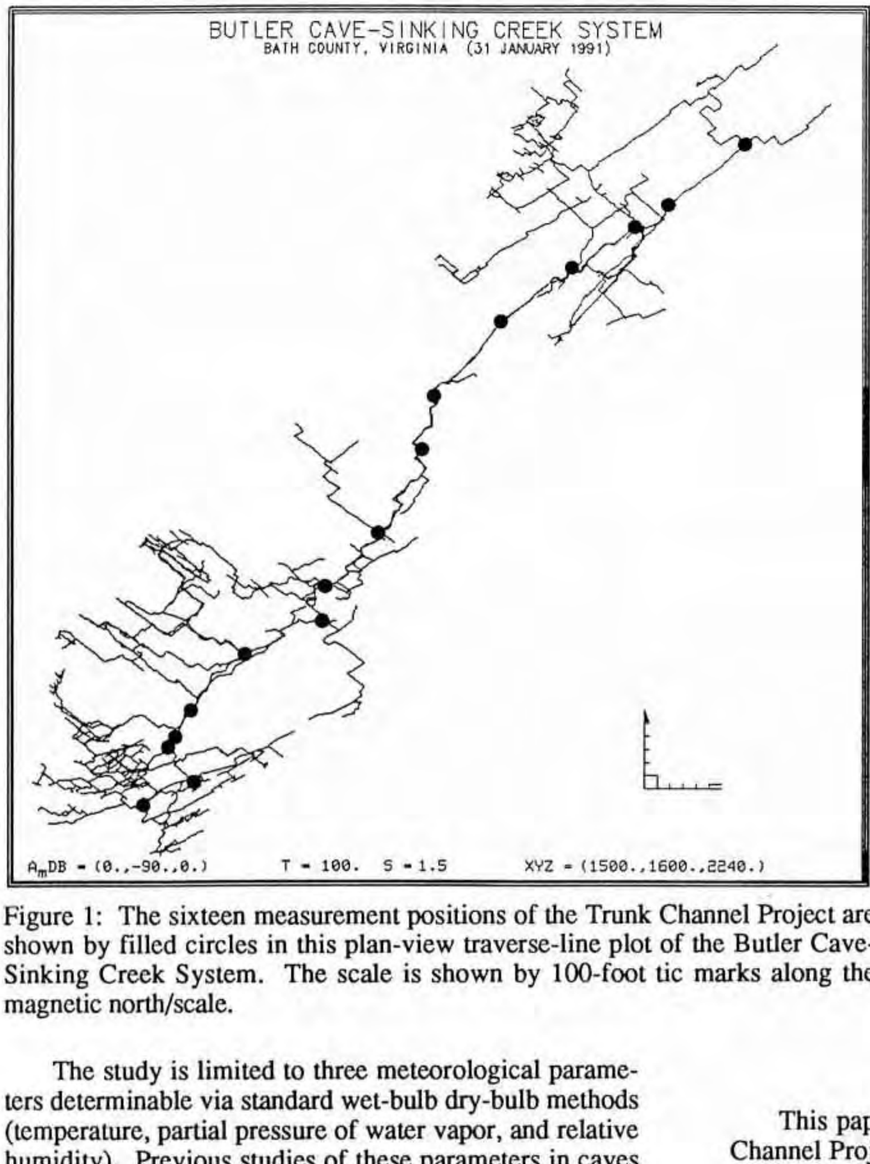

Figure 1: The sixteen measurement positions of the Trunk Channel Project are shown by filled circles in this plan-view traverse-line plot of the Butler Cave-Sinking Creek System. The scale is shown by 100-foot tic marks along the magnetic north/scale.

The study is limited to three meteorological parameters determinable via standard wet-bulb dry-bulb methods (temperature, partial pressure of water vapor, and relative humidity). Previous studies of these parameters in caves include: Davies (1960), Cropley (1965), Bamberg (1973), and Nepstad and Pisarowicz (1989), but there is little published information on meteorology of large cave systems.

Since the BCCS study was begun in April 1984, a total of 556 pairs of measurements have been made (as of 28 February 1991). The aspirated psychrometer allows the determination of the three parameters without the thermometer breakage usually attendant with the use of a sling psychrometer in a cave.

Because each parameter provides fundamentally different information, it is important to consider all three. The partial pressure of water vapor, often ignored in cave meteorological studies, is a direct indication of how much water vapor exists in the air. Relative humidity is the ratio (expressed as a percentage) of the actual partial pressure to the partial pressure at saturation, both partial pressures being determined at the measured temperature. Hence, the two parameters of temperature and partial pressure are combined into the relative humidity parameter.

During 1984 some initial exploratory measurements were made to establish that temporal and spatial differences in the three parameters could actually be seen in the cave (*see* Wefer, 1984). Also during 1984 two projects were defined, the Entrance Project and the Trunk Channel Project. Figure 2 shows a simplified schematic map of the parts of the cave system directly involved in these projects. The Entrance Project involves study of variations in the parameters between the Entrance and the Trunk Channel at Sand Canyon, a passage distance of approximately 1,800 ft (550 m). The Trunk Channel Project involves the study of variations in the parameters along approximately 8,000 ft (2,400 m) of the Trunk Channel from the upstream end at Penn State Lake to near the downstream end at the July 6th Room.

The period 1985 through 1989 saw: refinements in measurement techniques (detailed in Wefer, 1989a), continued measurements concentrating on the Entrance Project and in the Trunk Channel upstream from Sand Canyon (reported in Wefer, 1985), and calibration of the thermometers (described in Wefer, 1988). Results of the Entrance Project were presented at the 1989 NSS Convention (*see* Wefer, 1989b,c).

This paper presents preliminary results of the Trunk Channel Project in terms of temporal (seasonal) and spatial (along the Trunk Channel) variations in the three parameters: temperature (F), partial pressure of water vapor (in Hg, *i.e.*, inches of Mercury), and relative humidity (%). These are the units in which the instrument is graduated and in which the analyses are performed. Conversions to SI (metric) units have been provided for discrete numeric values in the text. In the figures below, 100 times the partial pressure of water vapor is plotted, conveniently bringing all three parameters into the same magnitude range. Whenever the term "partial pressure" is used below, the partial pressure of water vapor is meant.

Measurement Positions

Sixteen Trunk Channel measurement positions (hereafter referred to simply as positions) are shown by filled circles in Figures 1 and 2. The positions are roughly equally spaced along the passage and are all more than 1000 ft (300 m) from known air entry points. Figure 3 shows the elevations and depths of the positions plotted as

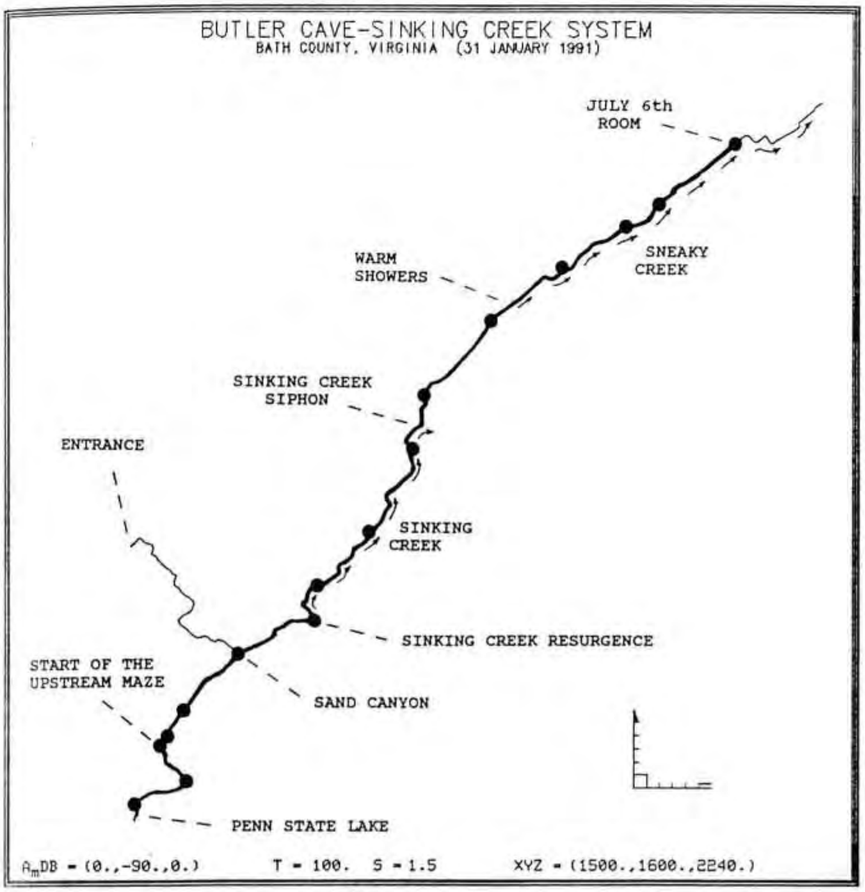

Figure 2: The parts of the Butler Cave-Sinking Creek System involved in the meteorological studies are shown in this simplified schematic map.

a function of distance along the passage from Sand Canyon. By convention, negative distances are upstream (southwest) of Sand Canyon whereas positive distances are downstream (northeast) of Sand Canyon. This upstream/downstream terminology is determined by the general direction of flow of Sinking Creek (*see* Figure 2) and is consistent with the plunge of the axis of the syncline.

The elevations of the positions are seen to decrease from approximately 2300 ft msl (feet above mean sea level) (700 m msl) at Penn State Lake, in a nearly monotonic fashion down to approximately 2000 ft msl (610 m msl) at the July 6th Room. The depths below the surface are seen to range from -150 to -300 ft (-45 to -90 m). The shallower depths downstream are due to a drop in elevation of the surface above the cave. The importance of these depths is that heat conduction effects from the surface are completely negligible since yearly (seasonal) variations as well as higher frequency (*e.g.*, diurnal) variations penetrate less than 50 ft (15 m) into the limestone (*see* Cropley, 1965).

The Butler Cave-Sinking Creek System is developed in the lower part of the 330 ft (100 m) thick Keyser Limestone. This limestone is divided into three distinct subdivisions by two sandstone units. These units are separated by about 80 ft (25 m) and are each about 13 ft (4 m) thick (*see* White and Hess, 1982 for a discussion of the geology). The lower sandstone unit forms the ceiling of the Trunk Channel from Penn State Lake to a point about 3500 ft (1070 m) downstream from Sand Canyon. Beyond that, the lower sandstone unit forms the floor of the Trunk Channel. The thick lines in Figure 3 indicate approximate locations of the two sandstone units. Thus, the five downstream-most positions are in a different layer of limestone than the rest of the Trunk Channel.

Sample Measurements

Figure 4 shows sample summer measurements made on 15 July 1989. Note the date, period of time spanned by the measurements, and initials of the caver making the measurements, all shown at the top of the plot. These eleven measurements were made during a 4.3-hour period, yielding an average total time per measurement of 24 minutes. It typically takes something like 15 to 30 minutes to pack up the instrument, travel to the next position, and set up the instrument again, plus 5 to 10 minutes to make the actual measurements, yielding an average total time per measurement in

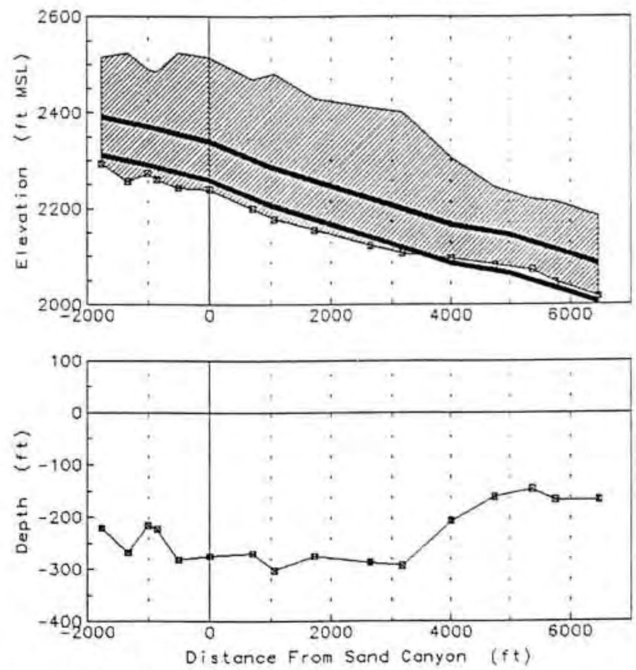

Figure 3: Elevations and depths as functions of distance along the Trunk Channel from Sand Canyon. The hatched area represents the rock layers above the cave.

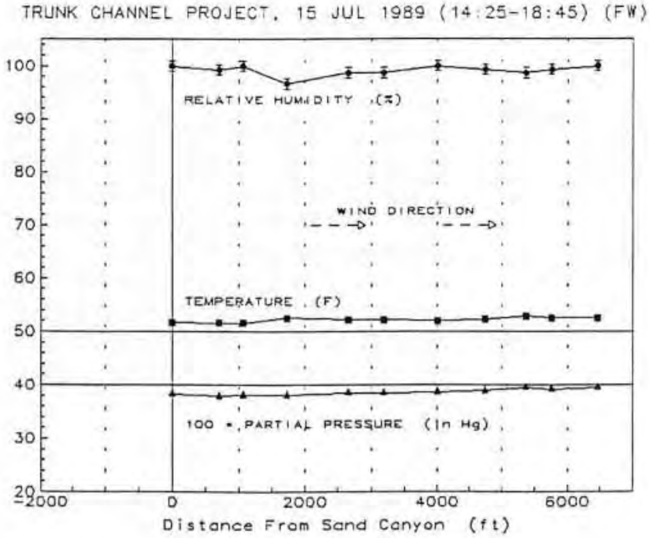

Figure 4: Sample summer meteorological measurements for the Trunk Channel.

the range of 20 to 40 minutes. The exact time between measurements depends on which positions are visited and the exact route taken through the cave, which can vary due to water conditions in the Trunk Channel. Contrast this with a total time of less than 9 minutes per measurement in the study of the tour part of Wind Cave, South Dakota by Nepstad and Pisarowicz (1989).

The temperature at each visited position is shown by a square in Figure 4. As detailed in Wefer (1989a), the wet-bulb and dry-bulb temperatures are recorded to the nearest 0.1 F. The mercury thermometers in the psychrometer are graduated in steps of 1.0 F, hence this precision is attainable only with considerable practice and skill. The uncertainty in the temperatures is thought to be about $\pm$ 0.15 F ($\pm$ 0.08 C), which is much smaller than the height of the squares in Figure 4.

The partial pressure of water vapor (actually 100 times the value) at each visited position is shown by a triangle in Figure 4. The uncertainty in the partial pressures (resulting from the above uncertainty in the wet-bulb and dry-bulb temperatures) is approximately $\pm$0.002 in Hg ($\pm$0.05 mm Hg), which is much smaller than the height of the triangles in Figure 4.

The relative humidity at each visited position is shown by a circle in Figure 4. The uncertainty in the relative humidities (resulting from the above uncertainty in the wet-bulb and dry-bulb temperatures) is approximately $\pm$ 1.0% (*see* Wefer, 1989a). This uncertainty is shown by a small error bar on each circle in Figure 4.

Straight lines connecting consecutive points of all three parameters indicate expected values in the Trunk Channel between measurement positions. Dashed arrows indicate the wind direction in the Trunk Channel.

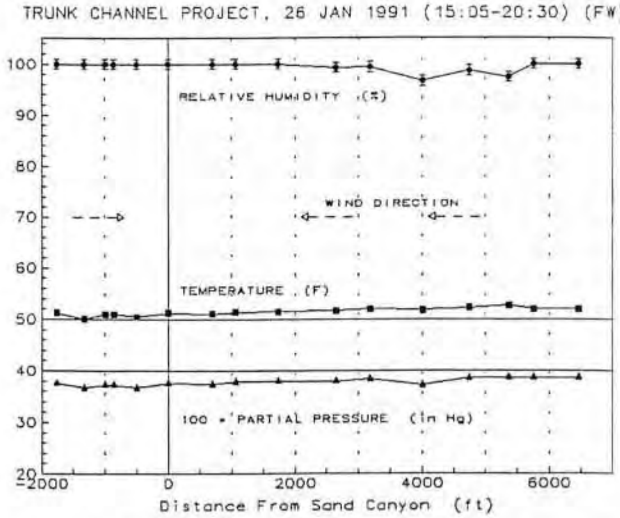

Figure 5: Sample winter meteorological measurements for the Trunk Channel.

Figure 5 shows sample winter measurements made on 26 January 1991 using the same format as in Figure 4. In these two figures all three parameters are nearly constant over the length of Trunk Channel visited on the two dates; however, variations considerably larger than the respective uncertainties are seen to exist. Note that both the temperature and partial pressure gradually increase as one proceeds downstream. Because these systematic variations are slight, horizontal lines have been drawn at ordinate values of 40 and 50 to make them easier to see.

Spatial Variations

Plots such as Figures 4 and 5 suffice to show that spatial variations exist in the three parameters, even on a single date. Of equal interest are trends in the parameters averaged over time. One way of displaying such trends is via a composite plot (Figure 6). On a composite plot, all curves (*i.e.*, data sets for which measurements are available at several consecutive positions along the Trunk Channel on a given date) for a single parameter are super-

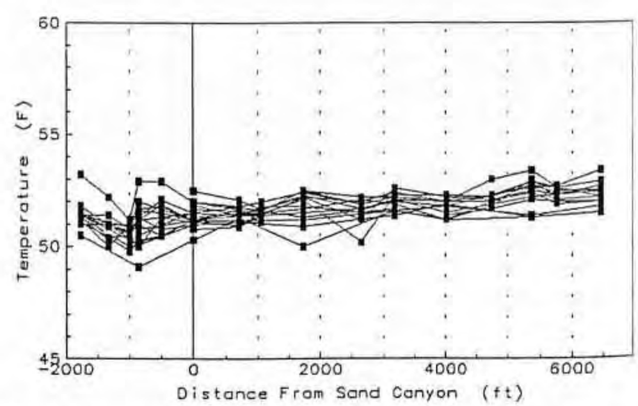

Figure 6: Composite plot of temperature variations along the Trunk Channel.

imposed. Because only one parameter is being plotted, a greatly expanded ordinate scale is practical.

This simple technique has been found to work well, particularly when curves are available throughout the year, as is the case in this study. Composite plots have apparently been previously used only in studies of the variations of the parameters with distance from the cave entrance (*see for example*, Gaum, 1952; Benedict, 1974a,b; Wefer, 1989c).

Another method of detecting trends in the parameters is to plot histograms of the distributions of the measured values (Figure 7). Ideally one would plot a histogram for each position; however, at this stage in the Trunk Channel Project only the position at Sand Canyon has a sufficient number of measurements to provide a smooth distribution. The Sand Canyon position has many more measurements than the other positions for two reasons - as part of the Entrance Project it was visited even when no other positions in the Trunk Channel were visited and it was normally visited multiple times on a single date.

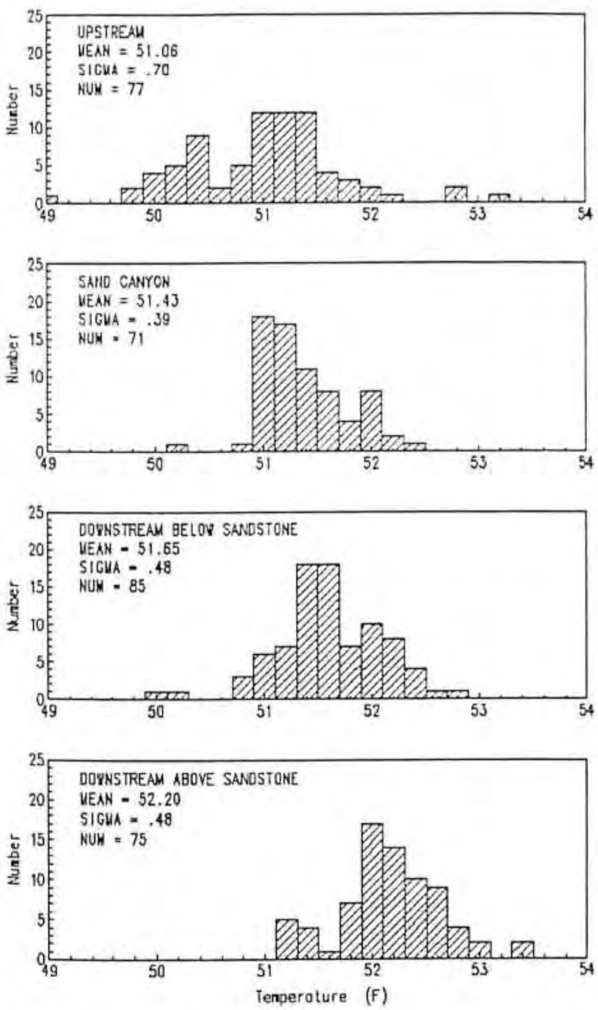

Figure 7: Distributions and statistics for all temperature measurements in the Trunk Channel.

Accordingly, data from the sixteen positions has been combined into four groups as follows: the five upstream positions, the position at Sand Canyon, the five downstream positions located below the lower sandstone unit (*see* Figure 3), and the five downstream positions located above the lower sandstone unit.

Whereas the composite plots show only data where consecutive measurements are available at a number of positions on each date, the distributions make use of all available data, even isolated measurements. The mean value (MEAN), standard deviation (SIGMA) and the number of measurements (NUM) in each group are listed in the figures.

Spatial Variations in Temperature

Figure 6 shows a composite plot of temperatures in the Trunk Channel. Because of the expanded ordinate scale (compared with Figures 4 and 5), the uncertainty in the temperature measurements is here equal to the half-height of the squares. The temperature in the Trunk Channel is seen to vary between approximately 49.0 and 53.5 F (9.4 and 11.9 C). Figure 7 shows the distributions of temperature measurements and the statistics of these distributions.

A major trend in both figures is the gradual increase in temperature as one proceeds downstream. Whereas the increase is only from approximately 51.1 to 52.2 F (10.6 to 11.2 C), it is, nevertheless, obvious in both figures. Possible origins of this increase are discussed in a separate section below.

A minor feature apparent in Figure 6 is systematically lower temperatures at the central upstream position located at the Start Of The Upstream Maze. The highest temperature recorded there is actually lower than the average temperature at Sand Canyon. Measurements at this position are largely responsible for the bimodality of the upstream distribution in Figure 7.

Two possible explanations for this minor feature have been considered. Firstly, it might have been caused by incomplete temporal coverage in the data (*e.g.*, the measurements might simply have been concentrated in the winter months); however, the measurements are, in fact, well distributed throughout the year. Secondly, it might have been caused by the position being located at a low point in the passage where a cold air trap could form; however, Figure 3 shows that the position is actually at a high point in the Trunk Channel. The actual cause of this minor feature remains unknown.

Spatial Variations in Partial Pressure

Figure 8 shows a composite plot of partial pressures in the Trunk Channel. The uncertainty in the measurements is shown by error bars on the triangles. Partial pressure in the Trunk Channel is seen to vary between

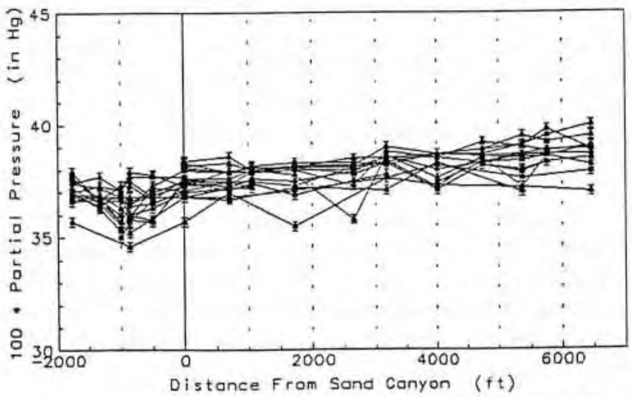

Figure 8: Composite plot of partial-pressure variations along the Trunk Channel.

approximately 0.345 and 0.400 in Hg (8.76 and 10.16 mm Hg). Figure 9 shows distribution of the partial-pressure measurements and statistics of these distributions.

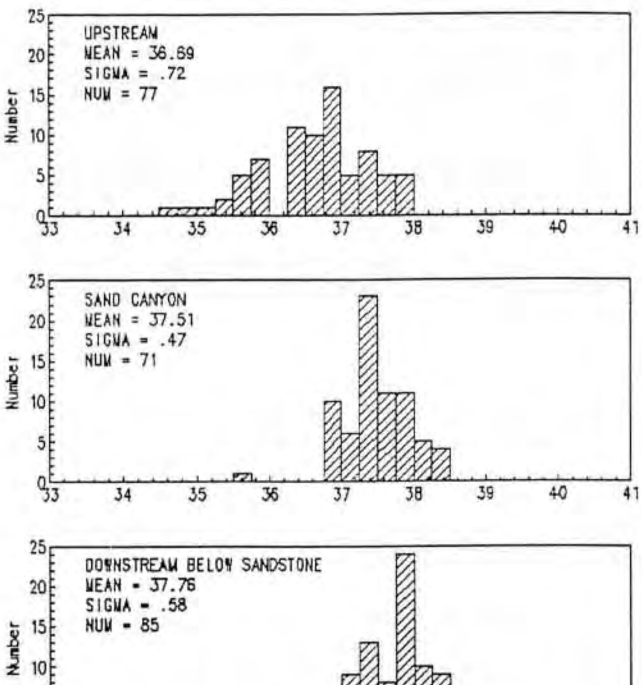

Figure 9: Distributions and statistics for all partial-pressure measurements in the Trunk Channel.

A major trend in both figures is the gradual increase in partial pressure as one proceeds downstream. Whereas the increase is only from approximately 0.367 to 0.385 in Hg (9.32 to 9.78 mm Hg), it is, nevertheless, obvious in both figures. Possible origins of this increase are discussed in a separate section below.

A minor feature apparent in Figure 8 is systematically lower partial pressures at the Start Of The Upstream Maze. Measurements at this position are largely responsible for the bimodality of the upstream distribution in Figure 9. The partial pressure is also unusually low at the adjacent position just downstream from the Start Of The Upstream Maze, an effect likely caused simply by the dry air flowing down the passage in that direction. The lower partial pressures probably share the same origin as the lower temperatures discussed above; however, that origin remains unknown.

Spatial Variations in Relative Humidity

Figure 10 shows a composite plot of relative humidities in the Trunk Channel. The uncertainty in the measurements is shown by error bars on the circles. Relative humidity in the Trunk Channel is seen to vary between approximately 92 and 100%. Figure 11 shows distribution of the relative-humidity measurements and statistics of these distributions.

A major feature in both figures is that the vast majority of relative-humidity values are greater than 97%, *i.e.*, the relative humidity is greater than 97% at most places and at most times in the Trunk Channel.

Two minor features are apparent in Figure 10. An unusually low relative humidity of approximately 92% was measured at Penn State Lake. This was during winter when cold, dry air enters the cave system from nearby Boundless Cave, flows down the Trunk Channel to Sand Canyon, then out the higher Butler Entrance.

The variation in relative humidity is approximately 3% at most positions; however, several show variations

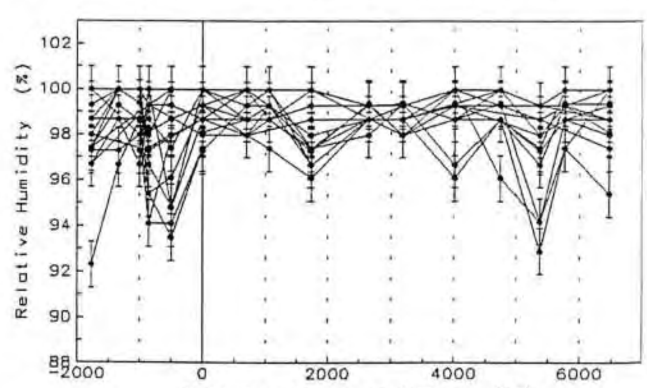

Figure 10: Composite plot of relative-humidity variations along the Trunk Channel.

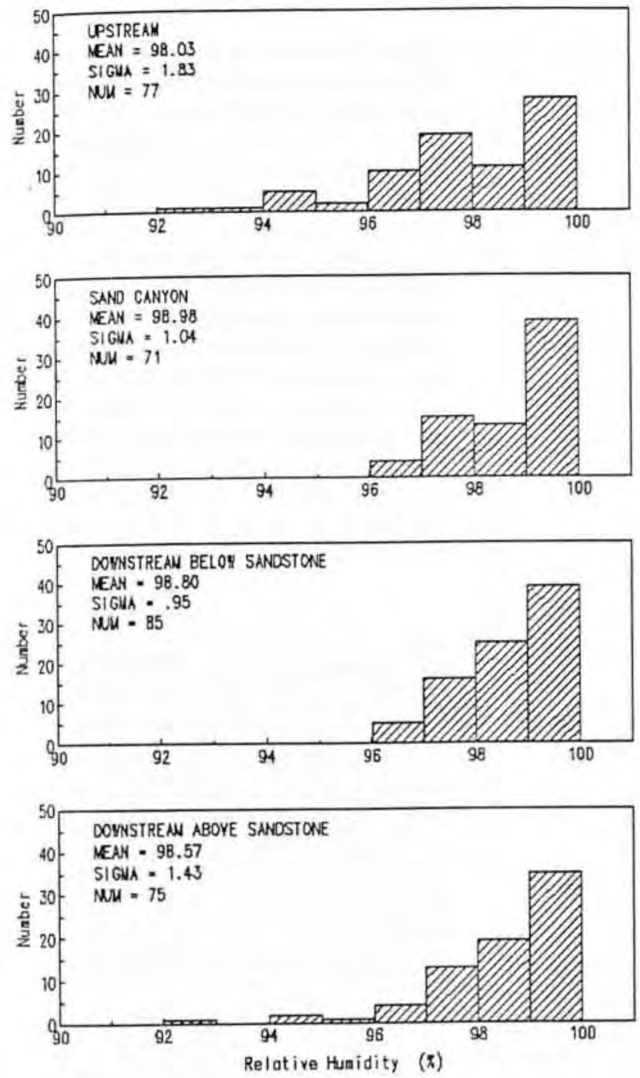

Figure 11: Distributions and statistics for all relative-humidity measurements in the Trunk Channel.

as high as 6% or larger. These positions are all at points where major side passages intersect the Trunk Channel, hence the variability could be caused simply by dry air flowing out of these sections of the cave. Note, however, that not all such positions show high variability. Sand Canyon (a major passage intersection) displays a low variability.

The two positions just upstream from Sand Canyon often exhibit low relative humidity. Both are near passage intersections, but far from known surface air-entry points. The low relative humidity tail of the upstream distribution in Figure 11 results largely from measurements at these two positions. Why some sections of the cave sometimes have lower relative humidities is unknown.

Temporal Variations

Temporal variations in meteorological parameters near the entrances of caves have been observed and studied (*see for example*, Gaum, 1952; Cropley, 1965; Benedict, 1974a,b; Wigley and Brown, 1976; Wefer, 1989b,c). One of the goals of the Trunk Channel Project is to determine whether systematic temporal variations can be observed far inside a cave, as contrasted with near the entrance. This involves looking for parameter variations in time at each position along the Trunk Channel.

As has been shown above, the observed variations tend to be small. For example, the distributions of Figures 7, 9, and 11 show that meteorological conditions at Sand Canyon are remarkably constant, with means and standard deviations as follows: temperature 51.4 ± 0.4 F, partial pressure 0.375 ± 0.005 in Hg, and relative humidity $99.0 \pm 1.0\%$. Hence a large number of measurements are required at each position in order to separate systematic variations from random statistical fluctuations.

Figure 12 shows the seventy-one measurements currently available (as of 28 February 1991) at the most frequently visited position, Sand Canyon. Most positions have less than a third of this number of points. In this figure the abscissa is the day number in the year, the meteorological parameters are indicated by the previously used symbols, and the average values are shown by horizontal straight lines across the plot.

Concentrating on temperature, which possesses the most obvious seasonal variation on the surface, it is apparent that measurements made during the winter tend to be below average, whereas those made during the summer tend to be above average. If one were to forge ahead and fit the data with a sinusoidal curve, the resulting amplitude would be only a fraction of a degree and the phase would not be well determined.

Ives (1964) and Sullivan and Moore (1964) presented evidence of phase lags in seasonal-temperature variations inside (but near the entrances of) caves compared with those on the surface. Although there appears to be no theoretical reason to suspect that appreciable phase shifts exist far inside caves, whether or not they do can be determined only when many more measurements are available.

Systematic Spatial Variations

As has been shown above, both temperature and partial pressure are seen to increase as one proceeds downstream in the Trunk Channel. There is, however, little reason to believe that distance from Sand Canyon is a fundamental independent variable in this process. As was shown in Figure 3, elevations of the positions decrease as one proceeds downstream. Relationships we see in Figures 6 through 9 may originate from decreasing elevations of the positions.

Figure 13 shows the temperature measurements plotted versus elevation. Points from the warmer months (day numbers 109 through 291) are plotted with "+" symbols,

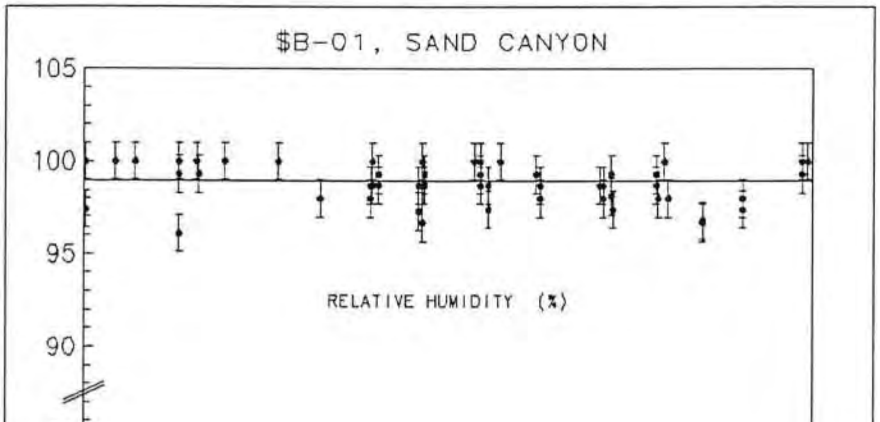

Figure 12: Temperature, partial pressure, and relative humidity as functions of day of the year for seventy-one measurements at Sand Canyon.

points from the colder months with "x" symbols. There appears to be a tendency for the "+" symbols to be at higher temperatures than the "x" symbols at the same position. Because the measurements are not evenly distributed throughout the year at all positions, only a single least squares fit has been made to the data. The resulting slope (SLOPE), standard deviation of the slope (SIGSL), correlation coefficient (CORR), and number of measurements involved (NUM) are shown on the plot.

One could, of course, prepare a plot similar to Figure 13 for the partial pressure. Recall, however, that partial pressure at saturation is a function of temperature only. Water vapor is available in sufficient quantities from Sinking Creek, Sneaky Creek, and the cave walls to keep the relative humidity in the Trunk Channel at essentially 100%, *i.e.*, at saturation. Thus the partial-pressure rate is driven by the temperature rate. A simple calculation shows that the observed partial-pressure rate of Figure 8 is consistent with the observed temperature rate of Figure 13, hence we need only be concerned with temperature observations.

As cave air moves along the Trunk Channel, the changing elevation subjects it to changing pressure. In an adiabatic process (*i.e.*, no heat enters or leaves the system) the rate of change of temperature with elevation is constant (over the limited range of temperature and elevation in the cave). This so called "adiabatic lapse rate" is approximately -5.4 F/1,000 ft (-9.8 C/1,000 m) in the Butler Cave-Sinking Creek System.

In the lower atmosphere of the earth, the observed "atmospheric lapse rate" is approximately -3.5 F/1,000 ft (-6.4 C/1,000 m). The atmospheric lapse rate is lower than the adiabatic lapse rate primarily because of the effects of water changing phase. Decreasing temperature with increasing altitude causes the relative humidity to increase and water vapor to condense into clouds. The heat of vaporization released in this process increases the local temperature, having the effect of decreasing the lapse rate.

In the Trunk Channel the relative humidity is nearly 100%, the walls are wet, streams cover much of the floor, and air, walls, and water are at nearly the same temperature. The net amount of water changing phase must be small, hence we expect the Trunk Channel lapse rate to be close to adiabatic, as indeed it is.

Although it is interesting that the measured lapse rate is close to adiabatic, the reader is reminded that these results are preliminary and that adiabatic compression is not the only process at work in the Trunk Channel. Possible sources (or sinks) of heat (that would make the process non-adiabatic) include heat from air and water entering (and exiting) the Trunk Channel, as well as heat from the geothermal gradient. Water-temperature measurements indicate, however, that neither Sinking Creek nor Sneaky Creek is a thermal stream (despite the intriguing name of the source of the latter).

Errors in the measurements must also be considered. Possible sources of errors include: errors in elevations of the positions, errors due to heat from the observer's body, and errors due to heat generated inside the psychrometer.

Because these sources are currently under review, their further discussion is beyond the intended scope of this paper.

The most serious consideration is that measurements at many of the far-downstream positions are not well distributed in the year. When this situation is rectified, it will make sense to compute separate least squares fits for the summer and winter measurements. This is expected to yield higher correlation coefficients and better values for the Trunk Channel lapse rate. Many more measurements are required before firm conclusions can be drawn about its numerical value and its meaning.

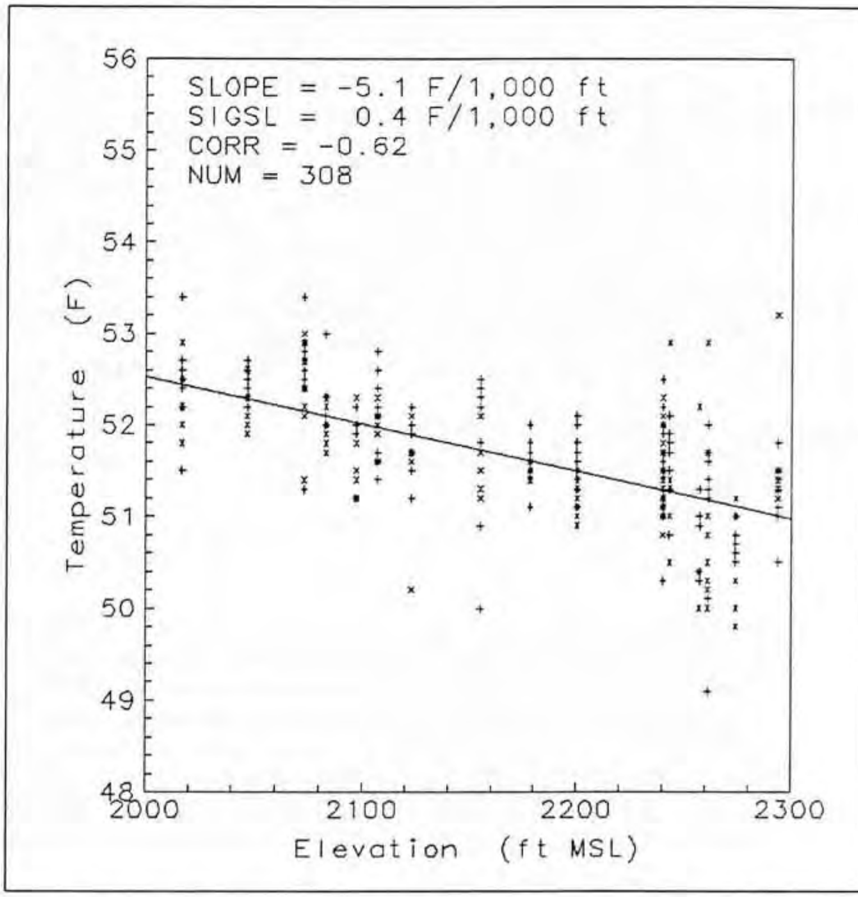

Figure 13: Temperature as a function of elevation for all temperature measurements in the Trunk Channel. Points from the warmer months are plotted with "+"s, points from the colder months with "x"s. A least squares fit to the data is also shown.

Conclusions

Temperature, partial pressure, and relative humidity are all nearly constant in the Trunk Channel far inside the Butler Cave-Sinking Creek System; however, systematic spatial and temporal variations do exist and are observable. That a systematic change in temperature with elevation is observed indicates that a very high precision has been attained in the measurements. More data is needed before a detailed explanation of this phenomenon can be advanced. Temporal variations far inside the cave are quite small, but appear to affect the interpretation of the spatial variations.

Perhaps the most important conclusion to be drawn from this preliminary report is that interesting and useful meteorological work can be performed by cavers with relatively simple and inexpensive instruments. Much more work remains to be done in this area of speleology.

References

Bamberg, S.A., 1973, Environment in Lehman Caves, Nevada, *National Speleological Society Bulletin*, v. 35, p. 35-47.

Benedict, E., 1974a, Winter temperatures and air flow, Malheur Cave, *Speleo Digest-1974*, p. 214-215.

Benedict, E., 1974b, Air temperatures in Owyhee River Cave, 1974, *Speleo Digest-1974*, p. 217.

Cropley, J.B., 1965, Influence of surface conditions on temperatures in large cave systems, *National Speleological Society Bulletin*, v. 27, p. 1-10.

Davies, W.E., 1960, Meteorological observations in Martens Cave, West Virginia, *National Speleological Society Bulletin*, v. 22, p. 92-108.

Gaum, C.H., 1952, Hydrologic and atmospheric studies in Schofer Cave, *National Speleological Society Bulletin*, v. 14, p. 49-57.

Ives, J.D., 1964, Cave migration of certain insects, *National Speleological Society Bulletin*, v. 26, p. 115-118.

Nepstad, J. and Pisarowicz, J., 1989, Wind Cave, South Dakota: Temperature and humidity variations, *National Speleological Society Bulletin*, v. 51, p. 125-128.

Sullivan, G.N. and Moore, G.W., 1964, Out of phase seasonal temperature fluctuations in Cathedral Cave, Kentucky (abstract), *National Speleological Society Bulletin*, v. 26, p. 84.

Wefer, F.L. and Nicholson, I.K., 1982, The exploration and mapping of the Sinking Creek System, *National Speleological Society Bulletin*, v. 44, p. 48-63.

Wefer, F.L., 1984, Cave meteorology, *BCCS Newsletter*, v. 10, p. 33-49.

Wefer, F.L., 1985, More on cave meteorology, *BCCS Newsletter*, v. 11, p. 25-56.

Wefer, F.L., 1988, Still more on cave meteorology, *BCCS Newsletter*, v. 14, p. 76-84.

Wefer, F.L., 1989a, On the measurement of relative humidity in cave meteorology projects, *Nittany Grotto News*, v. 36, p. 6-14; reprinted in *Speleo Digest 1989*, p. 395-399.

Wefer, F.L., 1989b, The meteorology of the Butler Cave-Sinking Creek System (abstract), *in* Eck, C. and Olsen, L. (editors), *1989 NSS Convention Program, 31 July - 4 August 1989, Sewanee, Tennessee*: National Speleological Society, Huntsville, Alabama, p. 49.

Wefer, F.L., 1989c, Yet still more on cave meteorology, *BCCS Newsletter*, v. 15, p. 18-38.

White, W.B. and Hess, J.W., 1982, Geomorphology of Burnsville Cove and the geology of the Butler Cave-Sinking Creek System, *National Speleological Society Bulletin*, v. 44, p. 67-77.

Wigley, T.M.L. and Brown, M.C., 1976, The physics of caves (Chapter 9), *in* Ford, T.D. and Cullingford, C.H.D. (editors), *The Science of Speleology*: Academic Press, London, p. 329-358.

Emerged Sea Caves and Coastal Features as Evidence of Glacio-Isostatic Rebound, Mount Desert Island, Maine

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ABSTRACT

Acadia National Park is situated on Mount Desert Island, Maine. A number of actively forming sea caves are present within the park. Two similarly appearing granite caves, one on Gorham Mountain and one on Champlain Mountain, have been identified at elevations well above the present sea level. An investigation was conducted to determine if these caves developed at a former sea level and were in fact sea caves. Previous studies had drawn conflicting conclusions as to whether the Gorham Mountain cave was indeed a sea cave, whereas the cave on Champlain Mountain was apparently not identified in the literature. A third feature, a pinnacled rock referred to as Pulpit Rock, situated on Day Mountain at the base of a cliff, resembles sea stacks of current-day rocky coasts. Limited debate is present in historic literature as to its origin.

A water-tube leveling survey was conducted to determine elevations of the two emerged caves. Cadillac Cliffs Sea Cave, on Gorham Mountain, occurs at an elevation of 231 to 238 ft msl. Champlain Mountain Sea Cave occurs at an elevation of 218 to 227 ft msl. Pulpit Rock extends from 204 to 230 ft msl. Thompson and others (1989) documented the maximum isostatic crustal uplift following deglaciation of Mount Desert Island as 231 ft msl, based on an ice-contact glaciomarine delta. This figure represents maximum glacio-isostatic rebound following the late-Wisconsinan retreat of the Laurentide Ice Sheet (~ 14 to 12 ka).

The limits of emerged boulder beaches on Day and Gorham Mountains were definable through additional survey work. The bottom and top of these beaches lie at elevations of approximately 204 to 245, and 173 to 198 ft msl, respectively. It had been assumed that these emerged boulder beaches were at approximately the same elevation. Instead, a continuum of glacio-isostatic rebound is portrayed as the depressed land surface rose following glacial unloading. The two emerged sea caves, the Pulpit Rock sea stack, the emerged boulder beaches, and nearby wave-cut platforms represent part of the geomorphic reconstruction. In addition, a revised maximum postglacial rebound figure is presented for Mount Desert Island.

Location

Mount Desert Island is located off the coast of Maine 110 miles northeast of Portland, Maine and 130 miles southwest of Saint John, New Brunswick, Canada. It is the largest and highest of many islands in this area. Acadia National Park, situated on Mount Desert Island, has long been a favorite haunt of tourists and geologists.

The Upper Marine Limit

As the Laurentide Ice Sheet retreated from Mount Desert Island and the Maine seaboard, the depressed land mass rebounded rapidly (Stuiver and Borns, 1975). Belknap and others (1987) found that glacio-isostatic uplift (land rise due to glacial unloading) occurred rapidly during deglaciation, continuing asymptotically thereafter. Even though sea level at the close of the Wisconsinan was lower than today's, sea level relative to the depressed land surface was considerably higher. As Raisz (1929) aptly states:

> "The height of this 'Upper Marine Limit' is one of the...outstanding problems of Mount Desert Island...The interest of...the question lies not only in its bearing upon the general...geological history of the

region, but also in the fact that...nearly all geologists who have studied the Island have ...expressed different opinions on the subject."

Shaler (1889) and Bascom (1919) identified numerous terraces on Mount Desert Island that they believed to be of marine origin. These terraces are reportedly found at various elevations up to and above 1000 feet, with a marine origin generally not given credence by geologists today. Stone (1899) placed the upper marine limit at 230 feet, while Fairchild (1919) placed it at 250-260 feet. Johnson (1925) questioned the marine origin of the higher cliffs of Mount Desert Island, including the Cadillac Cliffs. Other authors (*e.g.*, Smith, 1966, and Thompson and others, 1989) have conducted extensive studies that placed the upper marine limit for coastal Maine more in line with the work of Stone (1899) and Fairchild (1919). Many of these authors used an altimeter or barometer in order to determine the altitude of various features.

In order to provide a reasonable interpretation of the elevation of the upper marine limit, it is important to examine as many lines of evidence as possible, that together correlate the available data and support a unified theory. These data include sea caves, boulder beaches, sea stacks, benches, and marine deposits. It is necessary to understand the processes involved in the formation or deposition of these features. This paper attempts to correlate recent, still actively forming, coastal features with emerged coastal features.

Leveling Procedure

In order to assess and interpret various emerged coastal features, it was necessary to accurately determine their elevation. A systematic leveling survey was conducted utilizing a water-filled U-tube. A water-tube leveling survey is one means of rapidly determining elevational differences between locations with very small errors. The mean vertical error in loops for water tube surveys is 0.003 percent (Palmer, 1987). All survey elevations were determined to the nearest 0.01 foot. Accuracy is certainly valid to the nearest 0.1 foot, although elevations are reported here to the nearest foot. A U-tube is particularly useful in hilly, heavily vegetated, and very rough terrains.

A U-tube can be used very efficiently by two people, as it requires no timely or complicated setup procedures. Colored water at raised ends of the tube rests at an equal elevation. The range of the instrument utilized was 5.8 feet vertically and 50 feet horizontally. During the course of the survey a small number of permanent stations (such as the metal rung of a ladder along a trail) were established, separated by many temporary stations. Temporary stations consist of two limestone slabs, approximately seven inches in diameter, each with a drilled, 1/4-inch-deep by 3/4-inch-long groove in their center. Temporary stations were set by either firmly stomping a drilled rock into the ground or by placing it on a bedrock or other solid surface in such a manner as to guarantee no vertical or horizontal displacement.

Wooden rulers were placed in each of the drilled-rock grooves and manually clamped to each end of the U-tube. After removing a threaded metal cap from both ends of the U-tube, the fluid level in the tube was allowed to stabilize and the level above each station recorded. The difference in elevation between the two stations is the gain or loss in topographic elevation along the traverse. This operation was repeated twice, or if necessary more times between stations until a maximum difference of 0.01 feet between readings was achieved. A number of calculations were performed after each measurement was made in order to detect any errors. Periodically, the two ends of the U-tube were brought together to verify that the fluid levels were equal, thus insuring that no air bubbles had been introduced into the U-tube. Occasional placement of permanent stations reduces repeat surveying should air bubbles be detected. The survey continued in a step wise manner until the level of each desired survey point was established.

Boulder Beaches

For ease of the upcoming discussions, boulder beaches are discussed prior to sea caves. However, it was this author's firm belief that the Gorham Mountain cave was an emerged sea cave that led to the leveling survey of the emerged Day Mountain boulder beach.

Well-developed boulder beaches are often found at the base of steep coastal cliffs and cliff-bounded coves, where abundant source material is present. Rocks falling from these cliffs are constantly abraded, rounded, and polished through continuous wave action. A relatively shallow foreland gradient promotes boulder-beach development. Storm surges further round boulders through mechanical corrasion, often hurling boulders against a shore to elevations higher than the high tide mark. A period of only tens of years is needed to round quarried, rectangular blocks of roadbed materials, such as those with drill holes in Monument Cove.

Numerous examples of modern boulder beaches can be found along the coast of Mount Desert Island. The Monument Cove area exhibits many fine examples. Protected rock-wall bounded coves often host laterally continuous boulders across their shorelines. One example had a maximum vertical extent of approximately 35 feet, again reinforcing the concept that boulder beaches may represent a greater vertical extent than the difference between the high- and low-tidal range.

Three emerged boulder beach areas were examined. One excellent example lies on Gorham Mountain, slightly lower elevationally than the Cadillac Cliffs trail, directly east of a small trailside cave referred to here as Cadillac Cliffs Sea Cave. Coffin and others (1990) reference this

cobble- and boulder-beach deposit as a 75-foot-square area. Further examination revealed an extensive area of cobble and beach boulders in what was once a small cliff-bounded cove, similar in setting to many coastal coves of today. Shipp and others (1985) classify this type of geomorphic feature (*e.g.*, a small cove) as a pocket beach. Through careful observation of the heavily wooded beach, it was possible to delineate an upper and lower boundary. Boundary selection criteria included the presence of at least one cluster of clearly rounded and polished boulders, physically within close proximity of other more massive concentrations of boulders. The boulder beach top and bottom were determined to be at 198 and 173 feet msl, respectively. The boulder beach, perpendicular to the strandline, encompasses 160 feet, having a shore-normal gradient of 0.15. The lateral extent of the beach boulders, between nearby rock cliffs, is approximately 225 feet. This emerged cobble- and boulder-beach, complete with a sea cave in its upslope wave-cut cliff, may be the only known example anywhere of a former high-wave-energy pocket beach.

The best example of a high-energy emerged shoreline that is known in New England is found on Day Mountain below The Cleft (Coffin and others, 1990). Coffin and others cite this former beach as being nearly 1200 feet in lateral extent. Stuiver and Borns (1975) state that the beach developed in less than 600 years. The upper boundary was surveyed at 245 feet msl and the lower boundary at 204 feet msl. The survey of the Day Mountain boulder beach, Pulpit Rock, and a third boulder beach at the base of Pulpit Rock was tied to a U.S.G.S. benchmark along Route 3.

Actively Forming Sea Caves

A number of actively forming sea caves are present within the Park. Three of these caves were mapped with a Suunto compass and calibrated tape. Anemone Cave and Great Head Cave (Figures 1 and 2), located near the northwestern terminus of Oak Hill Cliff and at the base of the westerly facing sea cliffs of Great Head respectively, are enlarging at the present mean sea level. Stag Cave (Figure 3), located along the southeasterly facing sea cliff of Great Head, is also still forming.

The development of Anemone and Great Head caves along the bases of their respective cliffs is not simply a matter of chance. Both have formed as the result of constant wave action against highly fractured and intensely shattered country rock. Although slickensides were not observed, both caves appear to lie along low-angle thrust faults. More massive or less fractured bedrock is present toward the base of the respective caves. The undulating ceiling of Great Head Cave, possibly exhibiting well eroded slickensides, further suggests localized thrusting. The highly fractured bedrock, whether faulted or not, has provided a preferential pathway for attack by surging ocean waters. Close fracture spacing in both caves, and vein filling in Anemone Cave, permit the continued undercutting, erosion, and removal of relatively small blocks of rock. The ceiling of Anemone Cave supports active cliff swallow nests, suggesting that only storm tides result in complete inundation of the cave.

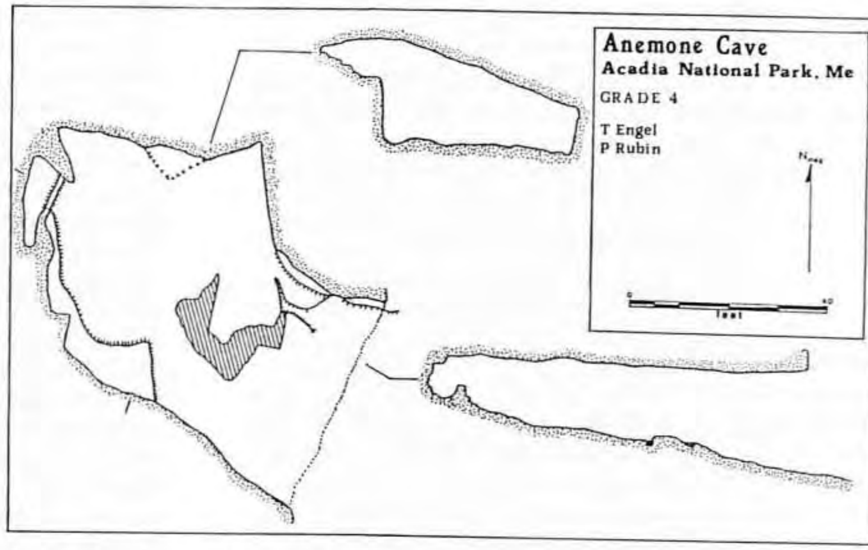

Figure 1: Map of Anemone Cave.

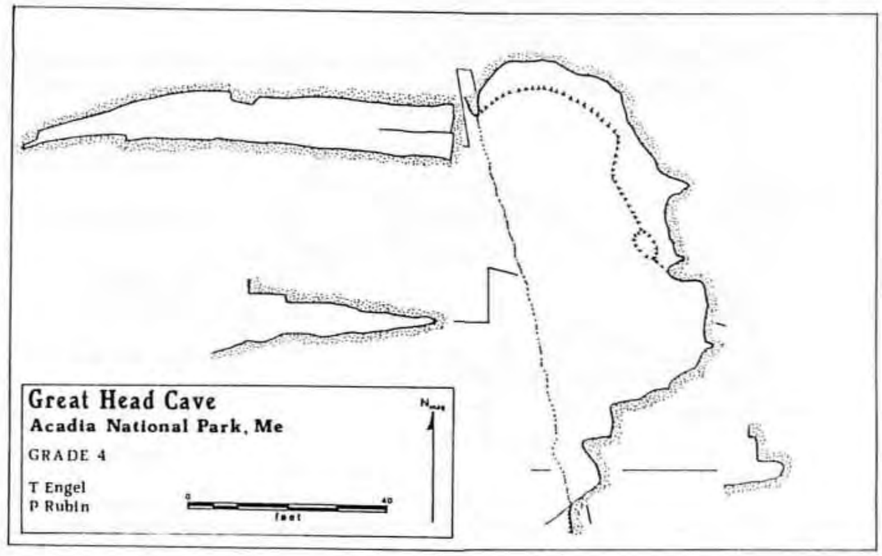

Figure 2: Map of Great Head Cave.

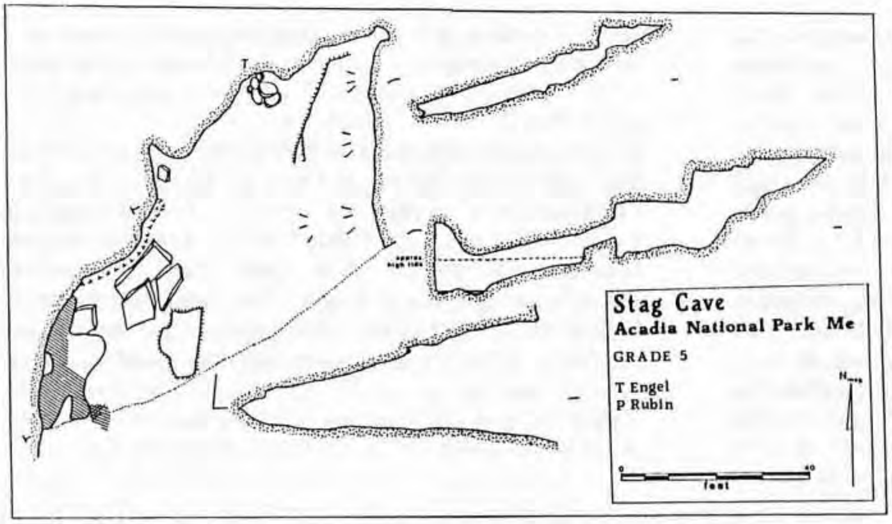

Figure 3: Map of Stag Cave.

Stag Cave has developed preferentially along a thrust fault. Extension veins, slickensides, and fracturing are present. Hundreds, if not thousands, of years of wave action and quarrying have produced the caves we see today.

Emerged Sea Caves

Cadillac Cliffs Sea Cave (Figure 4), located near the base of an easterly facing cliff on Gorham Mountain, is interpreted as representing evidence of large scale glacio-isostatic rebound following Wisconsinan glaciation. This cave is found along the base of the Cadillac Cliffs trail approach to the summit of Gorham Mountain. A leveling survey determined the elevation of the cave's floor to be 234 feet msl, with a ceiling elevation of 238 feet msl. The seaward-sloping floor at the entrance to the cave is 231 feet msl at a point 6 feet southeast of the dripline. The most reasonable location to initiate the leveling survey to this cave and the nearby boulder beach was from the summit of Gorham Mountain. The Seal Harbor 7.5-minute Quadrangle cites this as 522 feet msl. This is believed to be accurate to within one tenth of the published contour interval (± 2 feet).

The cave is formed in a relatively massive pink granite. The resistant nature of the bedrock is apparent as the cave is situated near the base of a high cliff. Granite is not a rock type typically associated with cave development because its component minerals are hard and weakly soluble. The physical presence of the cave so far above today's sea level has led geologists to speculate on whether it is an ancient sea cave. Johnson (1925) doubted that the Cadillac Cliffs were of marine origin. Bascom (1919) and Smith (1966) believed the cave to be of marine origin, whereas Johnson (1925) and Raisz (1929) found otherwise.

In addition to the fact that Cadillac Cliffs Sea Cave falls well within the former sea level suggested by the Day Mountain boulder beach, examination of various other lines of geological evidence also suggest that it is an ancient sea cave. The relatively short cave abruptly ends in a solid bedrock wall with no large avenues for water infiltration that could have led to cave development. A few small, isolated openings in the ceiling and behind a zone of rounded granite boulders appear to be incapable of providing an avenue for sufficient water discharge through the cave to account for its development. The only likely source of water capable of infiltrating downward to these openings is from very localized surface runoff through a large, vertical fracture a short distance west of the cave mouth. This vertical fracture extends downward from the top of Cadillac Cliffs, but does not appear to directly itersect the cave. In addition, the potential recharge area to

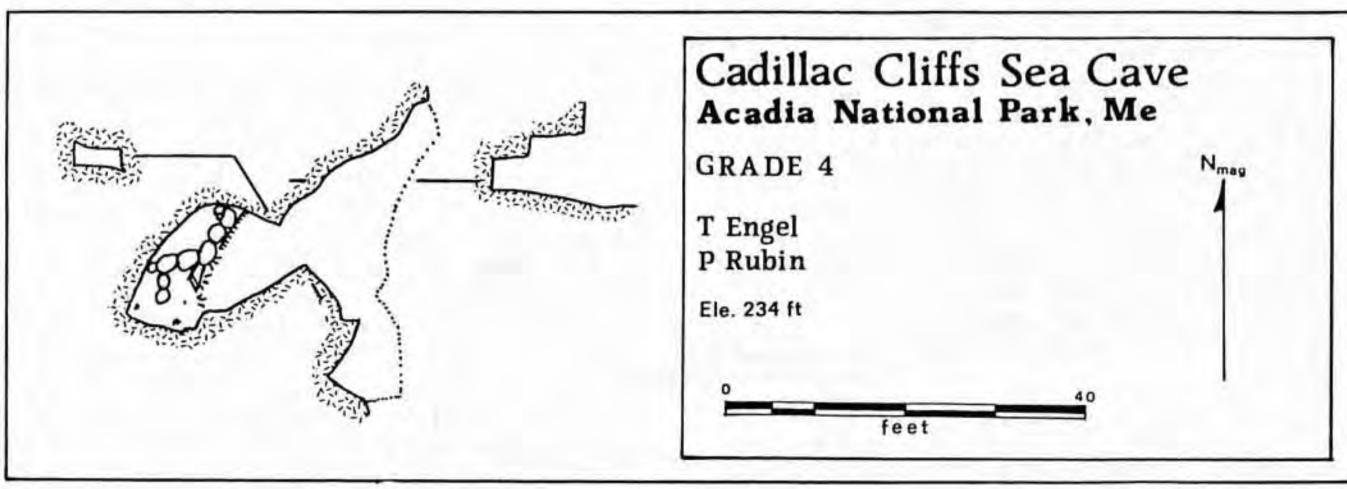

Figure 4: Map of Cadillac Cliffs Sea Cave.

this vertical fracture or to any other nearby fractures is extremely limited in areal extent because the fractures occur virtually at the top of Gorham Mountain's summit ridge.

Cadillac Cliffs Sea Cave's high elevation above sea level might, at first glance, suggest that it formed prior to the Wisconsinan glaciation. Perhaps the surface topography comprising the cave's potential recharge area was considerably larger prior to glacial abrasion. Kasycki and Shilts (1979) estimate Canadian shield erosion rates to be as low as 33 feet for each major glaciation. Gilman and others (1988) estimate that abrasion on Mount Desert Island may have lowered the average bedrock surface by one or two yards during the last glaciation. Thus, it is unlikely that Wisconsinan glacial abrasion or plucking removed a significant part of a formerly larger recharge area.

The physical characteristics of the cave itself do not support a fracture-infiltration/vadose-stream formation mode. Had a large enough recharge area been available to support an intermittent stream to the cave, the small-size opening in the ceiling would be expected to more closely conform to the cross-sectional dimensions of the cave. Instead, the cave and the cave's mouth are laterally wide and lack any evidence of channel incision. The limited recharge potential to the cave, coupled with the cave's blind terminus and lack of any incised floor channel, suggests a formation mode other than vadose (unsaturated zone) groundwater discharge.

Cadillac Cliffs Sea Cave bears geologic characteristics more in keeping with today's enlarging sea caves. The physical appearance of the cave is that of an undercut overhang, reducing in size vertically and laterally toward its terminus. The presence far back in the cave of well-rounded granite boulders, up to approximately eighteen inches in diameter, suggests aggressive wavecut erosion along either bedding planes and joints or along a localized fracture zone. The degree of erosion evident in the cave appears inconsistent with the more limited degree of natural weathering observed in such locations as the summit of Cadillac Mountain. A gentle gradient in the cave's floor from its terminus downward to the entrance is similar in nature to that observed in Great Head Cave. This cross-sectional rise in the elevation of the cave's floor, from front to back, is consistent with an expected reduction in erosive capacity as wave force attenuates toward the shoreline. The evidence indicates that this slope once graded eastward into the foreland of an ancient coastline.

Additional evidence that the Cadillac Cliffs Sea Cave is indeed a sea cave is found along the base of the Cadillac Cliffs, where numerous joints and fractures have been opened by wave carving. Excellent examples of wave-enlarged fractures can be found in the form of small caves (171 ft msl) near the Waldron Bates memorial plaque situated immediately northeast of the Cadillac Cliffs trail junction. Nearby wave-cut bedrock is suggestive of limited wave erosion.

Champlain Mountain Sea Cave (Figure 5) is situated on the northern flank of Champlain Mountain within sight of the Park Loop Road, a short distance west of the Champlain Mountain trailhead. Its elevation was determined by a leveling survey from the Bear Brook Pond. The pond's elevation is cited as 137 feet msl on the 1983 Seal Harbor Quadrangle. A beaver dam at the pond's outlet may possibly have raised the pond's elevation by 4 feet. The cave was found to occur at an elevation between 218 and 227 feet msl. This spectacular emerged sea cave is found at the back end of what was once a small rock-bounded cove. Massively fractured granite, with vein filling, provided a favorabie milling surface for breaking waves. This site would make an excellent tourist attraction.

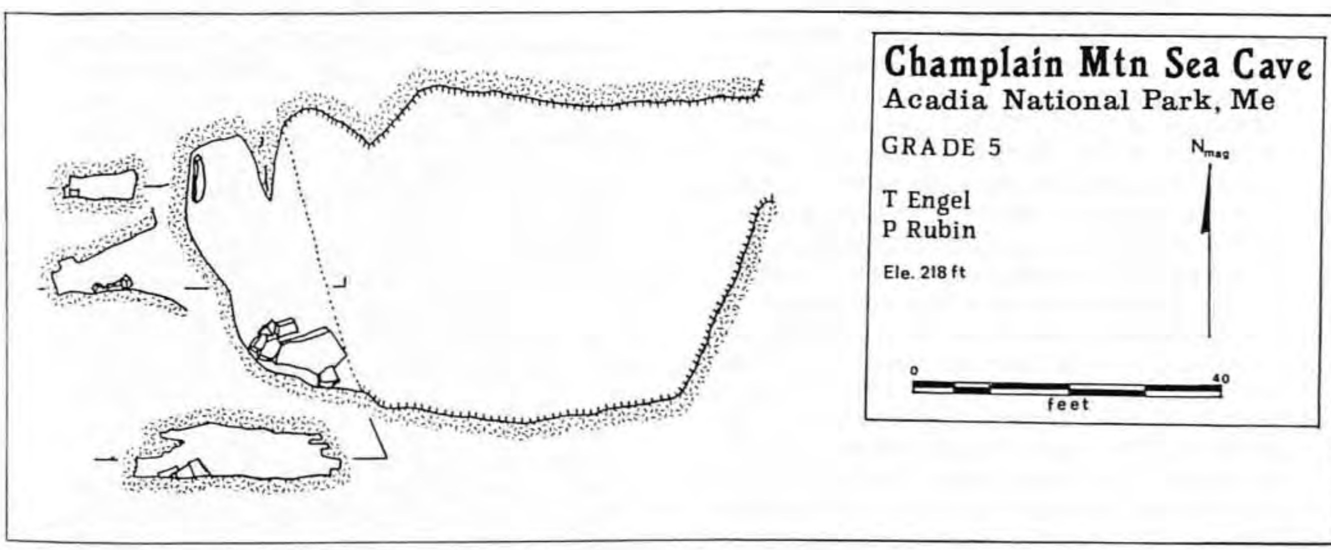

Figure 5: Map of Champlain Mountain Sea Cave.

Pulpit or Tilted Rock

Pulpit Rock is located at the base of a 19-foot cliff that is approximately 2000 feet southeast of The Cleft on Day Mountain and approximately 2900 feet from the summit of Day Mountain, along a bearing of S 34 E. Pulpit or Tilted Rock looks remarkably like a coastal sea stack, such as the one in Monument Cave. Various researchers assert that it *is* sea stack (Shaler, 1889; Bascom, 1919); others believe that it *is not* (Raisz, 1929; Johnson, 1925), and others believe that it *may be* (Coffin and others, 1990).

Pulpit Rock is pinnacle in shape, rises 25.9 feet from its lowest exposed point, and is separated from an upslope, laterally extensive cliff by several feet. The cliff base exhibits numerous wave-eroded joints and fractures, both adjacent to Pulpit Rock and at other locations along the cliff line. The top of Pulpit Rock was surveyed as 230 feet msl. Nearby clusters of rounded and polished beach cobbles and boulders were surveyed between 185 and 201 feet msl. The highest beach-cobble cluster found (201 feet msl) was only 3 feet below the base of Pulpit Rock. The nearby emerged beach boulders, the wave-cut cliff, the physical character of Pulpit Rock itself, the possible wave-turned upper stone of Pulpit Rock, and its elevation within known former shorelines argue that it is indeed an emerged sea stack.

Dating the Deglaciation of Mount Desert Island

During the Wisconsin glacial period, the Laurentide Ice Sheet covered Canada and parts of the United States to a maximum elevation of about 2.2 miles (3500 m), with about 2.6 miles (4200 m) of ice (Eyles and others, 1983). The land surface was depressed due to isostatic subsidence. Similarly, Quigley (1983) estimates that the Laurentide Ice Sheet may have been up to 3.1 miles (5000 m) thick at the time of maximum glacial advance, 20,000 to 18,000 years before present (20-18 ka). The time frame of maximum glacial advance prior to the onset of climatic warming and glacial retreat is further corroborated by the work of Morgan (1987). Morgan studied fossil assemblages of *Coleoptera* (beetles) in North America in order to establish the nature and timing of different paleoenvironments at the margins of the Laurentide Ice Sheet as it advanced and retreated. Morgan documents insect recolonization following retreat of the ice from its maximum at 18 ka and places the retreat of the ice front along the Maine coastline somewhere between 14 and 13 ka. Teller (1987) also places the southern margin of the Laurentide Ice Sheet north of Mount Desert Island by 13.5 ka. Gilman and others (1988) estimate that a continental glacier covered Mount Desert Island some 14,000 years ago, having receded by 12,500 years ago. Thus, Cadillac Cliffs and Champlain Mountain sea caves must be younger than approximately 14 ka.

Cadillac Cliffs and Champlain Mountain sea caves had to have formed at a time when either the elevation of the land surface was considerably lower or when the sea level was considerably higher. At the time of the glacial maximum, the glacio-eustatic (a change in the elevation of mean sea level caused by the growth and decay of ice masses) ocean-water levels were at least 394 feet (120 m) below present-day levels (Quigley, 1983). Gilman and others (1988) similarly place the minimum eustatic ocean-water level at 328 feet (100 m) below today's mean sea level. A slow rise in sea level, broken only by minor glacial re-advances, followed as continental ice melted and receded. As continental ice receded from Mount Desert Island, the depressed land surface began to rise, or rebound. Although rebound continued to occur for thousands of years following glacial recession, the greatest rebound occurred immediately after glacial recession. Gilman and others (1988) document an elevated delta (a shoreline feature) south of Jordan Pond at 230 feet (70 m) above today's sea level. Shells from nearby clay deposits, dated by radiocarbon methods, establish a date of approximately 12.25 ka for this former sea level. Gilman and others point out that the lack of erosionally modified rocks or clay deposits higher than the Jordan Pond delta suggest that the sea never rose above 230 feet (70 m). Stuiver and Borns (1975) dated shells and seaweed taken from marine sediment from 22 Maine locations interpreted as having been deposited near melting ice. They determined that between 12.2 and 12.6 ka, and more likely before ~ 12.7 ka, the ice sheet had retreated from its terminal position on the continental shelf and passed northward through central Maine. Andrews (1987) reports the oldest dated shell-remains from the deglaciation of the Maine coast to be 13.4 ka. The lack, at elevations above emerged marine sediments, of organic remains associated with emerged sea caves and boulder beaches may indicate that these features are somewhat older.

Gilman and others (1988) state that fossils from marine clays discussed above reveal that Mount Desert Island rose some 217 feet (66 m) above today's sea level between 12 and 11 ka. It is possible that the only evidence for a higher post-glacial rebound on Mount Desert Island is Cadillac Cliffs Sea Cave and the upper limit of the emerged Day Mountain boulder beach. Quigley (1983) presents the worldwide eustatic sea level curves of Kenney (1964) and Morner (1971). Kenney (1964) portrays an oscillating worldwide sea level ranging between 230 and 131 feet (70 and 40 m) below today's sea level between 15 and 12 ka. Morner's (1971) eustatic sea level curves show a sea level fluctuating between 249 to 197 feet (76 and 60 m) below today's sea level between 15 and 12 ka. Thus, the glacio-isostatic rebound suggested by the Day Mountain boulder beach (245 feet above today's sea level) is actually more on the order of 377 to 492 feet (115 to 150 m) above the glacio-eustatic sea level of some 14 ka. Cadillac Cliffs Sea Cave and the emerged Day Mountain boulder beach then formed at an elevation of 40 to 76 meters below today's sea level. It is likely that the cave

formed between 14 and 12 ka, coincident with the recession of the continental glacier.

Late Wisconsinan Reconstruction

The suggested Day Mountain upper marine limit (245 feet msl) coincides with massive talus lying at the foot of The Cleft, and may therefore not be truly representative of the maximum glacio-isostatic crustal uplift. It is interesting to note that the distance between the defined upper boundary of the Cadillac Cliffs Sea Cave's emerged boulder beach and the roof of the cave is 40 feet vertically and 89 feet horizontally. As discussed previously, the Cadillac Cliffs Sea Cave falls seven feet below the upper marine limit established on Day Mountain, thus indicating that the top of a boulder beach may not be truly reflective of an upper marine limit. Instead, the documented elevational continuum of well-rounded beach cobbles and boulders, sometimes with associated sea caves, may represent a period of relatively uniform glacio-isostatic rebound with accompanying marine transgression. It might be argued that the surveyed top of the Day Mountain cobble and boulder beach represents the uppermost elevation of a former storm berm in the supratidal region. The apparent elevational and lateral consistency of this emerged shoreline supports a mean high-water interpretation. Further credence is given to this interpretation by the 238-foot-msl elevation of Cadillac Cliffs Sea Cave, which must have been exposed to constant wave attack for some period of time. In light of the physical evidence available at this time, and in the absence of further verifiable evidence of a higher upper marine limit, the uppermost surveyed elevation of Day Mountain's emerged shoreline (245 feet msl) will be assumed to be coincident with the maximum glacio-isostatic rebound of Mount Desert Island.

Today, Mount Desert Island encompasses approximately 107.8 mi^2. By utilizing the 245-foot elevational contour, it is possible to reconstruct the early postglacial appearance of the Mount Desert Island area at ~14 ka. At this time, only 25 percent of today's land mass projected above sea level. Seven islands over 200 acres in size were present. Six of the largest of these islands encompassed Cadillac, Dorr, Day, Pemetic, Sargent, Norumbega, and Youngs Mountains (13,132 acres); Western and Bernard Mountains (1,732 acres); Champlain and Gorham Mountains (967 acres); Beech Mountain (741 acres); St. Sauveur Mountain (393 acres); and Acadia Mountain (213 acres). A number of other smaller islands were also present.

Day Mountain Cave - Evidence of the Upper Marine Limit?

It is possible that Day Mountain Cave might provide the only direct evidence of higher late-Wisconsinan glacio-isostatic rebound for coastal Maine. This small granite cave, 36 feet in length, is present virtually at the base of The Cleft on Day Mountain. Its elevation was recently surveyed, placing it at an elevation of 443-451 ± 5 feet msl. Day Mountain Cave has a maximum ceiling height of 8.7 feet near its terminus, and it is floored with a dry saprolitic granite-rich sediment. The rear and ceiling of the cave are uniformly free of any significant infiltration pathways, along which dissolution may have occurred. Its length is beyond that which might be expected from driving rain or freeze-thaw mechanisms. No viable Holocene mechanism appears available to account for soil and rock removal from the cave. In addition, no surficial-runoff or tributary-watershed explanation can provide infiltration to the cave. Figure 6 reveals the convex-upward profile at the terminus of the cave's longitudinal section, almost as if it was formed by waves crashing into it along a thin fracture.

Until a viable alternative explanation for the formation of Day Mountain Cave is forwarded, a milling or wave-attack origin cannot be discounted. Similarly, the ultimate placement of the upper marine limit for coastal Maine may also have to account for this cave. A brief literature search indicates that cavern development in granite is predominantly limited to solution within fractures along soluble vein fill or grussified (Esch, 1991) material open to erosional stripping. Finlayson (1983), Hose (1991), and Esch (1991) discuss various types of granite caves, all of which require the removal of weathered

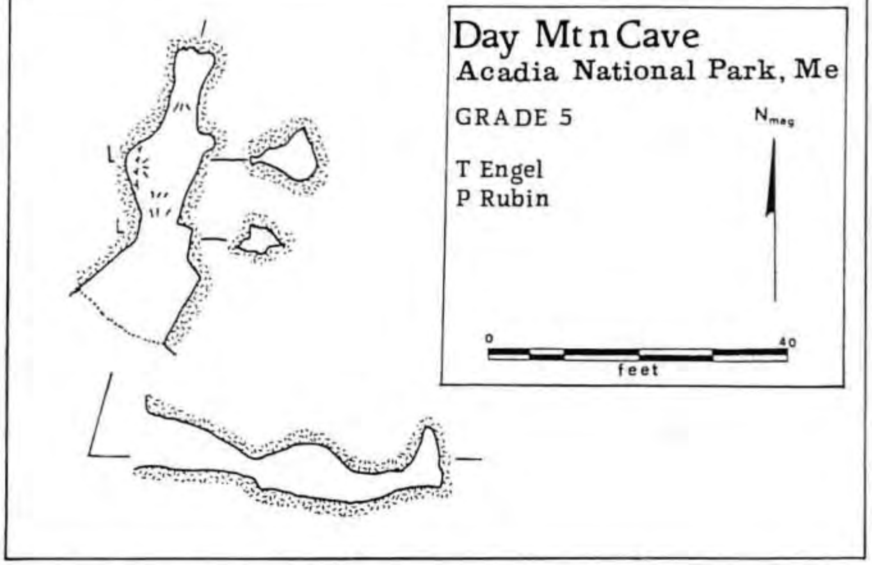

Figure 6: Map of Day Mountain Cave.

granite by flowing water. Alternately, Dragovich (1969) discusses the formation of sidewall and basal tafoni in granitic and gneissic rocks. Sidewall tafoni, found on vertical or near vertical rock faces, generally show forms ranging from a shallow ellipse to a near semi-circle. They form largely as a result of weathering by moisture and temperature variations, having been documented to depths of nearly eight feet.

Thompson and others (1989) have documented isostatic crustal-uplift increasing from coastal to interior Maine from 193 to 422 feet msl. Perhaps the initial glacio-isostatic rebound concurrent with a thinning or retreating ice sheet was rapid and sufficiently short-lived in the Mount Desert Island area so as to obviate the accumulation of any sedimentary record along small, cliffed islands. The combination of initial rapid uplift and exposed steep topography may account for the lack of sediment deposition or boulder-beach development (above 245 ft. msl) concurrent with ice sheet thinning or deglaciation. Perhaps the development of the Day Mountain boulder beach heralded a reduction in crustal-uplift rates from even higher initial rates. No additional speculation is warranted without further field work.

Conclusion

Thompson and others (1989) have assigned a maximum crustal-uplift value of 231 feet to Mount Desert Island, based on the elevation of an ice-contact glaciomarine delta. This study has established a slightly higher maximum crustal-uplift value based on the surveyed elevations of emerged boulder beaches and sea caves. Development of the Cadillac Cliffs Sea Cave must follow the retreat of continental ice from Gorham Mountain prior to substantial rebound and precede the lower sea-level stand documented in the elevationally lower clay deposits (12.25 ka). Because rapid crustal rebound occurred concurrently with ice thinning or deglaciation, it appears reasonable to attribute a date of between 14 and 12.7 ka to the emerged Cadillac Cliffs Sea Cave and the surveyed, upper limit of the Day Mountain boulder beach (245 ft msl). Additionally, this study has raised questions regarding whether a higher maximum glacio-isostatic rebound figure might be verified at Day Mountain Cave through further investigation.

Acknowledgments

The author extends thanks to Thom Engel who, even in pouring rain, ably manned the other end of the leveling tube. Thom is also responsible for the cave surveys and drafting of the maps accompanying this article. Thom provided good humor and company over the course of many fine days on Champlain's "Isle des Monts Deserts".

References

Andrews, J.T., 1987, The late Wisconsin glaciation and deglaciation of the Laurentide Ice Sheet, *in* Ruddiman, W.F. and Wright, H.E. Jr. (editors), *The Geology of North America, v. K-3, North America and Adjacent Oceans During the Last Deglaciation:* The Geological Society of America, p. 13-37.

Bascom, F. 1919. The geology of Mount Desert Island. *Geographical Society of Philadelphia, Bulletin 17*, p. 117-130.

Belknap, D.F. and others, 1987, Late Quaternary sea-level changes in Maine, *in* Nummedal, D.; Pilkey, O.H., Jr.; and Howard, J.D. (editors), Sea level fluctuation and coastal evolution: *Society of Economic Paleontologists and Mineralogists, Special Publication No. 41*, p. 71-85.

Coffin, T.E.; Tyler, Jr., H.R.; and Broyer, M.W., 1990, *An Evaluation of Sites Representing Emerged Marine Shoreline Features in the New England-Adirondack Region for Their Eligibility as National Natural Landmarks*, Prepared for Maine State Planning Office.

Dragovich, D., 1969, The origin of cavernous surfaces (tafoni) in granitic rocks of southern South Australia, *Zeitscrift für Geomorphologie*, v. 13, p. 163-181.

Esch, L., 1991, Speleogenesis in the Lost Creek Wilderness Area, Colorado, *Geo²*, v. 18, p. 31-35.

Eyles, N.; Dearman, W.R.; and Douglas, T.D., 1983, The distribution of glacial landsystems in Britain and North America, *in* Eyles, N. (editor), *Glacial Geology*: Pergamon Press Inc., p. 213-228.

Fairchild, H.L., 1919, Postglacial uplift of southern New England: *Geological Society of America Bulletin*, v. 30, p. 597-636.

Finlayson, B., 1983, The formation of caves in granite, *in Proceedings of the Anglo-French Karst Symposium, September 19-26, England and Wales, United Kingdom.*

Gilman, R.A.; Chapman, C.A.; Lowell, T.V.; and Borns, Jr., H.W., 1988, *The Geology of Mount Desert Island: A Visitor's Guide to the Geology of Acadia National Park*: Maine Geological Survey, Department of Conservation.

Hose, L.D., 1991, Millerton Lake Caves: A new model for granite cave development: *Geo²*, v. 18, p. 36-37.

Johnson, D., 1925, *New England Acadian Shoreline*: John Wiley and Sons, N.Y.

Kaszycki, C.A. and Shilts, W.W., 1979, Average depth of glacial erosion, Canadian Shield: *Geological Survey of Canada Paper 79-1B*, p. 395-396.

Kenney, T.C., 1964, Sea level movements and the geologic histories of the post-glacial marine soils at Boston, Nicolet, Ottawa and Oslo: *Geotechnique*, v. 14, p. 203-230.

Morgan, A.V., 1987, Late Wisconsin and early Holocene paleoenvironments of east-central North America based on assemblages of fossil *Coleoptera*, *in* Ruddiman, W.F. and Wright, H.E. Jr. (editors), *The Geology of North America, v. K-3, North America and Adjacent Oceans During the Last Deglaciation*: The Geological Society of America, p. 353-370.

Morner, N.A., 1971, Eustatic changes during the last 20,000 years and a method of separating the isostatic and eustatic factors in an uplifted area: *Palaeogeography, Palaeoclimatology, Palaeoecology*, v. 9, p. 153-181.

Palmer, A.N., 1987, Cave levels and their interpretation: *National Speleological Society Bulletin*, v. 49, p. 50-66.

Quigley, R.M., 1983, Glaciolacustrine and glaciomarine clay deposition: A North American perspective, *in* Eyles, N. (editor), *Glacial Geology*: Pergamon Press Inc., p. 140-167.

Raisz, E.J., 1929, The scenery of Mt. Desert Island: Its origin and development: *Annals of New York Academy of Science*, v. 21, p. 121-186.

Shaler, N.S., 1889, The geology of the Island of Mount Desert, Maine: *U.S. Geological Survey, Eighth Annual Report*, Pt. 2, p. 987-1061.

Shipp, R.C.; Staples, S.A.; and Adey, W.H., 1985, Geomorphic trends in a glaciated coastal bay: A model for the Maine coast: *Smithsonian Contributions to the Marine Sciences*, no. 25.

Smith, D.A., 1966, *Late-Glacial Emerged Shoreline Features of Mount Desert Island, Maine*: A report submitted to Acadia National Park, 55p.

Stone, G.H., 1899, The glacial gravels of Maine and their associated deposits: *U.S. Geological Survey, Monograph No. 34*, 499p.

Stuiver, M. and Borns, H.W., 1975, Late Quaternary marine invasion in Maine: Its chronology and associated crustal movement: *Geological Society of America Bulletin*, v. 86, p. 99-104.

Teller, J.T., 1987, Proglacial lakes and the southern margin of the Laurentide Ice Sheet *in* Ruddiman, W.F. and Wright, H.E., Jr. (editors), *The Geology of North America, v. K-3, North America and Adjacent Oceans During the Last Deglaciation*: The Geological Society of America, p. 39-69.

Thompson, W.B.; Crossen, K.J.; Borns, Jr., H.W.; and Andersen, B.G., 1989, Glaciomarine deltas of Maine and their relation to Late Pleistocene-Holocene crustal movements, *in* Anderson, B.G. and Borns, H.W., Jr. (editors), Neotectonics of Maine: *Maine Geological Society, Bulletin 40*, p. 43-67.

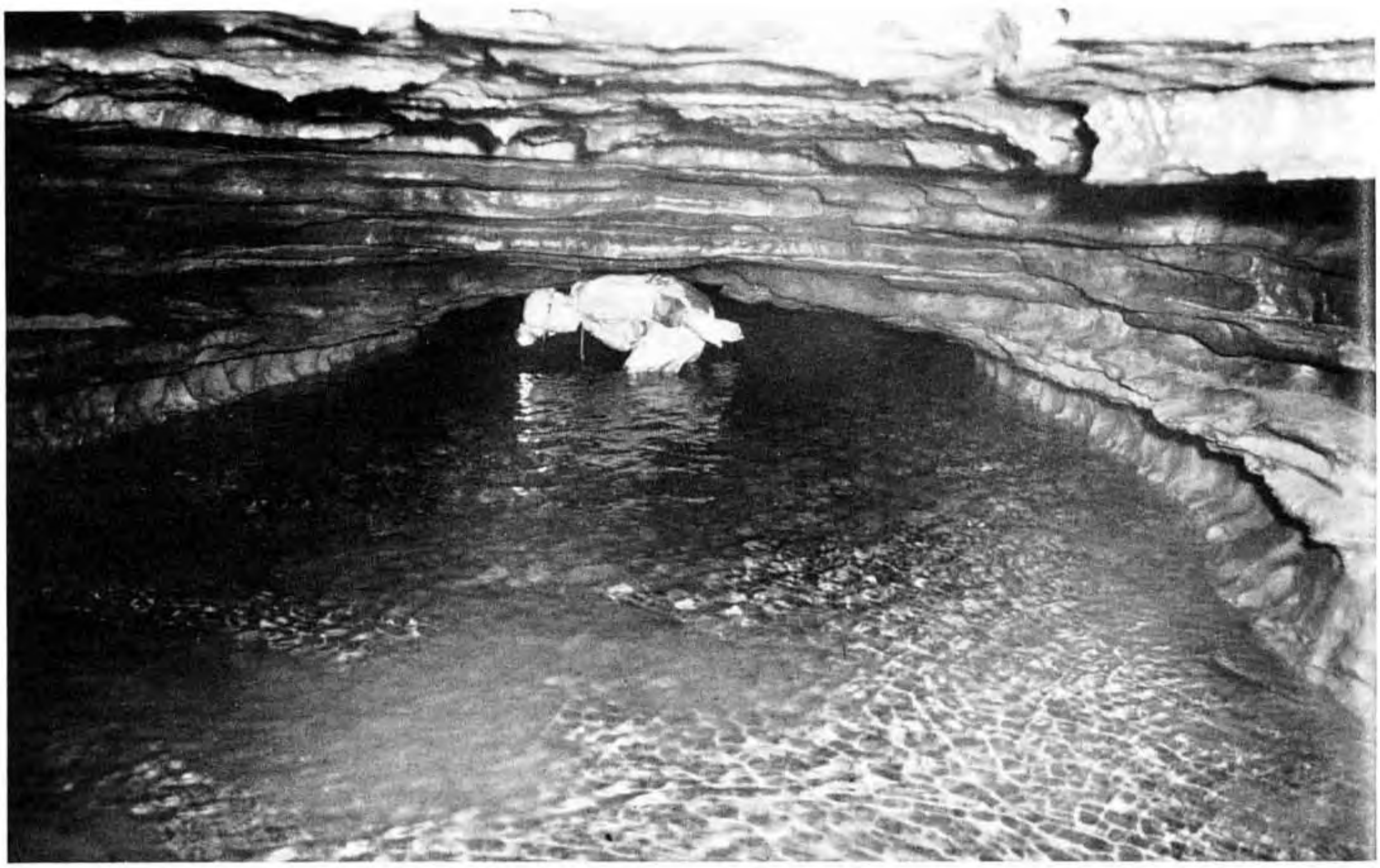

Plate B: The Sump in the northeastern section of McFail's Cave, Schoharie County, New York. Penetration of this nearly water-filled segment in the early 1960's led to the discovery of approximately five miles of cave passages. This cave stream flows for a distance of about 2.5 miles through explored sections of the cave. Scallops are visible on the walls just above the water level. For a map of the cave, *see* Figure 1A of Mylroie, this volume, page 87. *Photograph by Ernst H. Kastning.*

Cave Development in the Glaciated Appalachian Karst of New York: Surface-Coupled or Saline-Freshwater Mixing Hydrology?

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ABSTRACT

Recently, Panno and Bourcier (1990) proposed a new hypothesis for the mechanism of cave formation in the eastern midcontinental United States. They suggest that the dissolution of limestone occurred as a result of the mixing of saline formation waters with shallow, meteoric-derived fresh groundwater. Pleistocene glacial loading was suggested as the mechanism which caused discharge of the saline waters. Laboratory experiments and mixing-zone examples from the Yucatan were used to illustrate the dissolutional mechanism. Further, the occurrence in the midcontinent of carbonate and sulfate rocks, MVT-mineralization zones and basin margins, karst regions, and the limit of Pleistocene glaciations were all cited as field evidence that supports the hypothesis.

Unfortunately, the hypothesis is flawed because the field evidence does not in fact support what otherwise appears to be a viable model. The following are some of the major concerns: 1. Calculations of the amount of rock removed by the mixing process is trivial compared to the initial conditions. 2. Sulfate rocks cannot be treated as analogs of carbonate rocks in a discussion of mixing-zone dissolution. 3. MVT mineralization in the Midwest had nothing to do with Pleistocene glaciation. 4. The model does not explain passage levels or passage geochronology. 5. The relationship of karst development to the southern margin of glaciation reflects the mantling or quarrying of karst features by glacial action, not preferred karst development at the glacial margin. 6. The appearance of saline waters in modern caves is an effect, not a cause, of cave formation. 7. The morphology, plan, and profile of caves developed in or near glaciated areas does not differ from that of caves developed outside the range of influence of glaciations; caves developed in conditions of mixed-water (either H_2S/freshwater or saline/freshwater) are distinct from caves developed by meteoric water/CO_2-driven dissolution.

Comparison of cave morphologies from glaciated Appalachian karst of New York with cave morphologies from mixed-water caves of the Bahamas and Guadalupe Mountains illustrates several significant differences between them. Caves of mixed-water origin have globular, irregular chambers with numerous windows, thin wall-partitions, and blind passages. In comparison, New York caves resemble those from all over the world that have developed as a component of CO_2-charged meteoric-water infiltration, underground transport, and discharge back to the surface. These caves are dendritic, coupled to the surface hydrology, and contain little evidence of a mixed-water morphology.

Introduction

The dissolution of calcium carbonate on a scale necessary to form caves requires that water flowing within the carbonate aquifer be chemically aggressive during cave genesis. Bögli (1964) demonstrated that the mixing of waters saturated with calcium carbonate at different partial pressures of CO_2 results in renewed dissolutional aggressivity of the mixed water. Beginning in the mid-1970's, the importance of mixing waters of different salinity in the formation of caves was recognized (Plummer, 1975; Palmer and others, 1977a,b; Back and others, 1979). The development of secondary porosity, including caves, in carbonates as a result of mixing waters of different salinities is now a well-established phenomenon (Back and others, 1984, 1986; Mylroie, 1988; Proctor, 1988; Sanford and

Konikow, 1989; Stoessell and others, 1989; Mylroie and Carew, 1990; Vogel and others, 1990). Other investigations have studied other mixed-water phenomena that involve thermal, sulfate, and sulfide-charged waters (Egemeier, 1981; Hill, 1987; Bakalowicz and others, 1987; Palmer and Palmer, 1989; Ford, 1989). These are the *hypogenic* caves of Palmer (1991). The use of mixed-water models has helped explain the unusual morphologies of dissolutional caves found in such disparate settings as the Bahama Islands, Guadalupe Mountains of New Mexico, and Black Hills of South Dakota.

A novel new model that utilizes the effects of mixed waters for cave development was proposed by Panno and Bourcier (1990). They suggested that the caves along the southern margin of the Pleistocene glacial advances in the midcontinent of the United States were the result of glacial loading and meltwater discharge that flushed basinal formation brines to shallow depths in limestones. There, it was suggested, those brines combined with glacial meltwater and meteoric water to produce a mixed water with the aggressivity to dissolve caves. If their hypothesis is valid, then these caves should have a morphology common to that of mixed-water caves elsewhere, which is different from that seen in classic dendritic caves that are coupled to the surface hydrology. The caves cannot be older than the onset of the Pleistocene glaciations. The caves should be demonstrably different from caves formed in similar rocks well away from the influence of glaciation and possible expelled brines.

Examination of the arguments presented by Panno and Bourcier (1990), coupled with field observations, suggest that discharge of formation brines from intracratonic basins as a result of the effects of Pleistocene glaciation did not have a significant influence on cave development in the midcontinent of the United States. The caves that occur near the glacial margins of the midcontinent appear to have developed like most caves in the eastern United States. That is, as dendritic caves that function as part of the meteoric hydrologic cycle. Caves from the glaciated Helderberg Plateau of New York will be compared to freshwater/salt-water mixing caves of the Bahamas to demonstrate the morphological differences between caves produced by meteoric freshwater acting alone and those produced by mixed waters.

Problems with the Panno-Bourcier Model

The Panno and Bourcier (1990) model has a number of serious difficulties. These are listed and discussed below.

1. Based on their calculations, the *optimal* mixing ratio of brines to fresh water produces an additional dissolutional capability of 0.06 mmol/l. The length of time during which this activity occurred (and must have been less than that of the Pleistocene glaciations) was not discussed. Using a slightly lower value of 0.04 mmol/l, and an initial limestone porosity of 10% (high for Paleozoic limestones) they determined that 300 m^3 of limestone could be dissolved for each square kilometer of limestone. That is equivalent to a 2 m diameter cave passage 100 m long. This may at first seem to be significant cave development, but when the possible volume of limestone involved is considered, it is trivial. If we assume a mixing-zone thickness of 10 m, then a 1 km^2 surface area yields 1×10^7 m^3 of limestone. Removing, under optimal conditions, 300 m^3 results in an increase of porosity of 300 m^3/10,000,000 m^3 or 0.003%. That increase is negligible compared to the 10% porosity assumed as the initial conditions. With a 10% initial porosity, it is highly unlikely that the minor increased dissolutional potential they indicated would be concentrated to form one or more tubes of explorable size.

2. The model presents data on the distribution of carbonate and sulfate rocks in the midcontinent of the United States (their Figure 2A). The figure fails to differentiate between the locations of sulfates (gypsum and anhydrite), limestones, and dolomites. Further, the mixing model used in their arguments is not specific with regard to sulfate rocks, and does not deal with the difference between dolomites and limestones, which is necessary in light of recent work (Palmer and Palmer, 1989). Use of their Figure 2A to compare glacial margins, Mississippi Valley-type (MVT) mineralization, intracratonic basins, and karst development by basinal brines is diminished by the lack of specific data about soluble rock types in the figure.

3. The model uses MVT mineralization as a demonstration that brines can be expelled from intracratonic basins and interact with adjacent rocks. While that has merit, the model expands the idea and couples Pleistocene glaciation and cavern dissolution with MVT mineralization. This coupling seems contrived at best, and it does not match the distribution of glaciation and MVT deposits as seen in their Figure 2.

4. The paper describes caves of the midcontinent as "generally simple and consist of one or two tubular passages that have formed along joints and fractures" (Panno and Bourcier, 1990, p. 770). On the contrary, cave development within and near the Pleistocene glacial margin is much more complex than this description would imply. For example, Cold Water Cave in northeastern Iowa, Mystery Cave in southeastern Minnesota, and McFail's Cave and Skull Cave in east-central New York (Figure 1) are all from glaciated areas; all have long, large, and complex passage development. In addition, the Panno-Bourcier model does not explain this passage complexity, nor does it explain cave levels or passage geochronology.

5. The paper describes karstic terrain as concentrated around the southernmost extent of Pleistocene glaciation. This karst concentration is attributed to brine expulsion caused by the effects of glacial loading. Comparison of their Figure 2A with 2C and 2D demonstrates that karstic phenomena north of the glacial margin have most likely been glacially quarried or mantled with glacial drift. With-

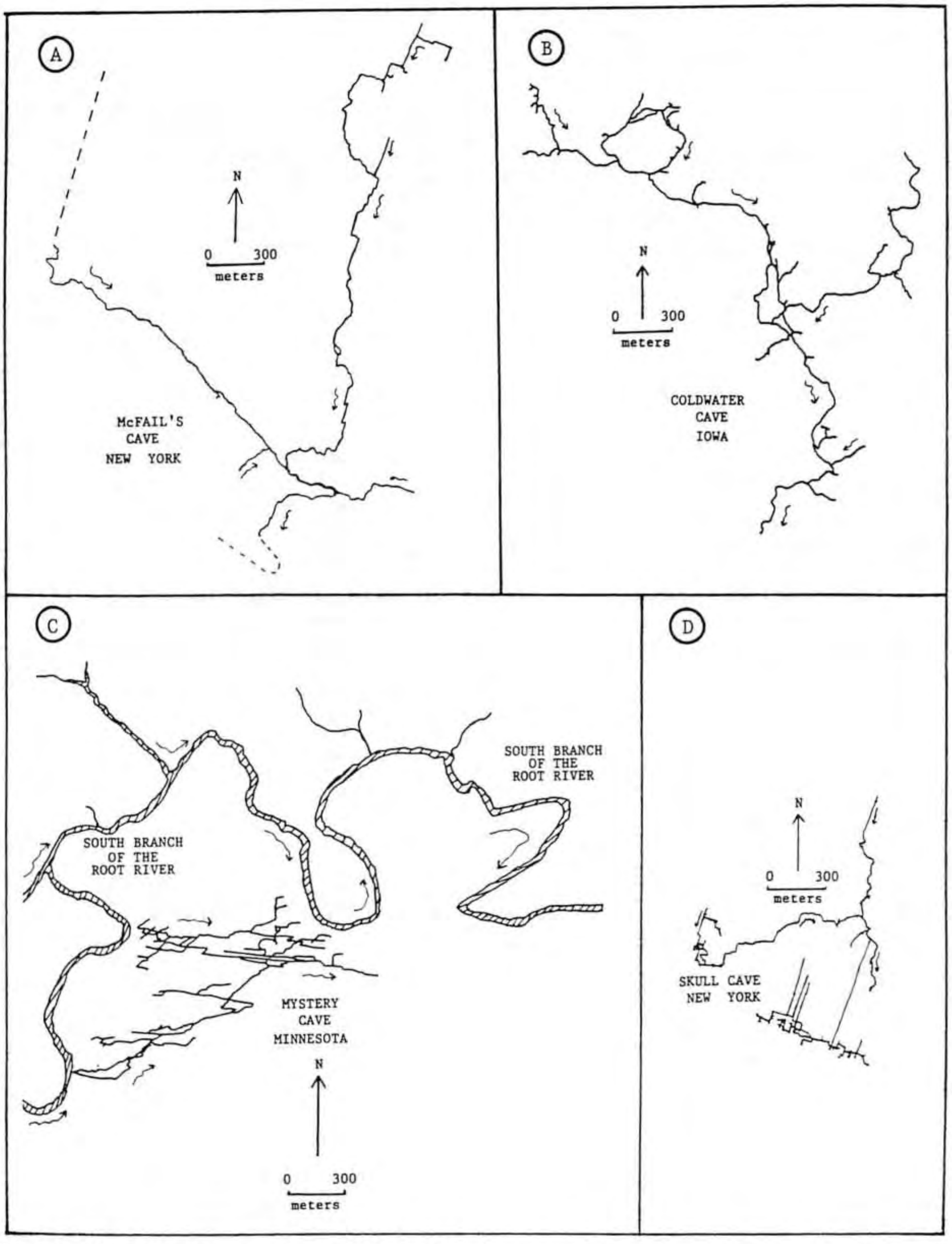

Figure 1: Maps of caves from glaciated regions of North America. Wiggly arrows indicate stream flow in the caves. A - McFail's Cave, New York, modified from Kastning, 1975. Dashed lines indicate recently discovered, unsurveyed passage. B - Coldwater Cave, Iowa, from Welch, 1989/90. C - Mystery Cave, Minnesota, from Alexander, 1989/90. The cave feeds water east across meander necks. River backflooding has produced maze development. D - Skull Cave, New York, from Kastning, 1975. The southern end of the cave is a backflooded maze caused by glacially-occluded resurgences.

out demonstration that the glacial margin karst is different from mantled karst to the north or glacially uninfluenced karst to the south, the model and its conclusions are not supported.

6. The occurrence of saline waters in glacial-margin karst systems is presented as evidence that brine mixing prduced the caves. It is equally probable, however, that the existing karst drainage systems have served as an escape path for brines, in which case the occurrence of the brines there is an effect and not a cause. Further, they did not discuss the abundance of saline water in karst springs in distant, unglaciated areas. Whereas the existence of halide-rich fluid inclusions in carbonate minerals in these caves does indicate exposure to halide-rich waters, it does not indicate that those waters were dissolutionally aggressive and helped promote cave development.

7. The key factor in their model is the use of a mixed-water mechanism to explain the formation of caves within and along the margin of the Pleistocene glaciations. Caves formed by mixed waters are generally unique in pattern, having a configuration of network, spongework, or ramiform mazes (Palmer, 1991). They rarely form dendritic systems. On the other hand, caves that developed in a groundwater regime coupled with the meteoric hydrologic cycle are usually dendritic. In those cases, maze configurations form only as an overprint in high discharge or back-flooded passages, or when surface recharge is extremely diffuse (Palmer, 1991). Caves along the glacial margin of the midcontinental United States are primarily dendritic and do not have the unusual patterns associated with cave systems of mixed-water genesis (Figure 1).

Comparison of Glaciated Appalachian Caves from New York with Mixed-Water Caves from the Bahamas

Glaciation has a dramatic impact on any landscape it directly affects, and through additional sea-level and climatic effects, can influence landscapes thousands of kilometers away from the ice margin. By exposing new limestone outcrops, incising valleys, and backflooding existing cave systems, glaciation has been demonstrated to have its own unique way of enhancing cave and karst development (Mylroie, 1984). Panno and Bourcier (1990) have proposed another type of glacially-enhanced karst development that involves glacial loading and subsequent brine expulsion into shallow limestone aquifers to produce mixing dissolution and cave formation. Support for their model hinges on whether caves from the area of Pleistocene glaciations have the characteristics of those of a mixed-water mode of initiation or development.

Mixed-water caves in the Bahamas have developed on a time scale (Late Quaternary) comparable to that of the proposed glacial-margin caves of Panno and Bourcier (1990). They also developed with minimal overprinting by hydrologic processes other than mixing of saline and fresh waters, and therefore represent a "type" sample, or end member, for this method of cave development. In contrast, caves of the Helderberg Plateau in New York developed in an environment that has been repeatedly glaciated. They are formed in dense Paleozoic limestones with properties similar to limestones found along the area of Pleistocene glaciation. If glaciation does produce mixed-water dissolution as proposed by Panno and Bourcier (1990), the Helderberg caves should have a mixed-water morphology. However, if mixed-water dissolution was not active, then the Helderberg caves should possess morphologies similar to that found in dense Paleozoic limestones of an area far from glaciation and possible brine expulsion, such as in Tennessee and northern Alabama.

The Bahama Islands contain a large number of caves that developed within a fresh or brackish water lens during times of past, higher sea levels. The islands are tectonically stable, and the limestones are Pleistocene in age (Mylroie, 1988). Therefore, dissolution that occurred during the glacio-eustatic sea-level high stands of the Pleistocene are responsible for the caves now found at 1 to 6 m above current sea level (Mylroie, 1988; Mylroie and Carew, 1990; Vogel and others, 1990). These caves have not been exposed to other hydrologic regimes, and therefore show minimal overprinting by other processes. These caves consist of a series of large chambers with numerous tubes that interconnect and end abruptly (Figure 2). The chambers are globular and irregular in shape and often separated from one another by extremely thin bedrock partitions through which small windows may have dissolved. This pattern is consistent with the ramiform or spongework pattern of mixed-water caves in high porosity, poorly jointed rock (Palmer, 1991). These caves are smaller than those found in the Guadalupe Mountains of New Mexico, and do not contain the abundance of sulfate mineralization found there, but in terms of passage morphology, cross section, and pattern, the Bahama caves are an excellent match for the hypogenic Guadalupe caves.

Caves of the Helderberg Plateau, on the other hand, exhibit a classic dendritic pattern (Figure 1). Glaciation has modified that initial pattern, primarily by shifting water input and output points (Mylroie, 1977). In some cases, backflood mazes have been superimposed upon a dendritic plan (Mystery Cave and Skull Cave, Figure 1). There is no evidence of the type of network development produced by hypogenic cave formation, such as that seen in the dense, well-jointed Paleozoic limestones of the Black Hills of South Dakota. Glaciation has also introduced sediment into pre-existing cave passages, and deposits that reflect ice advance, still-stand, and retreat are present (Mylroie, 1984). The arrangement of passages and their sediments indicate that the caves are or have been coupled to surface hydrology. Unusual ramiform or spongework passage development is not present. Network mazes are found, but their configuration and placement within cave passages indicate that they resulted from back-flooding of the cave system in response to obstructions by

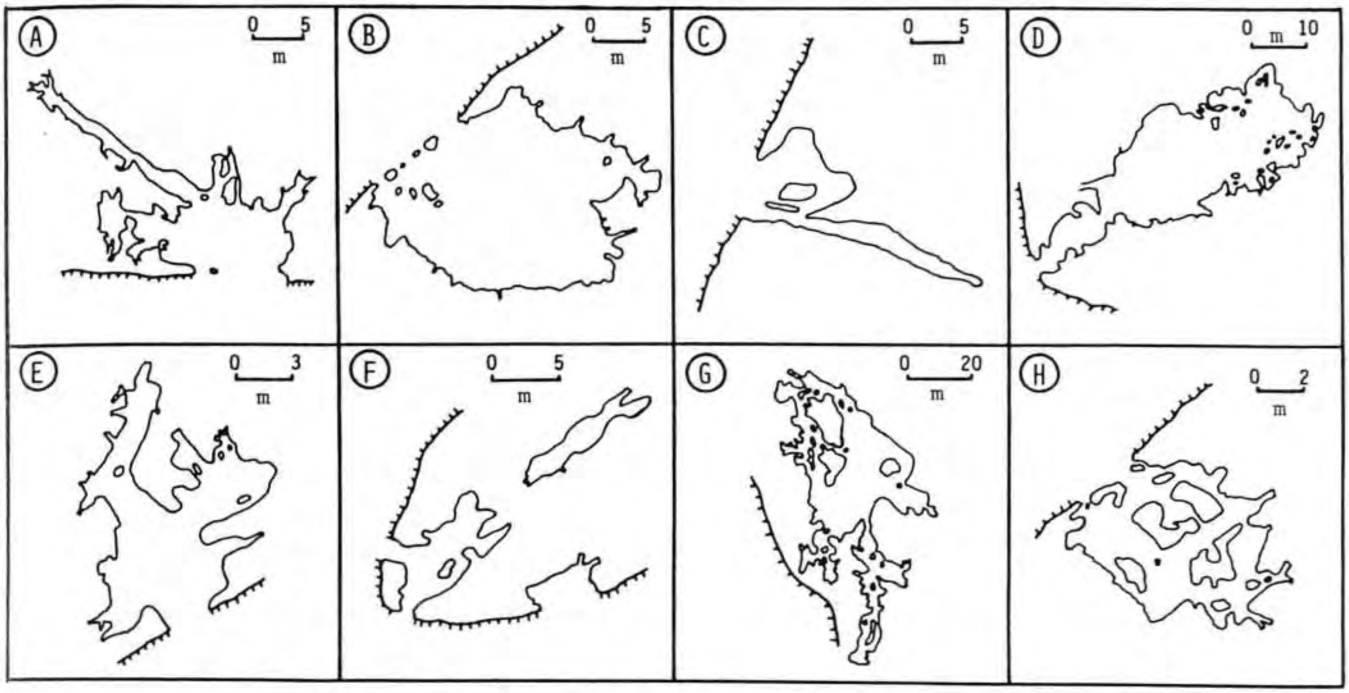

Figure 2: Selected maps of caves formed by mixed water in the Bahama Islands. North is to the top in all cases. Hachured line represents margin of lithified dune in which the caves are found. A - Dance Hall Cave, San Salvador Island. B - George Storrs' Cave, San Salvador Island. C - South Deep Creek Cave, South Andros Island. D - Beach Cave, San Salvador Island. E - Bug City Cave, San Salvador Island. F. - Maroon Hill Caves, Great Inagua Island. G - Lighthouse Cave, San Salvador Island. H - Reckly Hill Pond Cave, San Salvador Island. Note the variety of scales and the consistent general pattern of the caves. Modified from Mylroie and Carew, 1990.

glacial sediment. In the midcontinental United States, Mystery Cave, Minnesota, contains passages that are primarily a network maze (Figure 1C). The cave's setting, within a meander loop of a surface river, indicates that it is a meander-cutoff cave and its maze pattern is the result of river-induced backflooding (Mylroie and Mylroie, in press).

When compared to caves found in similar rocks in Alabama or Tennessee, the Helderberg caves are seen to be direct analogs. Despite glaciation, the New York caves share the overall dendritic pattern and morphology of their southern U.S. counterparts. In the Helderberg caves there are no passage morphologies, configurations, or deposits that relate to a mixed-water origin. There is no evidence of brines or the products of brines. All passage morphologies, configurations, and deposits are consistent with those produced by development of a dendritic cave system coupled to the meteoric hydrologic cycle. The Helderberg caves differ from their southern glacially-uninfluenced counterparts only by sediment and non-brine floodwater modifications produced by ice contact.

Conclusions

The role of glacially-expelled basinal brines on development of caves and karst in the midcontinental United States is minimal. The amount of extra rock that could be removed by that mechanism is insignificant. The relationship of differences in cave and karst development to the location of the Pleistocene glacial margin reflects the quarrying and mantling effects of glaciation not location of basin margins or MVT deposits. Co-occurrence of saline waters and caves today probably reflects the escape of formation brines through pre-existing conduits. The configuration, morphology, and pattern of caves in the proposed brine-expulsion area do not show the characteristics of cave systems of mixed-water genesis. Rather, those caves possess a dendritic pattern (often glacially modified, but not brine modified) that is consistent with development as part of a meteoric hydrologic pathway. The morphology of caves in the proposed brine-expulsion area do not differ from dendritic caves developed in areas far removed from glaciation.

Panno and Bourcier's (1990) model in which glacially-expelled basinal brines are an important cave-forming mechanism in the midcontinent is not supported by the field evidence. Glacially-induced expulsion of formation brines from intracratonic basins may have occurred, and may have added some minor component to the development of limestone porosity at the glacial margin. However, the effect is obscure and minimal compared to processes operating in the meteoric hydrologic cycle.

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References

Alexander, E.C., 1989/90, Karst hydrology of southeast Minnesota: *Geo*^{2}, v. 17, no. 2,3, p. 32-53.

Back, W.; Hanshaw, B.B.; Pyle, T.E.; Plummer, L.N.; and Weidie, A.E., 1979, Geochemical significance of groundwater discharge and carbonate dissolution to the formation of Caleta Xel Ha, Quintana Roo, Mexico: *Water Resources Research*, v. 15, p. 1521-1535.

Back, W.; Hanshaw, B.B.; and Van Driel, J.N., 1984, Role of groundwater in shaping the eastern coastline of the Yucatan Peninsula, Mexico, *in* La Fleur, R.A., *Groundwater as a Geomorphic Agent*, Allen and Unwin, Boston, Massachusetts, p. 281-293.

Back, W.; Hanshaw, B.B.; Herman, J.S.; and Van Driel, N.J., 1986, Differential dissolution of a Pleistocene reef in the ground-water mixing zone of Yucatan, Mexico: *Geology*, v. 14, p. 137-140.

Bakalowicz, M.J.; Ford, D.C.; Miller, T.E.; Palmer, A.N.; and Palmer, M.V., 1987, Thermal genesis of solution caves in the Black Hills, South Dakota: *Geological Society of America Bulletin*, v. 99, p. 729-738.

Bögli, A. 1964, Mischungkorrosion, ein Beitrag zur Verkarstungsproblem: *Erdkunde*, v. 18, p. 83-92.

Egemeier, S.J., 1981, Cavern development by thermal waters: *National Speleological Society Bulletin*, v. 43, p. 31-49.

Ford, D.C., 1989, Features of the genesis of Jewel Cave and Wind Cave, Black Hills, South Dakota: *National Speleological Society Bulletin*, v. 51, p. 100-110.

Hill, C.A., 1987, Geology of Carlsbad Cavern and other caves in the Guadalupe Mountains, New Mexico and Texas: *New Mexico Bureau of Mines and Mineral Resources Bulletin 117*, 150 p.

Kastning, E.H., 1975, Cavern development in the Helderberg Plateau, East-Central New York: *New York Cave Survey Bulletin 1*, 194 p.

Mylroie, J.E., 1977, Speleogenesis and karst geomorphology of the Helderberg Plateau, Schoharie County, New York: *New York Cave Survey Bulletin 2*, 336 p.

Mylroie, J.E., 1984, Pleistocene climatic variation and cave development: *Norsk Geografisk Tidsskrift*, v. 38, p. 151-156.

Mylroie, J.E., 1988, Karst of San Salvador, *in* Mylroie, J.E. (editor), *Field guide to the karst geology of San Salvador Island, Bahamas: 10th Friends of Karst Meeting*: Mississippi State University, p. 17-44.

Mylroie, J. and Carew, J., 1990, The flank margin model for dissolution cave development in carbonate platforms: *Earth Surface Processes and Landforms*, v. 15, p. 413-424.

Mylroie, J.E. and Mylroie, J.R., in press, Meander cutoff caves and self piracy: The consequences of meander incision into soluble rock: *National Speleological Society Bulletin*.

Palmer, A.N., 1991, Origin and morphology of limestone caves: *Geological Society of America Bulletin*, v. 103, p. 1-25.

Palmer, A.; Palmer, M.; and Queen, M., 1977a, Speleogenesis in the Guadalupe Mountains, New Mexico: Gypsum replacement of carbonate by brine mixing, *in* Ford, T.D. (editor), *Proceedings of the 7th International Speleological Congress, Sheffield, 1977*: British Cave Research Association, Somerset, England, p. 333-336.

Palmer, A.N.; Palmer, M.V.; and Queen, M.V., 1977b, Geology and origin of the caves of Bermuda, *in* Ford, T.D. (editor), *Proceedings of the 7th International Speleological Congress, Sheffield, 1977*: British Cave Research Association, Somerset, England, p. 336-339.

Palmer, A.N. and Palmer, M.V., 1989, Geologic history of the Black Hills, South Dakota: *National Speleological Society Bulletin*, v. 51, p. 72-99.

Panno, S.V. and Bourcier, W.L., 1990, Glaciation and saline-freshwater mixing as a possible cause of cave formation in the eastern midcontinent region of the United States: A conceptual model: *Geology*, v. 18, p. 769-772.

Plummer, L.N., 1975, Mixing of sea water with calcium carbonate ground water: *Geological Society of America Memoir 142*, p. 219-236.

Proctor, C.J., 1988, Sea-level related caves on Berry Head, South Devon: *Cave Science*, v. 15, p. 39-50.

Sanford, W.E. and Konikow, L.F., 1989, Porosity development in coastal carbonate aquifers: *Geology*, v. 17, p. 249-252.

Stoessell, R.K.; Ward, W.C.; Ford, B.H.; and Schuffert, J.D., 1989, Water chemistry and $CaCO_3$ dissolution in the saline part of an open-flow mixing zone, coastal Yucatan Peninsula, Mexico: *Geological Society of America Bulletin*, v. 101, p. 150-169.

Vogel, P.N.; Mylroie, J.E.; and Carew, J.L., 1990, Limestone petrology and cave morphology on San Salvador Island, Bahamas: *Cave Science*, v. 17, p. 19-30.

Welch, L., 1989/90, Coldwater Cave 1990: *Geo*^{2}, v. 17, no. 2,3, p. 90.

Modification of Preglacial Caves by Glacial Meltwater Invasion in East-Central New York

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ABSTRACT

Periods of high glacial meltwater have altered some preglacial cave-passage configurations. Floodwater and fossil karst features, whose formation cannot be explained based on available water from the surrounding watershed, are found superposed on actively forming cave passages. These features may be recognized through correlation of watershed boundaries, peak-runoff observations through a cave system, the presence of anomalous in-cave and surface features, and the geomorphic interpretation of the area in question. Knowledge of minimum rates of karstification may be used to infer climatic conditions, making possible the reconstruction of the hydrology associated with deglaciation.

Clarksville Cave, situated in the hamlet of Clarksville, New York, provides an excellent example of invasion by Wisconsinan meltwater on a preglacial cave system. Vadose development of a major part of the explored cave has occurred preferentially aslant a thrust-fault ramp, often along a calcite bed/limestone contact created by pressure solution. Other fault-related features include slickensides, extension veins, fault-bend folds, stylolites and the repeated basal Onondaga Limestone and impermeable Schoharie Formation thrust below the Onondaga Limestone stratigraphic column. An imbricate thrust east of the cave has upthrown the Esopus Shale against the Onondaga Limestone, forcing the development of an inefficient resurgence at the baselevel Mill Pond.

During the Wisconsinan glacial stage, subglacial meltwater formed a series of now abandoned bedrock channels and paleogorges that, due in part to topographic controls, found outlets along and over the flank of the Helderberg Escarpment. Some of this meltwater was pirated into Clarksville Cave where inefficient outlets resulted in the formation of higher in-cave "intermittent phreatic" levels not controlled by the thrust fault. These levels abruptly truncate and grade to lower vadose passages. The character of these upper levels, the paleogorge and related caves, and elevated paleo-insurgence points correlate with described alpine karst settings.

Physical Setting

Clarksville Cave is nestled under the flank of a low wooded ridge virtually in the center of the hamlet of Clarksville, New York (Figure 1). It is formed in the lower subunits of the Devonian Onondaga Limestone that were deposited approximately 380 million years ago. Its large passage size, up to 15 feet high and 40 feet wide, complete with multiple levels, makes it unique among other, usually smaller, Onondaga caves.

The Clarksville area lies at an elevation of 600 to 800 feet msl. It is situated within the foothills of the Helderberg Plateau, a part of the Appalachian Plateau physiographic province. Meyerhoff (1972) attributed the present-day drainage pattern of this region to the normal erosive processes of stream adjustment to structure. The Helderberg Plateau has been modified by stream incision, physical weathering, glacial and postglacial erosion, and deposition during the Cenozoic era (Dineen, 1987).

Dineen (1987) has determined that present-day drainage trends in the Hudson Valley were established before the Wisconsinan glaciation, sometime prior to 70,000 years ago. Glacial striations in two locations near Clarksville further indicate that today's drainage was in place prior to inundation by the Wisconsinan ice sheet. The direction of glacial movement was almost exactly north-south (S 13 W), with a maximum ice thickness on the order of one mile about 22,000 years before present (Dineen, personal communication). Late preglacial drainage along Onesquethaw Creek was probably little different from what it is

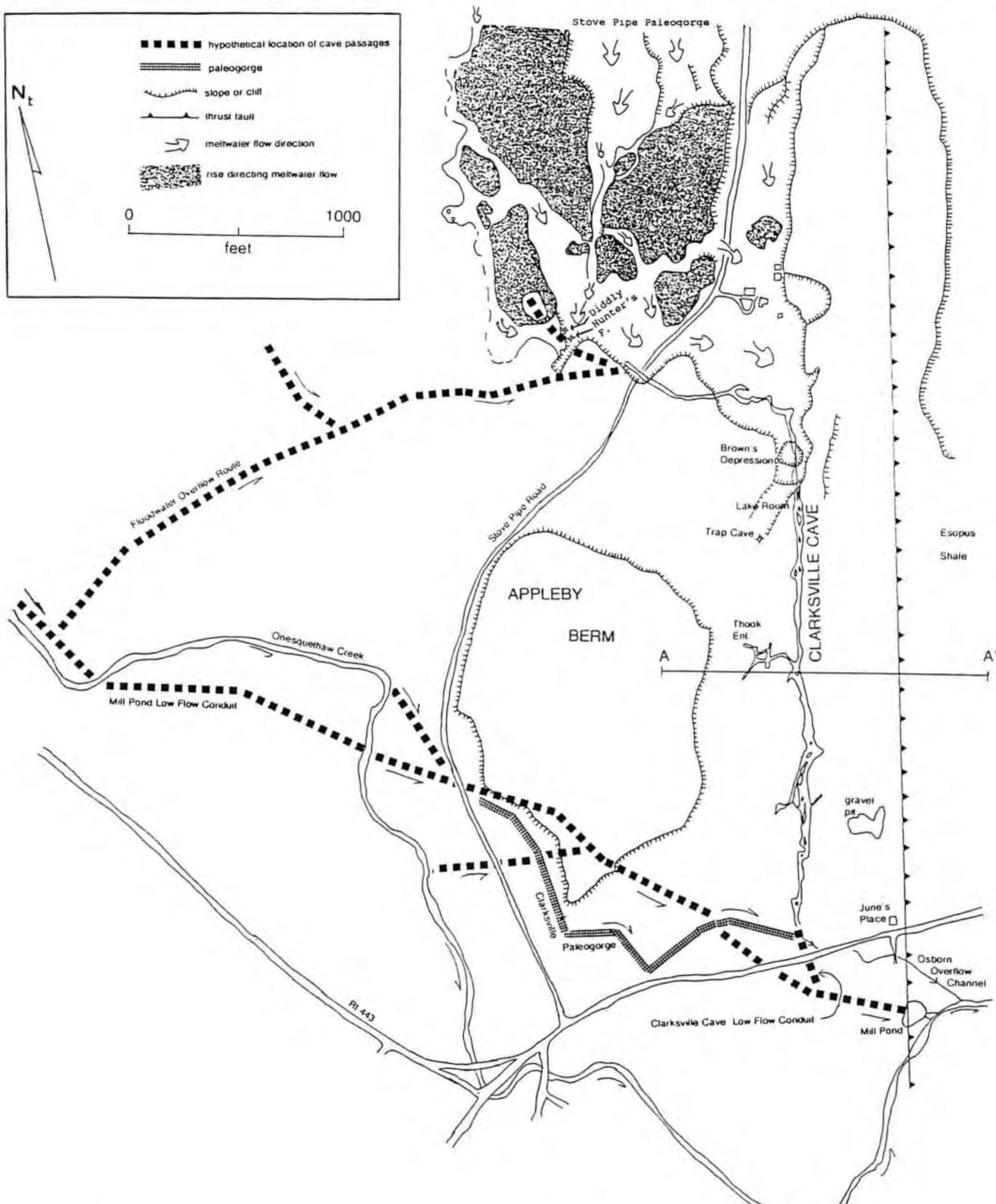

Figure 1: Configuration of drainage in the vicinity of Clarksville, New York. Shown are Clarksville Cave and present and past routes of flow. For a detailed map of Clarksville Cave, see Figure 3 of the following paper.

today. Evidence presented in this paper argues for pre-Wisconsinan cave development.

Structural Setting

The hamlets of Clarksville, Tarrytown and Feura Bush have all been subjected to extensive faulting. Marshak (1986), Marshak and Engelder (1987), and Cassie (1990) discuss structural deformation within parts of the Hudson Valley Fold-Thrust Belt (HVB). The HVB extends roughly from Kingston to Albany, New York, extending to a maximum of 20 kilometers east and west of the Hudson River (Marshak and others, 1986). The deformation may have occurred during the Acadian (Cassie, 1990) or Alleghanian orogenies (Geiser and Engelder, 1983), or during Mount Marion deposition (Murphy and others, 1980).

Faulting and deformation of the Esopus Shale, Schoharie Formation, and Onondaga Limestone, throughout the Clarksville area, may represent the farthest northwestern exposure of the Hudson Valley Fold-Thrust Belt. The extensive structural deformation present throughout Clarksville and the previously documented southern parts of the HVB are characteristic of deformation of sedimentary rock under relatively low pressure and temperature conditions (Marshak and Engelder, 1985). Mapping of the structural or bedrock geology in the area, both on the surface and in the cave, reveals that faulting in the Clarksville area is characteristic of either an imbricate thrust zone or a duplex.

At least six elongate ridges, trending north-south approximately along strike of the faults, are unevenly spaced throughout the Clarksville area. They often exhibit extensive fault-bend folding, slickensides, and in places an anticlinal structure. These limestone ridges, which are underlain by one or more basal thrust faults, can be mapped for distances of up to one mile. One such deformed ridge, situated at the eastern end of Clarksville, has been breached by Onesquethaw Creek. Perhaps the most prominent example is found in the upper Onesquethaw Creek gorge.

The upper Onesquethaw Creek gorge exhibits the best out-of-cave exposure of the repeated basal Onondaga Limestone and the impermeable Schoharie Formation (a quartz-rich limestone) thrust below the Onondaga Limestone stratigraphic column. Here much of the bed of Onesquethaw Creek is guided by fault-zone features. The thrust-fault ramp, associated thick calcite bed, and fault-bend folds are the same as those along which Clarksville Cave has developed, except that they are farther south along strike.

Structural Features Influencing Groundwater Flow in the Karst Aquifer

Faulting in and immediately east of Clarksville Cave has resulted in thrusting, deformation, and upward movement of impermeable bedrock units underlying the Onondaga Limestone (Schoharie Formation and Esopus Shale) into a position that makes the eastern escape of groundwater impossible. The gentle southwesterly dip of the bedrock of the Mill Pond aquifer (Figure 2) fails to direct all subsurface flow in this direction. Instead, significantly higher surface topography to the southwest (*e.g.*, Wolf Hill and Cass Hill) retards dissolution in this direction, in favor of the 1.3-degree apparent dip between Wolf Hill Dam and the base-level discharge point at Mill Pond. Tracer studies generally verify this predicted flow path, at least during periods of low discharge. However, tracer studies also document an unexpected easterly diversion of moderate- to high-discharge waters through Pauley Avenue in Clarksville Cave. This is significantly farther north than the Mill Pond. This easterly deflection of floodwaters may occur in response to an inefficient outlet and conduit leading to the Mill Pond.

Pauley Avenue floodwaters flow easterly until they become perched on a thin bed of impermeable Schoharie Formation that has been thrust below the basal, or lower, non-cherty subunit of the Onondaga Limestone. Cave diver John Schweyen (personal communication) reports the presence of the Schoharie Formation overlying the lower non-cherty subunit of the Onondaga Limestone approximately 700 feet west of the north-south trending Clarksville Cave. This number reflects a *minimum* westerly displacement of beds above the fault ramp. Floodwaters remain perched, flowing down the apparent dip of the Schoharie Formation, until they encounter a fractured zone along a more steeply inclined part of the fault ramp. Here, subsurface water is deflected sharply to the south and aslant the strike and dip of the inclined fault plane, with the possible localized exception of following a horse for 200 feet north of the Lake Room.

Pirated surface water must rise at the Mill Pond, because the impermeable Esopus Shale is thrust vertically upward against the cavernous Onondaga Limestone. The leading edge of this upthrown shale formation, an imbricate thrust sheet separate from the fault zone that Clarksville Cave formed along, trends roughly north-south (Figure 1). The Esopus Shale and thrust-fault-induced fault-bend folds in the Onondaga Limestone, present slightly west of the upthrown Esopus, ultimately form a wall or barrier to easterly karstic groundwater flow for a distance that is well in excess of one-half mile.

However, this geologic barrier has only retarded eastern groundwater movement in two locations: (1) east of the known parts of the Waterfall Passage and (2) at the Mill Pond spring resurgence. Formation of most of the north-south oriented cave occurred preferentially aslant an inclined ramp of a thrust fault. Deformation along this thrust plane has produced a fault zone with at least three easily discerned, slickensided surfaces. Although separate, they occur within a few feet of one another. In much of the cave, this fault ramp is accented by one or more prominent calcite beds, often accompanied by a zone of stylolites and calcite-filled extension veins. This calcite bed is

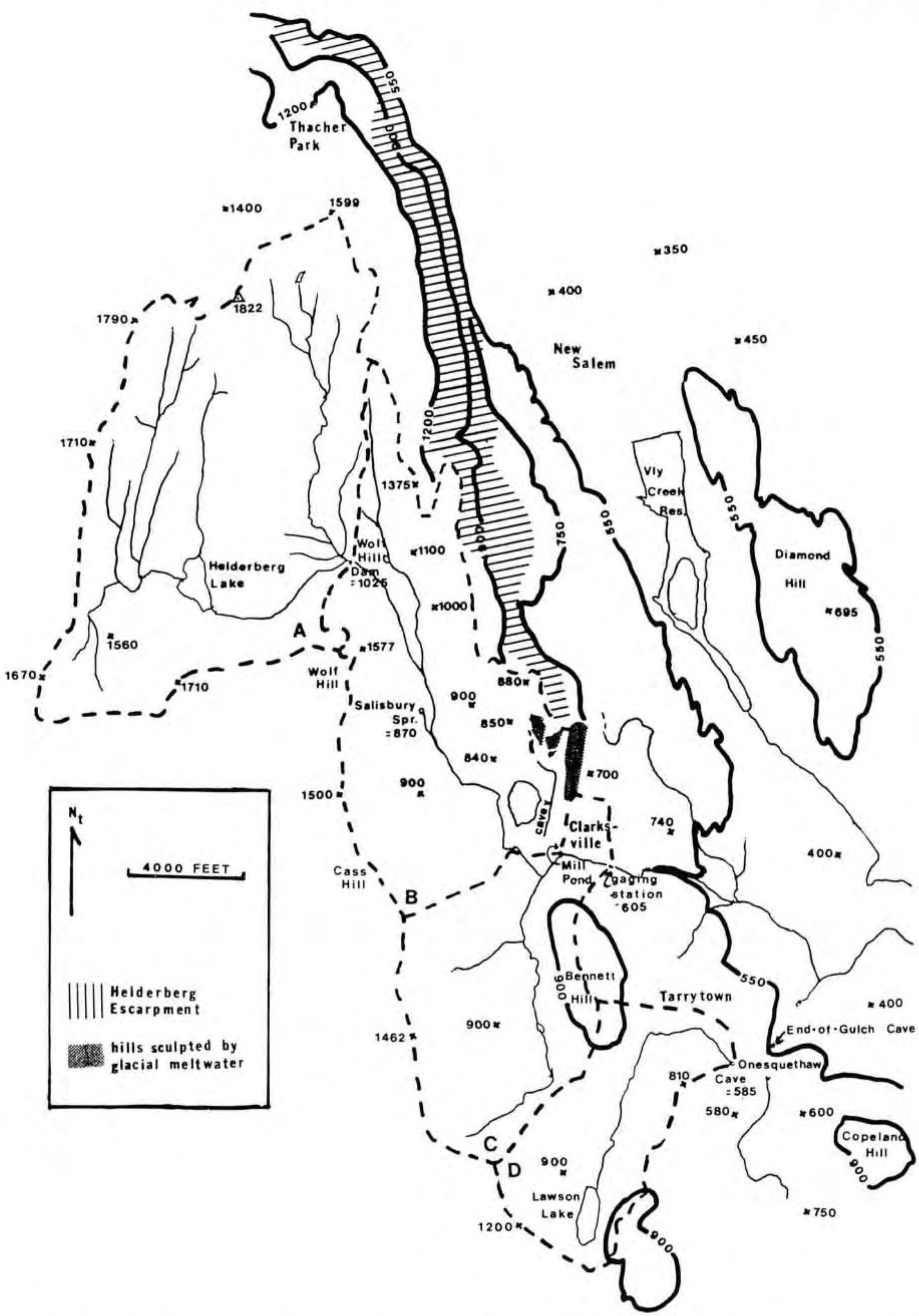

Figure 2: Topography, drainage basins, and selected features along the Helderberg Escarpment, Albany County, New York.

podiform in shape, with a central thickness ranging up to eight inches. Whereas the calcite forms a continuous bed of variable thickness aslant the strike of the thrust fault, it is most pronounced in the Gregory Section of the cave, where the vadose part of the cave is steeply inclined along the fault ramp. The ledge forming the outlet of the Waterfall Passage is the same buff brown to black, weathered Schoharie Formation, with underlying calcite bed, as seen in the upper Onesquethaw Creek gorge. The thicker zones of the calcite bed, where dissolution and crystallization are greatest, coincide with the more inclined segments of the thrust ramp, where the stress was highest. Occasional remnant calcite blocks, up to eight inches in thickness, in the Ward's Section of the cave provide the only evidence of the former presence of the thick calcite bed.

Ramsay (1980) provides evidence that similar "extension veins are formed by an accretionary process involving the formation of a narrow fracture followed by the filling of the open space by crystalline material, a mechanism termed crack-seal." Such stress-induced chemical transfer, or pressure solution of materials, seems to be relatively common (Ramsay, 1980). The characteristic crack-seal mechanism of repeated tectonic stress (Ramsay, 1980) is best illustrated in Onesquethaw Cave, situated 2 miles southeast of Clarksville, where calcite infilling aslant the ramp of a thrust fault reaches a maximum thickness of 27 inches. Here, insight into the fault style and repeated activation in the area is suggested by the presence of multiple calcite-vein infilling events along the fault ramp.

Successive cracks often occur along vein-matrix contacts of a previously sealed crack system, because this is mechanically the weakest surface in the rock (Ramsay, 1980). Fractures have been found to increase towards major faults. The higher the fracture frequency, the higher the percentage of calcite-filled fractures (Carrio-Schaffhauser and Gaviglio, 1990). It is a combination of this mechanically weaker calcite bed/Onondaga limestone boundary and related fault partings, all present along this inclined thrust ramp, that have served to orient the north-south segment of Clarksville Cave. Further structural and hydraulic control of cave-forming waters may also be locally attributed to perching on a fault-thinned Schoharie Formation. Similarly, much of Onesquethaw Cave has developed down and along the mechanically weaker vein-matrix contact. Both caves exhibit characteristic fault-bend folds, stylolites, and extension veins adjacent to the prominent thrust plane.

Of major importance to the development of both caves was Cenozoic structural deformation which provided a preferential solutional pathway along the inclined surface of a fault ramp. A steep hydraulic gradient was thus set up between infiltrating waters and their resurgence points along fault ramps. These faults may then be considered as both negative and positive influences on groundwater flow and cavern development: negative in the sense that downward dissolution did not readily penetrate far below the fault zone (Kastning, 1977 and 1984), and positive in the sense that almost the complete trend of the caves follows a structurally weakened zone of increased permeability.

Intermittent Phreatic Passages

Meltwater invasion of preglacial passages in Clarksville Cave occurred during glaciation, significantly enlarging the cave and its tributary conduits within the aquifer. Observation of the degree of flooding within the cave during major storm and runoff events reveals that only the lowest levels of the cave carry water along the fault zone. Two abandoned upper-level passages, both with relatively consistent ceiling elevations, were identified via a leveling survey. The level of these passages is determined by the relative uniformity of their ceiling heights. The highest of these two upper-level passages extends from the Lake Room, through the Big Room, until its truncation in the Pixie Passages immediately above the Corkscrew. (Note that a detailed map of Clarksville Cave is shown in Figure 3 of the following paper). This 714-foot level can roughly be characterized as the meandering upper level of Perry Avenue. In places, the lower ceiling elevation is controlled by chert beds. The 698-foot level extends from the Bathtub Passage through Upper Cook Avenue, where the passage is truncated by breakdown, and where flow had been diverted down a steeply dipping, fault-plane-controlled tube leading to Lower Cook Avenue.

These large upper-level passages are generally high and dry, and are sometimes tubular in cross-section, suggesting a water-table formation; they undulate upward and downward as is typical of phreatic passages, and have formed with only limited influence from the fault plane. However, they lack the low hydraulic gradient typical of phreatic or water-table origin, are discontinuous in size and extent, and truncate suddenly or become much reduced in cross-sectional area, grading systematically to active preglacial vadose passages. Relatively smaller drain size at the down-gradient end of these passages compared with the cross-sectional area of these floodwater conduits resulted in temporary phreatic conditions within the cave, quite dissimilar from the conditions of normal phreatic water. Palmer (1991) documents similar floodwater formation of conduits behind local passage impediments such as collapse debris, insoluble beds, or sediment fill, where aggressive water results in rapid passage enlargement. Formation of these "intermittent phreatic" floodwater conduits, coincident with the direct influx of large quantities of subglacial meltwater, occurred under alpine karst conditions.

A third solutionally developed upper level is also identifiable in the cave at 739 feet msl. This discontinuous level is represented in only a short segment of the cave proximal to the Root Room (northeast of the Lost Rockhammer Room) and some nearby domes. It clearly represents the maximum flood level attained in the cave. Solution domes at the 739-foot level are characteristic floodwater-injection features. Like the two lower abandoned levels, it is systematically graded to the actively forming

lowest level. These three abandoned levels are interpreted as reflective of different glacial discharges (perhaps seasonal fluctuations) coursing through the cave, rather than different time-based developmental stages. Variable discharges, perhaps influenced by variable outlet efficiencies and climatic conditions during glaciation, are inferred for formation of the levels.

There is no obvious mechanism present today that would explain both the elevation of these upper-level passages and their configuration, that exhibits little fault control. Other than the passage alignment, fault control appears to be a significant developmental factor only in the low-discharge lower-level passages. The strongest argument for a preglacial origin of the vadose-level passages in the cave stems from the recognition that the mean annual precipitation and the size of the Mill Pond watershed has probably not changed significantly in the last 10,000 years. Funk (1989), through the interpretation of archaeological sites, established that climatic conditions within the last 10,000 years were at times either dryer or similar to that of today. Thus, the availability of the significant discharges necessary to form the upper-level passages was not there postglacially.

These upper-level passages and related higher solution features on ceilings (upward to 739 feet) indicate that they are younger than many lower passages, having formed in response to aggressive glacial floodwaters behind an inefficient outlet. It is hypothesized that the lower-level Clarksville Cave passages served as a natural *in situ* drainage system for glacial meltwaters underneath a warm-based Wisconsinan ice sheet. Higher-level cave passages (*e.g.*, the 698- and 714-foot levels) formed in response to the massive influx of subglacial meltwaters behind inefficient, perhaps partially ice-blocked outlets. Similarly, the formation of high-ceiling solution domes, anastomoses, pendants, spongework-like dissolution, and diversion passages may be attributed to floodwater invasion. The gradation of upper-level conduits tributary to the lower levels of the cave, from relict meltwater infiltration points, also lends supportive evidence for a preglacial origin of the linear (N 12 E) Clarksville Cave passage.

Relict Karst

A number of relict karst features are present both proximal to Clarksville Cave and to the northwest within the same watershed. These include a number of small shafts and caves (*e.g.*, Trap Cave, North and Thook entrances) that receive only minor amounts of direct meteoric or snowmelt infiltration today. The most important relict karst feature is the Stove Pipe Paleogorge (Figure 1). This abandoned rockcut gorge grades directly into the preglacial Clarksville Cave via the North Entrance, Brown's Depression area, Trap Cave, and the Thook Entrance. Its channel is well defined for most of its course. The morphology of upper reaches of the gorge is characteristic of an ice-marginal meltwater channel with small-scale hanging valleys, rather than a well-graded stream bed which would be expected of a former channel of Onesquethaw Creek. Sugden and John (1976) describe the ice-flow dynamics which cause favorable formation of drainage routes in bedrock versus ice. In some places, the channel configuration is such that only large quantities of water would have been capable of filling the channel sufficiently high enough to overflow into sub-parallel channels. This paleogorge is sharply truncated to the north by a downwardly sloping limestone cliff. A negligible catchment area is present (Figure 2), certainly too small to carry any significant quantities of water or sediment into the cave as suggested by thick sediment banks and upper-level phreatic passages.

A second paleogorge, the Clarksville Paleogorge, proximal to Osborn Cave, may represent either a preglacial drainage route of the Onesquethaw Creek prior to glaciation or a channel carved around the Appleby Berm by glacial meltwaters. Gorges of this nature can form in a relatively short time if sufficient abrasive material is carried through it. Von Engeln (1911) documents the rapid formation of a rockcut marginal gorge at the outlet of the Hidden glacier in the Yakutat Bay Region of Alaska.

Brown's Depression (739 ft. msl) is an important location in that it received significant paleo-streamflow from the northwest (Figure 1). Stove Pipe Paleogorge streamflow, originating from a vast subglacial watershed to the north, incised a channel through the Onondaga Limestone from the northeast until it reached the Hunter's Fissure Cave and Diddly Cave area. Here, this paleo- streamflow was responsible for the formation of these caves. From Hunter's Fissure Cave the paleo-streamflow spread out to the southeast over the gently undulating topography. However, its course was partially constrained by the elevationally higher surface topography to the west, north and east. Thus, much of the Stove Pipe Paleogorge streamflow was funneled southeast into the Brown's Depression/ North Entrance (above Lake Room) area, where it entered Clarksville Cave.

During periods of low- to moderate-glacial discharge, meltwaters converged proximal to Brown's Depression where much of the southeastern discharge was retarded from flowing east by a low north-south trending limestone ridge (747 ft. msl). These meltwaters were pirated into Clarksville Cave through the Diddly Cave and Brown's Depression/North Entrance areas. Diddly Cave was recently dug open, increasing the length from 5 to 550 plus feet. A dive push, a short distance into the cave, led to a master conduit which may soon be linked to Clarksville Cave. The exposed limestone pavement, coupled with the steep hydraulic gradient present between insurgence and resurgence points, provided a favorable avenue for subsurface piracy.

High-discharge meltwater, with a glacial hydrostatic head, encountered the Clarksville Cave ridge and sought

the most efficient outlet, flowing southward and over the ridge barrier. However, the even higher surveyed elevations of a number of abandoned surface stream infiltration points and their conduits leading to Clarksville Cave provide information on the large magnitude of subglacial discharge necessary for their formation. Some of these features include Trap Cave (753 ft. msl), the Thook Entrance (766 ft. msl), and a deep solutionally enlarged joint near the Ward's Entrance (761 ft. msl). During periods of high glacial discharge, meltwaters probably flowed both within and outside the Stove Pipe Paleogorge channel. This meltwater splayed outward around, and possibly over, the Appleby Berm. The alignment of the Thook Entrance passages and the Pixie Passages suggests that meltwater coursing around both sides of the Appleby Berm sought to enter the preglacial Clarksville Cave through the most direct pathway. Meltwater thus entered the Clarksville and Stove Pipe paleogorges, encountering the Gregory Entrance to the cave, the Ward's Entrance, the Thook Entrance, Trap Cave, the North Entrance, Brown's Depression, and the jointed pavement above the cave.

Sediment Fill

Thick deposits of sediment in the cave provide direct evidence of the quantity of material carried down the Stove Pipe Paleogorge by glacial meltwaters. The point of entry of this material was largely through the Brown's Depression/North Entrance area. The finding of sediments in the newly discovered northwestern segment of the cave (Pauley Avenue) also argues for input via Hunter's Fissure and Diddly Cave. Thick remnant sediments reveal that at least the Ward's Section of the cave was once sediment filled. Because the physical opening of these sediment input points is believed to have been formed by subglacial meltwaters, a sedimentation, passage infilling, and re-excavation history may be constructed.

The thickness of the sedimentary column in contact with bedrock suggests that significant cave enlargement had occurred prior to sediment infilling, possibly during Illinoian and/or Kansan glaciations. The basal deposits on bedrock include imbricate shale-clast-rich sediments with small cobbles, indicative of rapid infilling. Once much of the cave became filled, floodwaters stagnated, leaving their signature in finely laminated sand, silt, and clay layers. These layers are seen near the ceiling in the Ward's Section and are interlayered with courser sediments in the Gregory Section (to at least the 714-foot level). At this time, bulk deposition of sediment is interpreted as occurring during early-Wisconsinan time, with re-excavation during late-Wisconsinan or post-glacial times. Lack of significant sediment cover and fill in paleo-channels and abandoned insurgence points supports this hypothesis. Limited deposition of sediment probably also occurred in the late Wisconsinan.

Partial plugging by sediment of Clarksville Cave's overflow outlets may have contributed to the degree of backflooding and upward passage development in the cave. Evidence is found for this in large glacial cobbles cemented in a clay matrix now terminating The Hidden Room. It is possible that sediments washing down the Clarksville Paleogorge partially or totally blocked the Gregory and Osborn Entrance overflow outlets for a period of time prior to being washed free again. Alternately, these outlets may have formed as a floodwater modification behind the inefficient Clarksville Cave low-flow outlet.

Glacial Geology

Two and possibly three glaciations are documented as far south as Corinth, New York (LaFleur, 1991, unpublished report). Approximately 14,700 years bp, the Wisconsinan ice sheet receded from the Helderberg Plateau (DeSimone and LaFleur, 1985). Dineen (1986) gives an extrapolated bog-bottom date of 15,060 ±1,000 year bp for the Great Bear Swamp situated somewhat west of Clarksville. This date further confirms the timing of the deglaciation of the Clarksville area. DeSimone and LaFleur (1985) provide a date of approximately 14,700 years bp for the recession of the Pine Swamp ice front from the Clarksville area. They depict the ice front as a lobe or tongue projecting southward to Stuyvesant, New York, with Clarksville situated along the southwestern flank of the ice margin.

Dineen (1986) documents ice thinning during glacial stagnation over the Helderberg Escarpment. Large quantities of meltwater flowed southward proximal to the southwestern flank of the Hudson Champlain Lobe of the Schoharie ice margin. Dineen describes deposition of sediments in multiple meltwater tunnels under stagnant ice. It thus appears that deglaciation from the Stove Pipe Road area was characterized by a southward thinning ice cover, with a southward meltwater flow direction. Free-surface flow was probably present in the paleogorge prior to the final retreat of the Wisconsinan ice sheet from the Clarksville area.

Implications of Wisconsinan Climates

Of even greater importance than the physical presence of the Stove Pipe Paleogorge is its relationship to Hunter's Fissure and Diddly Cave and the implication this has on interpretation of Wisconsinan, and possibly Illinoian and Kansan climate. Hunter's Fissure and Diddly Cave formed along the abandoned Stove Pipe Paleogorge. The presence of small scallop wavelengths in joint-controlled Diddly Cave indicates rapid streamflow along the base of the Wisconsinan ice sheet. Rounded stream cobbles in walking-sized passages in Diddly Cave provide clear evidence that a large stream once flowed through the paleogorge. The recent finding of bones of a varying hare and the extinct passenger pigeon within clay deposits in the cave may provide important scientific information on these species' recolonization during or following deglaciation, as well as dating of regional deglaciation. Additional field

work and radiocarbon dating are planned in this rare preservational location.

The Clarksville Paleogorge has probably not had a stream in it for the last 14,700 years, coincident with retreat of the Wisconsinan ice in the mid-Hudson Valley (DeSimone and LaFleur, 1985). Furthermore, Wisconsinan ice had retreated from the lower Hudson Valley 15,000 or 16,000 years ago (Connally and Sirkin, 1986; Dineen, 1986), with the ice front retreating to the St. Lawrence Valley by 13,000 years ago. Therefore, the maximum time frame for possible ice front stagnation in the Clarksville area during active deglaciation is on the order of 1,000 years.

Solutional cave formation will occur only where a pre-existing network of integrated openings connects the recharge and discharge areas (Palmer, 1991). This is a process that requires a minimum of 10,000 years (Palmer, 1984; Dreybrodt, 1987, 1990; Palmer, 1991) before passageways obtain sufficient size for human entry. Additional passage cross-sectional size requires additional time. A shorter time period, on the order of 5,000 years, may be possible depending on joint widths present in the bedrock prior to infiltration by glacial meltwaters. Thus, subglacial meltwaters apparently were not only responsible for the formation of Diddly Cave, but must have flowed for a minimum of 5,000 to 10,000 years in order for the cave to form. Because the maximum amount of time the retreating ice front could possibly have stayed in the Clarksville area was on the order of 1,000 years, it follows that Diddly Cave, Hunter's Fissure Cave, the Brown's Depression area and other southern infiltration points were receiving meltwater from below warm-based glacial ice for at least 5,000 to 10,000 years. It is likely that temperate climatic conditions were present during the early and later Wisconsinan.

Meltwater Features in Other New York State Caves

Many New York State caves need to be re-examined for evidence of glacial meltwater modification. Several caves in east-central New York exhibit features characteristic of meltwater invasion. Examples include Skull, Knox, Ella Armstrong, McFail's, Howe Caverns, Single X, Schoharie, Gage, Onesquethaw, and Surprise (Mystery) caves, all of which have one or more passage segments superposed above passages receiving Holocene peak floodwaters. For example, the upper levels of Skull Cave have aragonite speleothems that are incapable of surviving floodwater invasion. Similarly, caves such as Skull, Knox, and Ella Armstrong have watershed sizes too small to account for the volume of water necessary to form their observed vertical and areal extent. Other caves, such as McFail's and Surprise, carry underfit streams, yet exhibit anomalously large passage sizes. An artificially enlarged subglacial watershed would have been capable of providing the necessary recharge. Related fossil karst includes abandoned and sometimes glacial-debris-covered sinkholes and shafts which once served as infiltration points. Anomalous in-cave features such as abandoned pits and multiple-level, ungraded passages may also reflect meltwater invasion (*e.g.*, Surprise Cave). Similarly, significant cave development proximal to the headwaters of a drainage basin (*e.g.*, Gage Caverns and Phoebe Pit) may also reflect meltwater invasion from an expanded ice-sheet watershed. Other relict caves such as Knox, Salamander, several Saugerties-area caves, and Joralemon's (Engel, personal communication) are now abandoned and largely waterfree. Their derangement from active drainage patterns may portray development during a previous interglacial period or, more likely, may be a result of modification by glacial meltwaters.

The characterization of preglacial cave modification by glacial meltwater invasion poses many exciting geomorphic questions for researchers in New York State. Speleothem- and sediment-dating techniques may shed light on karstic evolution and modification through three or more glacial periods. A complete geomorphic interpretation must include an assessment of flow conditions and geologic features in a defined watershed, both on the surface and in the subsurface.

References

Carrio-Schaffhauser, E. and Gaviglio, P., 1990, Pressure solution and cementation stimulated by faulting in limestones: *Journal of Structural Geology*, v. 12, p. 987-994.

Cassie, R.M., 1990, Fault-related folding near Catskill, New York, *Northeastern Geology*, v. 12, no. 1, p. 19-27.

Connally, G.G. and Sirkin, L.A., 1986, Woodfordian ice margins, recessional events, and pollen stratigraphy of the mid-Hudson Valley, *in* Cadwell, D.H. (editor), The Wisconsinan Stage of the First Geological District, Eastern New York: *New York State Museum Bulletin 455*, p. 50-72.

DeSimone, D.J. and LaFleur, R.G., 1985, Glacial geology and history of the northern Hudson Basin, New York and Vermont, *in Field Trip Guidebook*: New York Geological Association, p. 82-115.

Dineen, R.J., 1986, Deglaciation of the Hudson Valley between Hyde Park and Albany, New York, *in* Cadwell, D.H. (editor), The Wisconsinan Stage of the First Geological District, Eastern New York: *New York State Museum Bulletin 455*, p. 89-108.

Dineen, R.J., 1987, Preglacial and postglacial drainage of the central Hudson Valley, *in Field Trip Guidebook*, New York State Geological Association, p. B-1 - B-30.

Dreybrodt, W., 1987, The kinetics of calcite dissolution and its consequences to karst evolution from the initial to the mature state: *National Speleological Society Bulletin*, v. 49, p. 31-49.

Dreybrodt, W., 1990, The role of dissolution kinetics in the development of karst aquifers in limestone: A model simulation of karst evolution: *Journal of Geology*, v. 98, no. 5, p. 639-655.

Funk, R.E., 1989, Some contributions of archaeology to the study of cave and rockshelter sediments: Examples from eastern New York: *Man in the Northeast*, no. 37, p. 35-112.

Geiser, P.A. and Engelder, T., 1983, The distribution of layer-parallel shortening fabrics in the Appalachian foreland of New York and Pennsylvania, *in* Hatcher, R.D., Jr.; Williams, H.; and Zietz, I. (editors), contributions to the tectonics and geophysics of mountain chains: *Geological Society of America Memoir 158*, p. 161-176.

Kastning, E.H., 1977, Faults as positive and negative influences on ground-water flow and conduit enlargement, *in* Dilamarter, R.R. and Csallany, S.C. (editors), *Hydrologic Problems in Karst Regions*: Western Kentucky University, Bowling Green, Kentucky, 481 p.

Kastning, E.H., 1984, Hydrogeomorphic evolution of karsted plateaus in response to regional tectonism *in* La Fleur, R.G. (editor), *Groundwater as a Geomorphic Agent*: Allen and Unwin, Inc., London, Boston, p. 351-382.

Marshak, S., 1986, Structure and tectonics of the Hudson Valley fold-thrust belt, New York: *Geological Society of America Bulletin*, v. 97, p. 354-368.

Marshak, S. and Engelder, T., 1985, Development of cleavage in limestones of a fold-thrust belt in eastern New York, *Journal of Structural Geology*, v. 7, p. 345-360.

Marshak, S., and Engelder, T., 1987, Exposures of the Hudson Valley Fold-Thrust Belt, west of Catskill, New York, *in Geological Society of America Centennial Field Guide*, Northestern Section, p. 123-128.

Marshak, S., and others, 1986, Structure of the Hudson Valley Fold-Thrust belt between Catskill and Kingston, New York, *in A Field Guide: Prepared for Geological Society of America, Northeast Section, 21st meeting, Kiamesha Lake, N.Y.*, 70 p.

Meyerhoff, E.A., 1972, Postorogenic development of the Appalachians: *Geological Society of America Bulletin*, v. 83, p. 1709-1728.

Murphy, P.J.; Bruno, T.L.; and Lanney, N.A., 1980, Decollement in the Hudson River Valley: *Geological Society of America Bulletin*, v. 91, Part I, p. 258-262, Part II, p. 1394-1415.

Palmer, A.N., 1984, Geomorphic interpretation of karst features, *in* La Fleur, R.G. (editor), *Groundwater as a Geomorphic Agent*: Allen and Unwin. Inc., London, Boston, p. 173-209.

Palmer, A.N., 1991, Origin and morphology of limestone caves: *Geological Society of America Bulletin*; v. 103, p. 1-21.

Ramsay, J.G., 1980, The crack-seal mechanism of rock deformation, *Nature*, v. 284, p. 135-139.

Sugden, D.E. and John, B.S., 1976, *Glaciers and Landscape*: Edward Arnold, London.

Von Engeln, O.D., 1911, Phenomena associated with glacier drainage and wastage, with especial reference to observations in the Yakutat Bay Region, Alaska, *Zeitschrift für Gletscherkunde*, v. 6, p. 104-150.

Plate C: Passage leading from the Bathtub to the Slickenside Block in Clarksville Cave, Albany County, New York (*see* Figure 1 of Rubin, this volume, page 103). The near-vertical fracture is a joint. The inclined fracture is a major thrust fault along which much of the cave has formed. Slickensides on the fault plane are readily visible nearby. Note that this passage has formed along the line of intersection of the fault and joint. These structures provided the initial avenues for groundwater flow resulting in later enlargement of the conduit to the dimensions visible here. View is to the northeast. *Photograph by Ernst H. Kastning.*

Flow Characteristics and Scallop-Forming Hydraulics within the Mill Pond Karst Basin, East-Central New York

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ABSTRACT

This is a study of the hydrology of Clarksville Cave and the headwaters of Onesquethaw Creek, situated in the hamlet of Clarksville, New York, specifically the Mill Pond karst basin. During most of the hydrologic year, water entering that part of the watershed that is downstream of the Wolf Hill Dam is pirated into the Onondaga Limestone. Tracer tests and in-cave stream gaging indicate that extreme conduit conditions are present in the aquifer, with a maximum water velocity on the order of 5.3 km/hr.

It has been hypothesized that a submerged conduit must lie covered by breakdown blocks at the cave's northern terminus. Having established a known peak flow, a modified version of the Darcy-Weisbach equation was used to accurately calculate the minimum diameter of this conduit. Knowledge of the structural geology throughout the watershed, coupled with a detailed leveling survey in the cave, permitted reasonable estimates to be made for the two unknowns in the equation. A submerged conduit was subsequently opened and explored.

Scalloped cave walls are present in Perry Avenue at a key stream gaging location. Backflooding occurs behind inefficient passage constrictions a short distance downstream of, but not up to, this station. Evidence exists that documents that only long return-interval flood stages cause backflooding to this station. This situation permits a reasonable estimate of the maximum discharge and flow velocity responsible for scallop formation. Scallop wavelengths were measured below the elevation of peak floodwaters. By inputting measured values for discharge and flow velocity into published equations, it was possible to back-calculate scallop Reynold's numbers that favorably correlate with measured flow velocities and discharges. A possible revision of the scallop Reynold's number is suggested when it is utilized in the determination of paleoflow velocities. It also appears that scallop wavelength is partially determined by the properties of the rock comprising the walls of the conduit.

Location and Watershed Boundaries

A broad karst aquifer is present in the Clarksville area. Its boundaries extend north and northwest of Mill Pond, situated less than 120 meters south of the restaurant, June's Place (*see* Figure 2 of the preceding article). The farthest boundary of the Mill Pond karst basin lies about 3.9 kilometers to the northwest, proximal to the Wolf Hill Dam on Onesquethaw Creek. The elevation of the basin ranges from 1822 feet msl atop the Helderberg escarpment to approximately 645 feet msl at the Mill Pond. The boundaries of the catchment basin are depicted in bold dashed lines. These boundaries were defined through the use of low-altitude stereo aerial-photography, U.S.G.S. topographic maps, tracer studies and, in places, detailed structural geologic mapping.

The Mill Pond watershed may be subdivided into two parts: A) that part of the watershed located upstream of Wolf Hill Dam (1,245 hectares), and B) that part of the watershed located downstream of Wolf Hill Dam (829 hectares). The downstream part of the Mill Pond watershed exhibits features characteristic of karst terranes. These include sinking streams, limited surface drainage, solutionally enlarged joints, sinkholes, and the Clarksville/Diddly cave system. Structural deformation throughout the region has resulted in extensive jointing and faulting, providing solutional pathways for infiltrating waters.

Piracy of Onesquethaw Creek Waters

Most of Onesquethaw Creek downstream from Wolf Hill Dam and upstream of Mill Pond is a losing stream, with a substantial amount of surface flow lost to solutionally enlarged joints in the stream bed. During most of the hydrologic year little or no surface flow occurs in the area downstream of Wolf Hill Dam, located nearly on the Marcellus Shale/Onondaga Limestone contact. Subsurface piracy of water into the Onondaga Limestone below Wolf Hill Dam occurs via numerous joints in the stream bed. The water briefly surfaces at Salisbury Spring, only to again sink into joints in the stream bed. The volume of water flowing in the stream bed and the relative efficiency of the often partially sediment-choked joints governs the distance water may be found flowing on the surface downstream from Wolf Hill Dam and the Salisbury Spring. The greater the discharge of the stream, the farther its flow is capable of traveling prior to complete subsurface piracy. During periods of low or moderate discharge, all Onesquethaw Creek surface flow is pirated into the karst network prior to where the bed of the Onesquethaw Creek passes beneath Rt. 443 (*see* Figure 1 of the preceding article). Only large storm and snowmelt events generate enough surface flow in the watershed to cause Onesquethaw Creek to flow throughout its course. This represents a very small part of the hydrologic year. Surface-stream flow is short-lived even after major storm events.

Tracer Tests

A series of uranine-tracer tests have permitted partial delineation under varying conditions of discharge of the subsurface flow paths throughout the Mill Pond drainage basin. Uranine is a non-toxic tracer frequently used in karst investigations (Smart, 1984). It was injected into various joints in the Onesquethaw Creek stream bed that were pirating water. Activated-carbon detection bugs were placed at all likely resurgence points, collected later, and chemically elutriated with Smart solution (Quinlan, 1986).

Tracer testing has revealed that dye injections from 3.2 km upstream of the Mill Pond resurgence remain perched above the Upper Cherty Subunit for at least 1.6 km before breaching chert beds that overlie the lower, more massive, non-cherty subunits. Water pirated into the Onesquethaw Creek stream bed immediately downstream of Wolf Hill Dam and upstream of Rt. 85 remains in one or more subsurface conduits, until surfacing briefly at Salisbury Spring, only to again sink into joints in the stream bed downstream. Salisbury Spring is located on the western side of Rt. 443, approximately 0.6 km southeast of Rt. 85. It is set back some distance from the road. It is likely that piracy of the Salisbury Spring discharge into the bed of Onesquethaw Creek is roughly coincident with the point at which this water breaches the Upper Cherty Subunit of the Onondaga Limestone. Thus, one major tributary conduit to the system is likely to become physically impassable within 1.4 km northwest of the Lake Room in Clarksville Cave (Figure 1). However, stream gaging and tracer studies indicate the presence of a second low-flow conduit entering the known parts of Clarksville Cave from the large, heavily jointed watershed to the north-northwest.

All subsurface flow resurges at the Mill Pond. The relative inefficiency of the outlet of the Mill Pond conduit may be due to structural problems resulting from the upward thrusting of the impermeable Esopus Shale against the cave-bearing Onondaga Limestone (*see* previous paper). The presence of impermeable Esopus Shale in the bed of Onesquethaw Creek at and immediately downstream of Mill Pond forces all subsurface flow from the karst aquifer to surface at Mill Pond. This author established a gaging station downstream of this point (*see* Figure 2 of the preceding article).

Tracer tests and discharge measurements throughout the watershed indicate that during periods of low discharge, pirated Onesquethaw Creek waters do not travel through Clarksville Cave. Surface and subsurface stream gaging and tracer tests establish the intersection of the pirated Onesquethaw Creek low-flow conduit with the Clarksville Cave low-flow drainage conduit to be located between the southern end of the cave and Mill Pond (*see* Figure 1 of the preceding article). Hereafter, the conduit that resurges at Mill Pond and is physically separate from the Clarksville Cave conduit north of Osborn Cave, is referred to as the Mill Pond low-flow conduit. Although the exact elevation of the lowest drain point in Clarksville Cave remains to be surveyed, it lies slightly below an elevation of 660 ft msl. The hypothesized flow routes of unentered parts of the network are portrayed in Figure 1 of the preceding article.

Tracer tests verify that after a certain critical subsurface discharge is reached, coincident with piracy of increasingly greater amounts of surface flow into the subsurface conduit system, the efficiency of the Mill Pond low-flow conduit is exceeded and surplus water is shunted to the Lake Room in Clarksville Cave. The Mill Pond low-flow conduit utilized today, which bypasses Clarksville Cave, may be the original flow route, with the flow route leading to the Lake Room (via Pauley Avenue) forming as a floodwater-overflow route. Alternatively, the flow path to the Lake Room may represent the original subsurface flow route that was later abandoned due to further stream piracy, possibly coincident with lowering of the regional base level. Under this genetic interpretation, diversion of waters from the Mill Pond low-flow conduit to Pauley Avenue would occur behind an immature drain. During periods of base flow, it appears that only water from the north-northwestern part of the Mill Pond watershed rises in the Lake Room. Moderate and high discharge in the subsurface causes a significant backup of water behind the Mill Pond low-flow conduit, resulting in large overflows to the Lake Room. The greater the flow in Onesquethaw Creek, the more water is lost through joints in the stream bed, and the greater is the discharge that appears in the Lake Room.

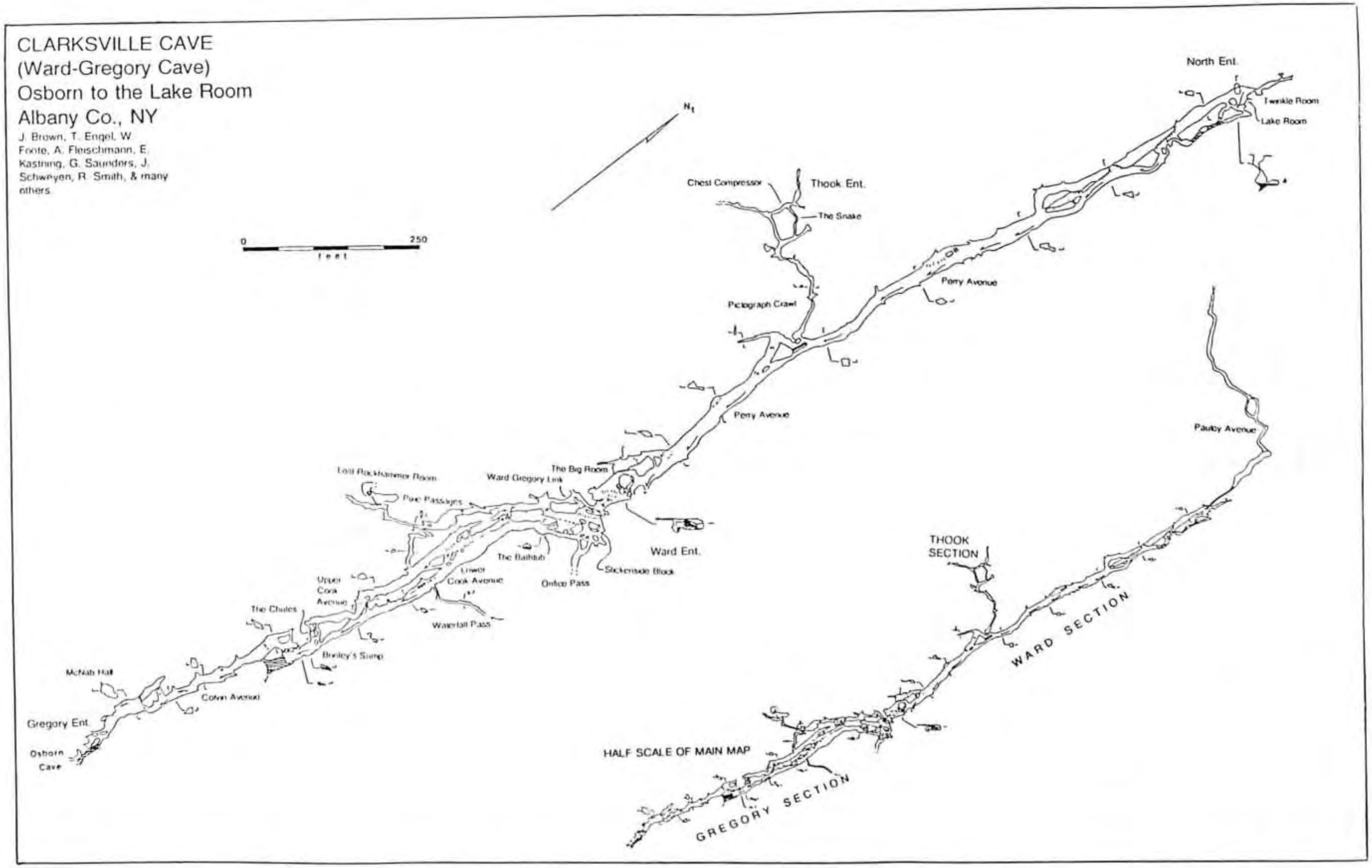

Figure 1: Map of Clarksville Cave, Albany County, New York. Notes: 1) This map is based on map of Kastning (1975). New passages added are Pauley Avenue, the Pinch Passage, the Thook Section, the Orifice Passage, and part of the Bathtub Feeder. 2) Stream flow is perennial, but varies from a low of 2 gpm to in excess of 60,000 gpm. 3) The Gregory Entrance, Brinley's Sump, and the stream access immediately east of McNab Hall are subject to flooding. During periods of extreme flood, the Bathtub and the downstream end of Perry Avenue also sump shut. 4) From the Lake Room south, the entire cave is Grade 5 except the Pinch Passage (grade unknown) and the Orifice Passage (Grade 2). Upstream of the Lake Room the underwater section is Grade 3; the rest as far as the Loop is Grade 5. Beyond the loop it is Grade 1.

From the Lake Room the water flows south through the cave where some of it joins, in a tributary manner, the Mill Pond low-flow conduit somewhere between the downstream end of the cave and Mill Pond (*see* Figure 1 of the preceding article). As the discharge of floodwaters within Clarksville Cave increases, the hydraulic efficiency of the branched conduit leading to Mill Pond is exceeded. The remaining water that cannot be handled by Clarksville Cave's low-flow subsurface conduit and the Mill Pond low-flow conduit backs up within the cave as temporary storage. After a critical flow on the order of 2.7 cfs is reached, excess floodwaters are discharged along the Osborn Cave overflow route (677 ft msl) to the surface. Osborn Cave is situated directly south of the Gregory entrance and is physically connected to Clarksville Cave by a water-filled conduit. Figure 1 of the preceding article shows the Osborn overflow channel, that sometimes carries large quantities of water.

Karst Basin Characterization

Tracer tests conducted in parts of the Mill Pond aquifer reveal that all subsurface waters reappear or resurge at Mill Pond. During periods of low flow, all surface and ground water downstream of Salisbury Spring and upstream of the bridge crossing Rt. 443 discharge through conduits in the karstic aquifer at Mill Pond. During periods of moderate- to high-subsurface discharge, part of the subsurface flow is shunted through Clarksville Cave. All flow throughout the Mill Pond watershed thus surfaces either in Mill Pond or Osborn Cave, where, for much of the year, it comprises the headwaters of continuous surface flow of Onesquethaw Creek. At times this flow is supplemented by water from the Clarksville South Road and western Bennett Hill Road sub-watersheds.

Water in Onesquethaw Creek, from that part of the watershed upstream of Wolf Hill Dam that is not artificially diverted to the Vly Creek Reservoir, also sinks into the subsurface downstream of Wolf Hill Dam. Much of the flow in the karst network originates as diffuse infiltration outside the Onesquethaw Creek corridor. Virtually all meteoric water and snowmelt contacting the heavily-jointed, generally thin-soil-mantled limestone pavement within the Mill Pond watershed is pirated into subterranean limestone conduits. Geologically, water entering the soluble Onondaga Limestone must stay within it, because it is underlain by approximately 1 m of the Schoharie Formation (a quartzitic limestone) and approximately 30 m of impermeable Esopus Shale.

Physically unentered segments of the conduit network may be envisioned as being similar to a tree, where all branches coalesce downstream toward the trunk. Palmer (1991) describes such branchwork caves as the most common type. Water infiltrating from different segments of the aquifer's recharge area converges as higher-order passages that decrease in number and generally increase in size in the downstream direction. It is likely that the large northwestern part of the Mill Pond aquifer is branchwork in nature, with many tributaries coalescing downstream toward larger, master passages. It is also likely that segments of the conduit system directly underlie the bed of Onesquethaw Creek, whereas others extend far to the northwest. Still other segments must enter from the northwest where runoff from the Marcellus and Hamilton beds of Wolf and Cass hills sinks near the Onondaga Limestone contact and is rapidly pirated into the system. Recently, exploration via the newly opened Diddly Cave entrance yielded approximately 0.5 km of large stream passages extending north into the Mill Pond karst basin. These passages area branchwork in character, and if connected to Clarksville Cave would bring the cave's length to greater than 2 km. The dashed lines on Figure 2 of the preceding article portray a simplified version of the hypothesized configuration of conduits in the eastern end of the system.

Subsurface Travel Times

The combined flow from stream losses and diffuse-fracture infiltration is documented as moving very rapidly through the karst system. Although effort has not been made to absolutely quantify the rate of subsurface flow in the aquifer, the timing of two tracer tests provides some insight on the situation. Under moderate flow conditions present on February 23, 1990, uranine tracer was injected into a joint in the bed of Onesquethaw Creek, 3.2 km northwest of Mill Pond. At this time, all surface flow in the upper reaches of Onesquethaw Creek was being pirated into this joint. Tracer-detection bugs were collected from Clarksville Cave at 4:00 p.m. on February 24, 1990, about 22 hours after the tracer injection. All were positive for uranine. Thus, a subsurface groundwater transit time in excess of 150 meters per hour was documented.

A similar trace was conducted in October 1988 under low-flow conditions. In this instance, the tracer injection and sinking of the stream occurred farther northwest than during the above trace. In this second test the tracer-detection bug was removed from a location proximal to Mill Pond 27 hours after tracer injection. After elutriation, the detection bug was positive for uranine. A subsurface groundwater transit time in excess of 120 meters per hour was documented for low-flow conditions.

In contrast, during a time of peak flow within the aquifer (March 15, 1986 at 1:45 a.m.), the discharge and velocity of flow within Clarksville Cave were measured. The velocity was recorded as 1.48 meters per second. This equates to 5,328 meters per hour (5.3 kms/hr) and may be considered as indicative of the peak velocity of potential groundwater movement within the aquifer and of extreme conduit conditions. At times of peak flow, groundwater may move from end to end through the karst aquifer, a distance of approximately 3.9 km, in less than one hour.

The rapid hydraulic response to significant precipitation or snowmelt within the watershed has been repeatedly

documented with stream hydrographs both in Clarksville Cave and in Onesquethaw Creek. Subsurface conduit flow in the Mill Pond aquifer is roughly analogous to open channel flow in a surface stream. A thin soil-moisture bank over much of the watershed's limestone pavement further permits rapid infiltration of meteoric waters and snowmelt, thus bolstering subsurface transit times. Flood pulses throughout the karstified system are flashy, providing evidence of mature conduit development. Rapid flow characteristics present within the 2,074-hectare Mill Pond watershed, especially that part downstream of Wolf Hill Dam, make it and Onesquethaw Creek extremely sensitive to infiltration of contaminants.

During much of the hydrologic year, discharge from Mill Pond acts as the sole source of water to the upper reaches of Onesquethaw Creek. During periods of base flow this discharge has been gaged at less than 0.1 cfs. The recent zoning of land central to the karst aquifer as rural commercial may have severe effects on both the aquifer and Onesquethaw Creek if untreated waste streams or septic infiltration are permitted (Rubin, 1990b).

In-Cave and Onesquethaw Creek Flow Calculations

Measurements of discharge and streamflow velocity have been made periodically in Clarksville Cave since 1983. Over 99% of the water flowing through Clarksville Cave rises in the Lake Room. This water has been gaged during both low and high flow at discharges ranging between 0.002 and 111 cfs. A maximum water depth of 63.5 cm was measured during the storm of March 15, 1986. Discussions with Ed Gregory revealed that the flood discharge component in the cave, associated with the 1938 failure of the Helderberg Lake Dam, was significantly greater than the above maximum-gaged amount. Gregory reported that floodwaters were ponded to an elevation of approximately 719 feet msl, a short distance down the entrance slope inside the Ward's Entrance. The elevation of the cave passage in upstream Perry Avenue, approximately 11 meters south of the Lake Room, lies between 715 and 708 feet msl, thus indicating that all of Perry Avenue was flooded during this event. Confirmation of this flood level, and possibly another in 1903, is manifested in a thick mud film covering historic names and dates chiseled near the passage ceiling.

A gaging station was established in Onesquethaw Creek (*see* Figure 2 of the preceding article) in order to examine the relationship between in-cave discharge and surface-watershed discharge. This was monitored twice daily for 15 months, more frequently during flood events, and periodically for 4 years thereafter during major runoff events. Stream discharge was gaged at 13 different stages. Curvilinear regression was then utilized to establish a series of multi-order equations that could be used to correlate stage height with discharge. The greatest discharge recorded for Onesquethaw Creek during the course of this study was approximately 1337 cfs. This occurred on March 15, 1986 at 3:00 am following heavy rains (~7.0 cm) on a 38-centimeter snow pack. Temperatures up to 40 F accompanied the coastal storm of March 13-14, 1986. Daily monitoring of stream stage in Clarksville Cave for the same 15-month period revealed that a direct correlation exists between this discharge and that in Onesquethaw Creek. Approximately 8 percent of flood-peak discharge in Onesquethaw Creek flows through Clarksville Cave.

Knowledge of expected flood-return intervals and their magnitude in the Mill Pond karst basin was found to be essential to both the interpretation of how abandoned upper-level passages in Clarksville Cave formed and an understanding of the dynamics controlling scallop formation. The limited data for statistical comparison among hydrologic years in the Mill Pond karst basin necessitated examination of another roughly comparable basin in order to assess flood-return intervals. The farthest headwater gaging station on Schoharie Creek at Prattsville was selected. Many inherent differences occur between the basins, notably elevation, geology, regolith thickness, size, and location. The Prattsville and Onesquethaw Creek gaging stations are approximately 48 kilometers apart. However, the Prattsville and Mill Pond watersheds are comparable under conditions of a saturated soil-moisture bank, high runoff, and similar storm systems. Eighty-two years of data were examined at the Prattsville, New York station.

A Log-Pearson-Type-III and Gumbel-distribution statistical comparison of historic peak flow of Schoharie Creek gaging data with this study's hydrograph information for Onesquethaw Creek indicates that the largest Onesquethaw Creek peak of record (March 15, 1986) has a return interval on the order of 30 to 47 years. This corresponds to a Prattsville hydrologic-year peak discharge of 54,900 cfs. Thus, if 40 years was the expected flood-return interval, 25 floods of this magnitude could be expected every 1000 years. Reconstruction of the 1903 peak discharge at Prattsville (approximately 63,000 cfs), the highest on record, further reveals that a cave discharge well in excess of 111 cfs may also have occurred in 1903. The 1903 discharge, three standard deviations greater than the mean-annual Prattsville peak flow, has a predicted flood-return interval on the order of 47 to 90 years. Although the magnitude of this flood was larger than the 1986 flood, it probably was not as great as during the dam-failure flood in 1938. These infrequent storm or runoff events reasonably represent a near-maximum quantity of water available in the watershed under ideal, thin-soil-mantled, rapid-infiltration conditions. Therefore, it is difficult to explain the "intermittent phreatic" upper-level passages in Clarksville Cave without a substantially greater quantity of water. A subglacially enlarged watershed, as discussed in the preceding article, appears to be the only viable explanation.

Lake Room Submerged Conduit

Measured, statistically predicted, and inferred (*e.g.*

mud-covered historic names and dates) high discharges rising from the Lake Room indicated that an obscured conduit was present that must be capable of transmitting large discharges. It was thus hypothesized (Rubin, 1989) that a submerged conduit must lie covered by breakdown blocks, below the water surface in the Lake Room. By making a number of reasonable assumptions, it was possible to calculate the minimum diameter of an assumed circular conduit capable of discharging a given flow. A modified version of the Darcy-Weisbach equation

$$r = \left(\frac{Q}{\pi \sqrt{\frac{4g}{f}} \sqrt{\frac{\Delta h}{L}}} \right)^{2/5}$$

was successfully utilized to examine the size of the, until recently, unentered upstream segments of cave conduit, north of known parts of the cave. Calculations were confined to a circular conduit capable of discharging between 111 and 222 cfs (Q). The latter value was considered a reasonable approximation for the 1938 dam-failure discharge. A friction factor (f) of 0.1 was used. The two unknowns in the equation were the change in head (elevation of water upstream of the lake versus the elevation of the lake, Δh) and the length of flooded passage upstream of the lake (L) during flood events. A wide range of values of 1.5 to 30 meters, and 6 to 1219 meters were tested, respectively, for these unknowns. Although some of the values tested were likely to be extreme in nature, they were selected based on knowledge of the structural geology within the watershed, coupled with a detailed leveling survey throughout the cave. It was believed possible that significant backflooding might be occurring behind the Lake Room breakdown.

Insertion in the modified Darcy-Weisbach equation of a reasonable range of values for the change in head and the length of flooded passage suggested that the minimum diameter of a circular-conduit tributary to the Lake Room is between 0.6 and 2.4 meters. Recent excavation and penetration of a formerly blocked and water-filled conduit extending north and west of the Lake Room verified the calculations (Rubin, 1990a). The length of the water-filled passage was found to be 61 meters. The actual Δh value is probably no more than 3 meters. The smallest diameter found in these newly discovered passages was 1.4 meters. Maximum cross-sectional area found in the approximately 366 meters of conduit beyond the Lake Room that have been entered thus far is on the order of 8 square meters. These passage dimensions attest to mature conduit development in the carbonate aquifer within the catchment basin.

The assumed friction factor of 0.1 was found to accurately reflect the flow conditions through the Lake Room breakdown. Approximately two vertical meters of clean-washed angular breakdown, interspersed with minor quantities of rounded glacial cobbles, were excavated. The heterogeneous mixture of breakdown blocks ranged in size from several centimeters in length, width and height to approximately one meter. The water's approach angle, toward the lake surface, rises at approximately 30 degrees for the last 3 meters before reaching an irregular constriction (1.2 meters by 0.5 meters) in the breakdown. Prior to the last 3 meters, the submerged conduit is generally horizontal. The maximum conduit depth below the surface of the lake was found to be approximately 4.3 meters.

Scallop-Forming Hydraulics

Phases of this study focused on defining the Mill Pond karst basin, the relationship between flow in the karst aquifer versus that in Clarksville Cave, the expected return interval of peak discharge inside and outside Clarksville Cave, and flow conditions peculiar to Clarksville Cave. Specifically, a range of stream discharges and velocities were measured in an air-filled segment of linear passage and rectangular cross section. Water depth was recorded, as well as scallop wavelengths within the zone of the 30- to 47-year flood-return interval.

Paleoflow information that researchers hope to reconstruct, based on scallop wavelengths and dimensions of an abandoned passage, is either empirically measured or reasonably constrained. Measurements in Clarksville Cave permitted a cave-specific evaluation of Blumberg and Curl's (1974) scallop Reynold's number.

One potential problem with characterization of the physical conditions under which scallops form is defining the discharge, or range thereof, responsible for scallop development. It was possible to define minimum and maximum discharge limits leading to scallop formation at a key stream-gaging location in Perry Avenue. Here, cave walls in a fossiliferous sparite are scalloped. Stream flow across the width of the cobble floor does not become deep enough, or of sufficient discharge to form scallops until the water is approximately 18 centimeters deep. The 30- to 47-year flood (111 cfs) of March 15, 1986 resulted in a stream depth of 63 centimeters, but decayed in 34 hours to 10 cfs with a stream depth of less than 13 centimeters.

Backflooding occurs behind inefficient passage constrictions at the Big Room, approximately 120 meters downstream of this same key stream-gaging location. The level of backflooded waters, as measured on March 15, 1986 at the stream's surface, was only 48 centimeters lower in elevation than ponded water at the Perry Avenue gaging station. A small additional discharge amount, such as that probable in 1903, or certainly in the 1938 flood, would have substantially reduced the stream velocity here and its ability to form scallops. Thus, the 111 cfs measured on March 15, 1986 represents a value that is close to the maximum possible for discharge capable of forming scallops at the gaging station. This situation allows for a reasonable estimate of the maximum discharge and flow velocity responsible for scallop formation.

Statistical analysis of Clarksville Cave flood-return intervals indicates that cave discharges in excess of two standard deviations about the mean-annual peak discharge may be sufficiently short-lived and of infrequent recurrence to form the observed scallops. The frequency of shorter-term flood intervals is greater, and perhaps it is these events which are recorded as scallops rather than very short duration, high-discharge long-return-interval floods. Although the shorter-return-interval floods are also relatively short-lived, it may be the combined contact time of water with bedrock of many similar magnitude floods that is of importance. The relationship among stream depth, discharge, and flood-return interval in the watershed, as partially indicated in the table below, suggests that scallop formation in Clarksville Cave may occur during flood intervals that range between one and two standard deviations of the mean-annual peak discharge.

Scallop wavelengths were measured below the elevation of peak floodwaters. By inputting measured discharge and flow velocity numbers into published equations, it was possible to back-calculate scallop Reynold's numbers that favorably correlated with measured flow velocities and discharges. This procedure involved measuring scallops, stream flow, and stream velocity and examining the likely range of scallop-forming conditions utilizing published equations. For the rectangular Perry Avenue conduit:

The Sauter mean was used to calculate mean scallop wavelengths of scallop groups within 63.5 cm of the cave floor:

$$\mathbf{L}_{32} = \frac{\sum li^3}{\sum li^2}$$

(Curl, 1974)

A range of in-cave flow conditions was examined. The three flow conditions presented in Table 1 bracket the minimum and maximum stream discharge and velocity believed to be responsible for scallop formation at the key Perry Avenue gaging station.

"The scallop Reynold's number, N^*_R based on friction velocity, is a universal constant for scallop formation and was determined from model experiments (Blumberg and Curl, 1974) to have the numerical value $N^*_R = 2200$" (White, 1988).

Scallop formation is controlled, in part, by a dimensionless Reynolds number:

$$N_R = \frac{\mathbf{v}\mathbf{L}_{32}\rho}{\eta}$$

where: $\mathbf{v}$ = mean velocity of fluid flowing past scallop in cm/sec
$\mathbf{L}$ = mean scallop length in cm
ρ = density of fluid ≈ 1.0 gm/cm for 5 C and 10 C
η = fluid viscosity ≈ 0.015 gm/cm/sec for 5 C and ≈ 0.013 gm/cm/sec for 10 C

Thus, examining the specific flow conditions in a), b), and c) above (*see* Table 1) using $\mathbf{L}_{32}$ = 7.49 cm and η = 0.015 gm/cm/sec., a range of site specific Reynold's numbers was obtained:

a) N_R = 34,254
b) N_R = 48,186
c) N_R = 73,801

Curl (1974) provides the limiting geometry for a rectangular cave passage:

$$N_R = N^*_R\left[2.5\left(\ln\frac{D}{2\mathbf{L}_{32}} - 1\right) + B_L\right]$$

By inserting the range of N_R's above into Curl's equation, we can examine N^*_R, the scallop Reynold's number based on a range of actual flow velocities:

a) N^*_R = 2,341
b) N^*_R = 3,293
c) N^*_R = 4,337

Blumberg and Curl (1974) derived a universal constant for the scallop Reynold's number, based on plaster model studies, of 2200. Based on this study, it appears that N^*_R may actually not be a constant, but instead may best be characterized by a range of values. These values, based on this cave-specific study of a rectangular conduit, appear to be from 1 to 2 times the accepted constant. Empirical observation of the flow dynamics in Clarksville Cave, coupled with a characterization of flood-return intervals within the catchment basin, suggest that a scallop Reynold's number on the order of 3300 might fit the cave-specific conditions. It is possible that constants in accepted equations may lead to an underestimate of paleo or recent flow velocities and discharges. Further studies of the actual hydrologic conditions in which scallops form are warranted. It should be noted that B_L, another constant in the Reynold's number equation (which deals with wall rough-

Channel Width (m)	Stream Depth (cm)	Discharge (cfs)	Velocity (cm/sec)
a) 3.3	18.3	14	68.6
b) 3.3	27.2	30	96.5
c) 3.3	63.4	111	147.8

Table 1: Three flow conditions in Perry Avenue, Clarksville Cave.

* The wall material subject to scallop formation may influence the value of the scallop Reynold's number. Different types of surfaces, like limestone, ice, and plaster, may respond different-ly to water scour.

ness) was accepted on face value.

Acknowledgments

Heartfelt thanks are extended to the many northeastern cavers who have contributed to the various activities associated with the study of the Clarksville/Diddly Cave system. Different aspects of the project have included stream gaging, tracer tests, photography, leveling, digging, diving, surveying, drafting, and wall scrubbing. Kevin Downey and Kevin Harris deserve special thanks for their many hours of appreciated cave photography. Clayton Pauley, now deceased, was a true friend with whom I spent many fine evenings leveling through Clarksville Cave's labyrinth. Special thanks go to Thom Engel, the unsung hero who is always there to help survey, stream gage, draft maps and formulate ideas. The study of the Mill Pond karst basin is northeastern caving at its finest.

References

Blumberg, P.N. and Curl, R.L., 1974, Experimental and theoretical studies of dissolution roughness: *Journal of Fluid Mechanics*, v. 75, p. 735-742.

Curl, R.L., 1974, Deducing flow velocity in cave conduits from scallops: *National Speleological Society Bulletin*, v. 36, no. 2, p. 1-5.

Kastning, E.H., 1975, Cavern development in the Helderberg Plateau, East-Central New York: *New York Cave Survey Bulletin 1*, 194 p.

Palmer, A.N., 1991, Origin and morphology of limestone caves: *Geological Society of America Bulletin*, v. 103, p. 1-21.

Quinlan, J.F., 1986, Water-tracing techniques in karst terranes, revised extract from *Practical Karst Hydrogeology, with Emphasis on Groundwater Monitoring*: National Water Well Association, Dublin, Ohio.

Rubin, P.A., 1989, Clarksville Cave - 1990: *The Northeastern Caver*, v. 20, no. 4, p. 122-125.

Rubin, P.A., 1990a, Breakthrough in Clarksville Cave: *The Northeastern Caver*, v. 21, no. 1, p. 3-8.

Rubin, P.A., 1990b, *Clarksville Cave and Mill Pond Watershed Critical Area Definition*: Report to the Town of New Scotland, 25 p.

Smart, P.L., 1984, A review of the toxicity of twelve fluorescent dyes used for water tracing: *National Speleological Society Bulletin*, v. 46, p. 21-33.

White, W.B., 1988, *Geomorphology and Hydrology of Karst Terrains*: Oxford University Press, New York, 464 p.

Replacement Mechanisms among Carbonates, Sulfates, and Silica in Karst Regions: Some Appalachian Examples

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ABSTRACT

Every carbonate rock formation contains examples of replacement among carbonate minerals, sulfates, and silica. Although the net geomorphic effect is rarely significant, porosity and permeability can be greatly modified in this way. Recognition of these processes is also a powerful tool for interpreting past geochemical conditions. Examples can be observed in many areas of the Appalachians: in caves, paleokarst zones, and carbonate oil reservoirs. Some mechanisms are well known, but their recognition in the geologic record is not. Replacement of sulfates by carbonates usually involves common-ion effects, *e.g.* where dissolution of gypsum by calcite-saturated water causes precipitation of calcite. Evidence for former evaporites includes sutured, lath-shaped, double-terminated and lozenge-shaped calcite crystals, scimitar and anastomotic veins, multi-stage breccias, boxwork, nodular or cauliform textures, and authigenic carbonate sediment. Dolomitization and dedolomitization depend on the relative solubilities of calcite vs. dolomite. Below 23 C, dolomite is more soluble than calcite, but the relationship is reversed at higher temperatures. The sluggish kinetics of dolomite near saturation make it unlikely for dedolomitization to be a major karst process, except in sulfate-rich solutions, which greatly boost the solubility of dolomite. Dedolomitization, recognized by scattered rhombs of calcite, is therefore another hint of former sulfates.

Silica easily replaces carbonates and sulfates. Carbonates and silica dissolve and precipitate under the opposite conditions: rising pH and temperatures increase the solubility of silica but decrease the solubility of limestone and dolomite. Silica replacement of either carbonate or sulfate minerals is usually very selective, so it is unlikely that such a broad-scale process as cooling of high-temperature fluids could be the main mechanism. Closed-system dissolution of carbonates isolated from carbon dioxide sources can raise the pH well above 9, allowing much silica to dissolve. Aeration, evaporation, and exposure to local acidity cause silica to precipitate. These conditions are common in aerated caves and in zones rich in sulfates or organics.

Introduction

Certain karst-related porosity and rock textures cannot be accounted for by simple dissolution of carbonates. Replacement of one mineral by another is responsible for many of these features, but evidence can be obscure. Only through recognition of diagnostic textures and mineral associations can these processes be reconstructed. Porosity is usually generated because of differential solubility of the various minerals or by changes in volume caused by contrasts in mineral density.

This paper outlines the common replacement processes in karst regions, describes some Appalachian examples, and summarizes the petrographic evidence for mineral replacement. The processes described here include: (1) dolomitization and dedolomitization; (2) replacement of gypsum or anhydrite by calcite, or vice versa; and (3) silicification of carbonate or sulfate minerals. All three are commonly associated with breccia and vuggy porosity caused by the recrystallization, dissolution, and movement of evaporites.

Field Data

Much of our recent field work has been in the western United States, where late Paleozoic carbonate rocks have

been modified by present or former evaporites. Concurrently we have encountered several examples of porosity and mineralization in the Appalachians, some of which clarify the mechanisms observed in the western examples, and some that would not be understood without the more extensive evidence from the West. Part of the western work is summarized in an article on the Black Hills caves of South Dakota (A. Palmer and M. Palmer, 1989). Field sites in the Appalachians are located in Figure 1 and are described below. Because of the consistent relationships between the replacement features and the geologic setting, it is possible to extend the ideas presented here to many other similar locations.

Camp's Gulf Cave, Tennessee

Camp's Gulf Cave, in Middle-Upper Mississippian carbonate rocks of the Cumberland Plateau, Van Buren County, Tennessee, contains unusually extensive cave breakdown. The collapse zones provided views of carbonate rock at distances from the original solutional cave wall ranging up to several tens of meters. Veins and nodules of white material have the appearance of evaporites, but instead they consist of fibrous quartz and calcite. Many of the veins are thick and either discordant to the limestone bedding along fractures or concordant to the bedding. Other veins are thin, with an interconnected "chicken-wire" pattern. Needle-shaped anhydrite inclusions are abundant in all of them, suggesting that they were once evaporite bodies that have been replaced. In places the veins and nodules consist entirely or partly of gypsum and/or anhydrite in various stages of replacement by the less soluble quartz

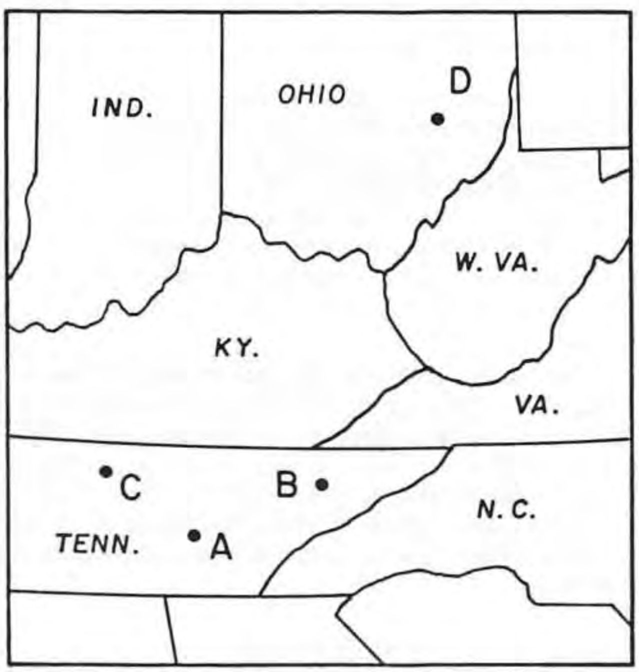


Figure 1: Location map of field areas described in this paper: A = Camp's Gulf Cave, B = Jefferson City, C = Elmwood, D = gas field of eastern Ohio.

and calcite. It appears that exposure to freely circulating groundwater has favored the replacement. We had previously observed nodular quartz veins in Mammoth Cave, Kentucky, but until the replacement process was actually caught in the act in Camp's Gulf Cave, their origin was just speculation.

Sulfide Zones in Carbonates of Eastern Tennessee

Zinc mines of eastern Tennessee (*e.g.* at Jefferson City) are located in early Ordovician limestones of the Knox Group in the vicinity of a lower Ordovician paleokarst. Similar mines in central Tennessee (*e.g.* at Elmwood), though not in the Appalachians, are in the same stratigraphic horizon and provide additional information. The limestone has been brecciated, dolomitized, and rendered highly porous. Sulfide minerals line many of the voids. Although sulfates are absent, breccias and bedrock contain many textures associated with vanished evaporites (M. Palmer and A. Palmer, 1989). In some respects they resemble those of the sub-paleokarst zones in the Madison Limestone of the northern Rockies.

There are two stages of breccia, only the later of which has been mineralized. Breccia bodies are interconnected, irregular, and discordant to the bedding of the carbonate rocks, with diameters on a scale of meters to many tens of meters. They resemble evaporite-induced breccia bodies in sub-paleokarst zones of the Madison of the northern Rockies (A. Palmer and M. Palmer, 1989). The later ore breccia is more extensive than the earlier one and follows roughly the same trends. Sphalerite and other sulfides in the later breccia are Mississippi Valley-type hydrothermal ores (Kyle, 1983; Ohle, 1985). The mineralized breccias post-date the paleokarst surface, extending both below and above it. Although the breccia texture resembles that of evaporite breccia (with bedrock clasts "floating" in a matrix of dolomite and other minerals), there is no concrete evidence that the late breccias were caused by evaporite-related processes. The mineralizing fluids caused solution and dolomitization of limestone and disaggregation ("sanding") of dolomite. Sulfide ore both replaces the bedrock and occupies pre-ore void spaces. Limestone in the early breccia has been partly replaced by quartz, some of which contains shrinkage cracks, that suggest an original amorphous phase. In contrast, certain euhedral quartz crystals are doubly terminated Herkimer "diamonds", and others are pseudomorphs after anhydrite needles and are oriented in flow patterns typical of former evaporites. Some of the early breccias have been dolomitized, especially those in the Valley and Ridge Province. Chert containing criss-crossing silica laths lines the bottoms of some ore zones in the mines. All of these characteristics are typical of former evaporite zones, suggesting that the breccias themselves are at least partly the result of evaporite disruption of carbonates by expansion, flow, and replacement. Unmineralized, bedded breccia in the Knox form extensive aquifers throughout many eastern and east-central states.

Gas Reservoirs in Ordovician Carbonates of Eastern Ohio

In 1987 we were contracted by Stone Resource and Energy Corp. (Worthington, Ohio) to investigate the origin and distribution of porosity in the Ordovician gas field of the Appalachian Plateau in eastern Ohio. The exact location and geological setting are proprietary. To supplement available information, a 10-cm-diameter core was drilled through the pay zone at a depth of 2053-2071 m. Porosity is limited to specific stratigraphic horizons and has textures diagnostic of former sulfates. Porosity consists of vugs, intercrystalline voids, and small-scale breccias. The vugs are cauliform, representing former evaporite nodules, and are partly filled with sanded dolomite. White nebulous masses of dolomite are surrounded by porous dolomite. Anhydrite, locally replaced by quartz, occludes some of the pores. Chips from other wells in the most productive zones consist of dolomite and authigenic quartz crystals containing tiny dolomite rhombs with ragged dissolved edges. The chips also contain large amounts of unoxidized pyrite.

Geochemistry

The porosity and rock textures in these field examples are the result of mineral replacements and selective removal and precipitation of minerals of contrasting solubility. Several mechanisms are responsible, all of which relate to the original geologic setting. By understanding the reactions involved, it is possible to interpret former geochemical environments.

In the following sections, geochemical relationships were obtained by a home-made iterative computer program (SOLEQUIL) designed to calculate carbonate equilibria in the presence of other solutes. It uses the equilibrium constants and their temperature dependence recommended by Plummer and Busenberg (1982) for calcium-carbonate solutions and by Langmuir (1971) for dolomite, and thermodynamic values given by Wagman and others (1982) for other chemical species. The extended Debye-Huckel equation is used to correct for activity-concentration relationships. All relevant ion pairs are included in the calculations.

Relative Solubility of Calcite and Dolomite

Calcite and dolomite have approximately equal solubility in groundwater, but the exact relationship depends on the way in which solubility is measured. As shown in Table 1, if calcite and dolomite are dissolved independently at temperatures less than 60 C, dolomite has a lower molar solubility but a larger solubility in terms of mass per volume. Dolomite is also volumetrically more soluble than calcite below 40 C. The saturation values for dolomite are rather approximate because of the difficulty of achieving dolomite equilibrium in the lab.

The ions liberated by dissolution of dolomite include those of calcite. Therefore, calcite approaches saturation as dolomite dissolves, even in the absence of solid calcite. Below 20 C, calcite becomes supersaturated as dolomite approaches saturation. Pure solutions of dolomite are rare. Natural water almost always encounters a combination of limestone and dolomite, so the degree of calcite saturation is generally higher than in either rock alone. Karst groundwater is typically less saturated with dolomite than with calcite (Thrailkill, 1977).

As temperature increases, both dolomite and calcite become less soluble. Dolomite solubility decreases at a faster rate, so the opportunity for dolomitization is enhanced at high temperatures, particularly since the normally sluggish kinetics of dolomite precipitation would be speeded. Dolomitization and dedolomitization involve the following reaction:

$$2CaCO_3 + Mg^{++} \Longleftrightarrow CaMg(CO_3)_2 + Ca^{++} \quad (1)$$

with an equilibrium constant, equivalent to $(Ca^{++})/(Mg^{++})$, of roughly 1.67 at 25 C. This equation does not apply if the water is undersaturated with the replacing mineral. Replacement of one mineral by the other is often achieved simply through dissolution of one and later precipitation of the other. In such a case the process is not considered true dolomitization or dedolomitization.

Although dolomite and calcite dissolve at roughly the same rate at low saturation ratios, the dissolution rate of dolomite drops dramatically below that of calcite as saturation is approached (Herman and White, 1985). Because dolomite dissolves so slowly near saturation, dedolomitization must be very slow or negligible in carbonate waters, and absent entirely at temperatures above 20 C.

	Calcite			Dolomite			
Temp (C)	mmol/L	mg/L	cc/L	mmol/L	mg/L	cc/L	$SI_{calcite}$
0	2.34	234	0.0864	1.6	291	0.10	+0.10
10	1.94	194	0.0717	1.3	230	0.081	+0.06
20	1.61	161	0.0595	0.99	180	0.064	+0.01
40	1.11	111	0.0408	0.66	120	0.043	-0.08

Table 1: Relative solubility of calcite and dolomite at $P_{CO_2} = 0.01$ atm. Each mineral is considered independently, with no mixture of the two. The values for dolomite are less accurate than those for calcite. The saturation index of calcite is shown at dolomite saturation. The saturation concentrations of calcite and dolomite increase greatly with P_{CO_2}, but the calcite SI values do not change significantly.

When waters at or near saturation with respect to calcite encounter gypsum or anhydrite, the saturation ratio of calcite increases because of the presence of the common ion Ca^{++} (Figure 2). Calcite usually precipitates as gypsum dissolves.

From Table 1, it appears that dedolomitization (simultaneous replacement of dolomite by calcite) is favored by low temperatures, with dolomite dissolving incongruently as calcite precipitates (*cf.* Evamy, 1967). However, in the presence of calcium sulfate, dolomite is much more soluble than calcite at all temperatures (Figure 3). Mg^{++} is depleted because of its strong affinity for sulfate ions, forming the ion pair $MgSO_4^0$. Ca^{++} is also depleted by forming $CaSO_4^0$, but not enough to offset the increase in Ca^{++} from gypsum or anhydrite. Therefore the saturation index of calcite rises much more rapidly than that of dolomite as gypsum or anhydrite is added to the solution.

Dedolomitization is more likely to occur in the presence of dissolved sulfates and at low temperatures, as shown by the contrast in solubilities between dolomite and calcite in Figure 3. The effect of calcium sulfate on dedolomitization has been noted previously by Yanat'eva (1955) and DeGroodt (1967) but attributed to the high Ca/Mg ratio forcing calcite to replace dolomite, as in Equation 1.

Open-System vs. Closed-System Carbonate Dissolution

Where infiltrating water passes through soil directly into carbonate rocks, the water approaches saturation with respect to the bedrock at or near the high P_{CO_2} values of the soil. Equilibrium pH is around 7.5-8.5, P_{CO_2} is high, and $CO_3^=$ activity is low. If, on the other hand, the water passes through a permeable but insoluble rock such as

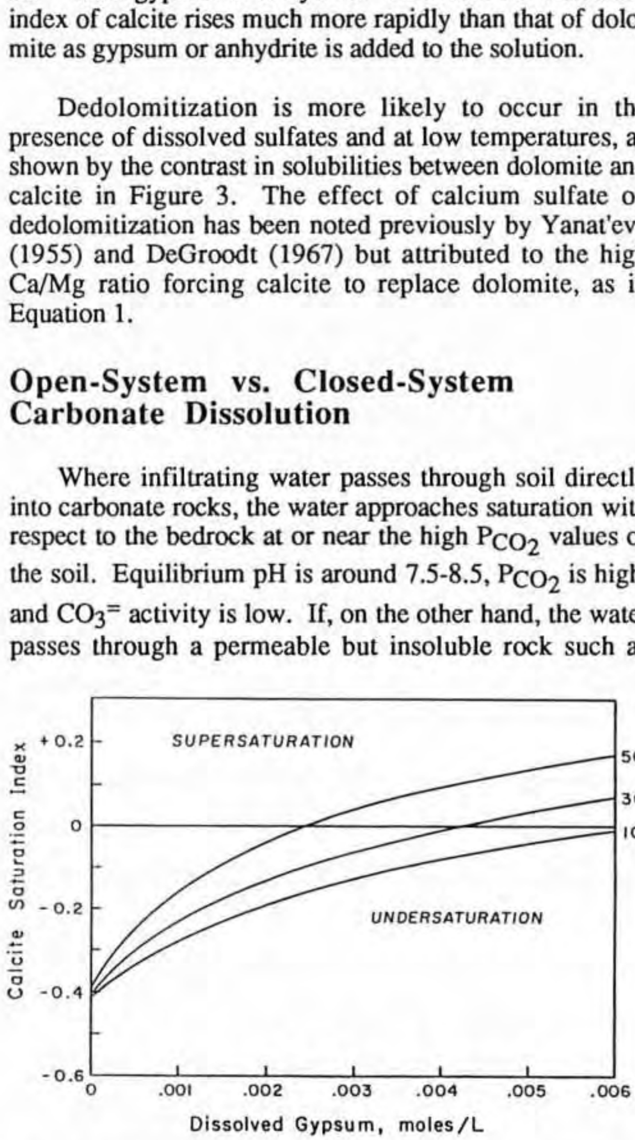


Figure 2: Increase in calcite saturation caused by dissolved gypsum (or anhydrite). Initial solution is 70% saturated with calcite at zero sulfate concentration. Saturation index = log (IAP/K).

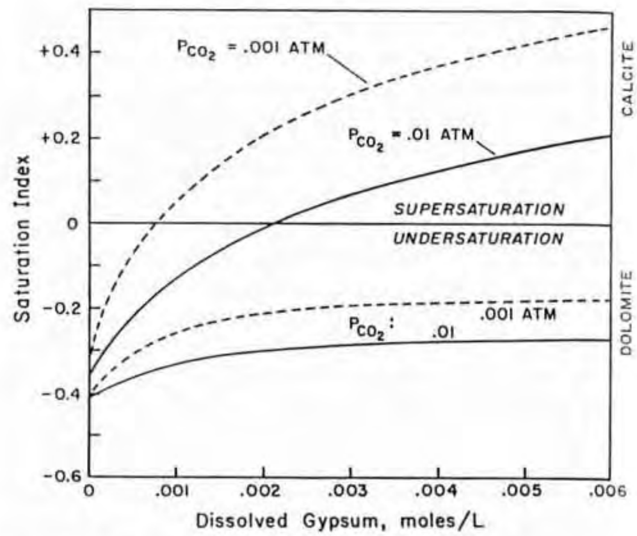


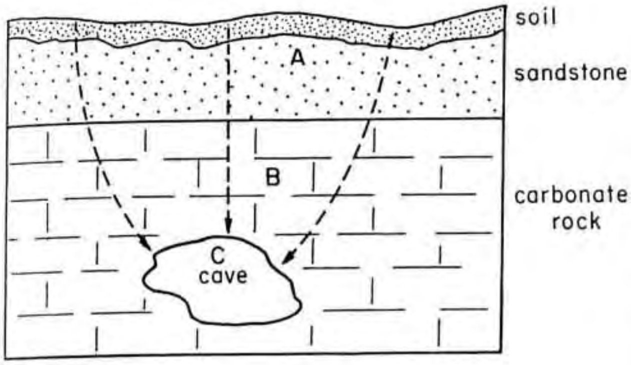
Figure 3: Contrast between calcite and dolomite saturation caused by dissolved gypsum (or anhydrite) in a solution initially 70% saturated with dolomite. Ionic ingredients of calcite are contributed by the dissolution of dolomite. Saturation index = log (IAP/K) for calcite and log $(IAP/K)^{0.5}$ for dolomite (to make the two SI values comparable).

sandstone before it encounters the carbonate rock, the water is effectively cut off from its original CO_2 source and dissolves the carbonate rock in a system nearly closed to CO_2 (A. Palmer, 1987). Equilibrium pH can rise above 9, while P_{CO_2} drops nearly to zero (Figure 4). The $CO_3^=$ concentration is an order of magnitude greater than in the open system. This effect is reduced if the water becomes partly saturated with carbonate before it reaches the sandstone.

If the water then reaches an aerated zone such as a cave or emerges at the surface, it encounters a P_{CO_2} higher than the negligible values shown in Figure 4. CO_2 is rapidly absorbed, and there is a sudden drop in both pH and (perhaps surprisingly) $CO_3^=$. The effects of these changes are described in the next two sections.

Replacement of Calcite and Gypsum by Silica

The solubility of silica (either crystalline or amorphous) accelerates greatly as the pH rises above 9 (*see* Drever, 1982, p. 92). The high pH values and availability of silica where water infiltrates through alternating sandstones and carbonates is favorable to dissolution of much SiO_2. When the water encounters an aerated zone such as a cave, or a zone of locally low pH, such as in an organic or sulfate/sulfide-rich zone, silica precipitates (Figure 4). This process is most effective at low initial P_{CO_2}, which would be typical of the loose sandy soils that develop on sandstone. M. Palmer (1986) has observed replacement of cal-



1. If carbonate rock = pure limestone:

	A	B	C
P_{CO_2}	0.001	9×10^{-7}	0.0005 atm
pH	5.37	9.9	8.17
C_s quartz	3.6	5.9	3.6 mg/L
$CO_3^=$	0.0066	0.0101	1.4×10^{-5} millimole/L

2. If carbonate rock = pure dolomite:

	A	B	C
P_{CO_2}	0.001	5.4×10^{-7}	0.0005 atm
pH	5.37	10.14	8.29
C_s quartz	3.6	7.5	3.6 mg/L
$CO_3^=$	0.0066	0.069	1.4×10^{-5} millimole/L

Figure 4: Chemical evolution of diffuse vadose water in quasi-closed conditions with respect to CO_2 at 10 C. Water passes through sandstone, becoming isolated from CO_2 in the soil, then encounters carbonate rocks. Capillary and gravitational seepage into a dry cave causes a sudden increase in CO_2. P_{CO_2} of soil and cave is assumed to be 0.001 and 0.0005 atm (typical of caves in such settings). Saturation with carbonate bedrock is assumed by the time the water reaches point C. C_s = saturation concentration. Amorphous silica behaves in a similar manner to quartz, but with a higher C_s. To achieve saturation at B, quartz must be available as insolubles within the carbonate, or carbonates must be interbedded with the sandstone below A.

cite in small amounts by silica in weathering rinds in the walls of certain caves overlain by quartzose sandstone.

Silica solubility also increases with temperature and therefore, if present, silica may be precipitated by rising, cooling groundwater. Precipitation should be more widespread than by the mechanism described above and would not necessarily depend on local changes in pH or on replacement of other minerals.

Carbonate/Sulfate Replacement

Gypsum and calcite are able to replace each other, depending on the $(CO_3^=)/(SO_4^=)$ ratio:

$$CaSO_4 \cdot 2H_2O + CO_3^= \Longleftrightarrow CaCO_3 + SO_4 + 2H_2O \quad (2)$$

This reaction has an equilibrium constant of $10^{-3.85}$ at 25 C (*see also* White, 1976). If $(CO_3^=)/(SO_4^=) > 10^{-3.85}$, calcite will tend to replace gypsum. The opposite holds true at ratios less than $10^{-3.85}$. A similar equilibrium can be stated for anhydrite and calcite, with an equilibrium constant of $10^{-4.0}$ at 25 C. These statements are valid only if the supposed replacing mineral is supersaturated.

Where water reaches equilibrium with dissolved carbonates in the quasi-closed conditions described in the previous section, and then encounters sulfates, the great $CO_3^=$ activity helps to keep the $(CO_3^=)/(SO_4^=)$ ratio high, which favors the replacement of gypsum or anhydrite by calcite (A. Palmer, 1987). Dissolution of gypsum or anhydrite by a carbonate-rich water would boost the Ca^{++} activity enough to precipitate calcite by the common-ion effect alone. The replacement would not take place on the molecule-by-molecule basis suggested by Equation 2.

Where the water emerges into an aerated zone, $CO_3^=$ decreases rapidly, and the $(CO_3^=)/(SO_4^=)$ ratio plummets. As a result, the ratio may easily drop below the threshold for calcite replacement, causing gypsum to replace calcite (A. Palmer, 1987). This is almost always valid only where evaporation is sufficient to drive gypsum to supersaturation. Most gypsum crusts on cave walls represent an actual replacement of calcite, rather than simple deposits of gypsum on the limestone surface.

Petrography

Recognition of replacement textures is best accomplished by microscopic examination. The following paragraphs outline some of the most diagnostic clues to the occurrence of carbonate-sulfate-silica replacement.

Dolomite bedrock is generally considered to be large-scale replacement of pre-existing limestone. In the early stages, euhedral dolomite crystals replace calcite in a scattered manner and are sprinkled throughout the limestone. The crystal shapes of the dolomite are independent of the crystals being replaced and do not produce pseudomorphs of the replaced mineral. Extensive dolomitization destroys most of the original sedimentary textures of the original limestone. Sharply defined, scattered rhombs are nearly always dolomite rather than calcite. Evaporites commonly

occur in dolomite bedrock because dolomitization also occurs in evaporitic marine environments.

Dedolomitization leaves the edges of dolomite crystals shredded and irregular, often with selective replacement of concentric zones within the crystals. Common byproducts include dolomite sand (where the process has not progressed far), pink calcified clumps, and iron-oxide residues. Evidence for former gypsum or anhydrite is common, as dedolomitization is most vigorous in a sulfate-rich environment. In some cases dolomite is replaced by gypsum, which is later replaced by calcite. Quartz may replace gypsum within or between the original dolomite crystals, and when the carbonate or sulfate minerals are dissolved, the residue is very porous, forming a friable mesh of quartz. This situation is typical of the bedrock in boxwork zones.

Quartz is a great mimic, taking on pseudomorphic textures of the minerals it replaces. Replacement of anhydrite crystals often produces needle-shaped or lath-shaped quartz crystals. This provides the most conclusive evidence for former evaporites. Early quartz replacement of bedrock is common in arid meteoric environments, and so quartz is often found replacing evaporite minerals.

In meteoric water, calcite typically replaces evaporites. Euhedral pseudomorphs are rare, but monoclinic and lozenge-shaped crystals of calcite appear to mimic former evaporite crystals. Calcite more commonly replaces large areas of former evaporites, forming sheets of sutured spar with banded inclusions of insoluble residue and wedges of bedrock, floating dedolomite and dolomite sand. Clasts appear to "float," and calcite fills anastomotic veins that have been repeatedly fractured. Calcite eventually replaces evaporite-cemented breccias and former evaporite nodules.

Gypsum and anhydrite are so soluble that they usually disappear entirely in near-surface humid environments. Their former presence must be inferred from relict textures such as those in Figure 5 (M. Palmer, 1988). The most diagnostic textures are: pseudomorphs of evaporites after quartz or calcite, lozenge-shaped calcite crystals, doubly terminated calcite and quartz crystals (Herkimer diamonds), breccia that includes upward-wedged clasts, crackle breccia (in which the fragments have barely moved relative to each other), fossil filaments of iron-fixing bacteria, multi-stage calcite veining, scimitar-shaped and anastomotic veins, boxwork, dedolomitization, sutured-clast contacts, cauliform and nodular textures, and authigenic carbonate sediment. No single feature is conclusive, but in combination their evidence is overwhelming.

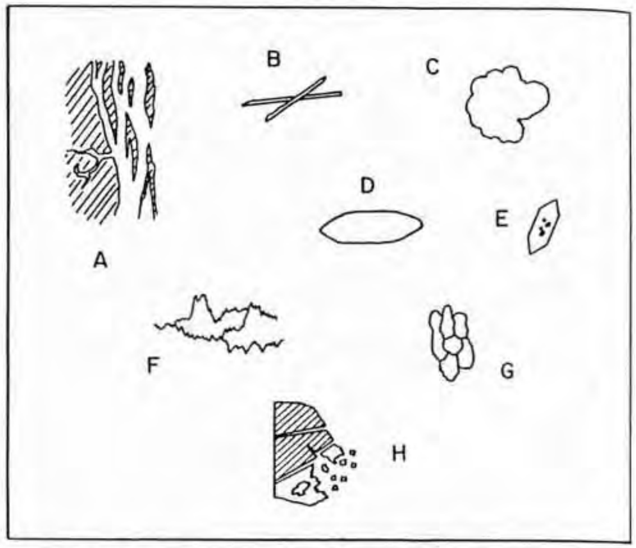


Figure 5: Small-scale and microscopic textures associated with former evaporites now replaced by other minerals. Scales are approximate. A = Anastomotic calcite veins, wedge-shaped slivers of bedrock and semi-circular cracks (width = 1 cm); B = needles and lath-shaped pseudomorphs of calcite or quartz after anhydrite (length = 100 microns); C = nodular or cauliform textures, (width = 2 cm); D = lozenge-shaped calcite crystals (length = 500 microns); E = doubly terminated quartz crystals with inclusions (length = 200 microns); F = sutured stylolitic patterns (length = 3 cm); G = "chicken-wire" anhydrite texture (width = 1 cm); H = dolomite clasts (shaded) assimilated by poikilotopic calcite spar after gypsum. Ragged bedrock remnants and dolomite sand converted to dedolomite and assimilated by spar (width = 1 mm).

Conclusions

Much of the apparently karst-related porosity and mineralization at depth in the Appalachians is actually related to former evaporites. Observations of mineral replacement and differential solubility help to interpret the origin and distribution of deep porosity zones. It also allows more accurate projection of pay zones, interpretation of the local geochemical history, and stratigraphic correlation. If replacement is not recognized, the original rock composition cannot be determined.

References

DeGroodt, K., 1967, Experimental dedolomitization: *Journal of Sedimentary Petrology*, v. 37, p. 1216-1220.

Drever, J.I., 1982, *Geochemistry of Natural Waters*: Prentice-Hall, Englewood Cliffs, NJ, 399 p.

Evamy, B.D., 1967, Dedolomitization and the development of rhombohedral pores in limestones: *Journal of Sedimentary Petrology*: v. 37, p. 1204-1215.

Herman, J.S. and White, W.B., 1985, Dissolution kinetics of dolomite: Effects of lithology and fluid flow velocity: *Geochimica et Cosmochimica Acta*, v. 49, p. 2017-2026.

Kyle, J.R., 1983, Economic aspect of subaerial carbonates, *in* Scholle, P.A.; Bebout, D.G.; and Moore, C.H.

(editors.), *Carbonate Depositional Environments*: American Association of Petroleum Geologists, Tulsa, p. 73-92.

Langmuir, D., 1971, The geochemistry of some carbonate ground waters in central Pennsylvania: *Geochimica et Cosmochimica Acta*, v. 35, p. 1023-1045.

Ohle, E.L., 1985, Breccias in Mississippi Valley-type deposits: *Economic Geology*, v. 80, p. 1736-1752.

Palmer, A.N., 1987, Gypsum replacement by alternating open and closed systems in the vadose zone: *Cave Research Foundation, 1986 Annual Report*, St. Louis, p. 27-28.

Palmer, A.N. and Palmer, M.V., 1989, Geologic history of the Black Hills caves, South Dakota: *National Speleological Society Bulletin*, v. 51, p. 72-99.

Palmer, M.V., 1986, Siliceous crusts on cave walls (abstract): *National Speleological Society Bulletin*, v. 48, p. 39.

Palmer, M.V., 1988, Influence of former evaporites on South Dakota caves (abstract): *National Speleological Society, Program of 1988 Annual Convention, Hot Springs, SD*, p. 31.

Palmer, M.V. and Palmer, A.N., 1989, Paleokarst of the United States, *in* Bosak, P.; Ford, D.C.; and Horacek, I. (editors), *Paleokarst: A Systematic and Regional Review*: Prague, Elsevier and Academia, Amsterdam and Prague, p. 337-363.

Plummer, L.N. and Busenberg, E., 1982, The solubilities of calcite, aragonite, and vaterite in CO_2-H_2O solutions between 0^o and 90^oC and an evaluation of the aqueous model for the system $CaCO_3$-CO_2-H_2O: *Geochimica et Cosmochimica Acta*, v. 46, p. 1011-1040.

Thrailkill, J., 1977, Relative solubilities of calcite and dolomite, *in* Tolson, J.S. and Doyle, F.L. (editors.), *Karst Hydrogeology*, International Association of Hydrogeologists, 12th Memoirs, p. 491-500.

Wagman, D.D. and 7 others, 1982, The NBS tables of chemical thermodynamic properties: Selected values for inorganic and C1 and C2 organic substances in SI units: *Journal of Physical Chemistry, Reference Data 11, Supplement 2*, p. 1-392.

White, W.B., 1976, Cave minerals and speleothems, *in* Ford, T.D. and Cullingford, C.H.D. (editors), *The Science of Speleology*: Academic Press, London, p. 267-327.

Yanat'eva, O.K., 1955, Effect of aqueous solutions of gypsum on dolomite in the presence of carbon dioxide: *Akad. Nauk. SSSR, Doklady*, v. 101, p. 911-912.

Fracture Controls on Groundwater Flow and Cave Development in Northern Greenbrier and Southern Pocahontas Counties, West Virginia

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ABSTRACT

Fractures fulfill a number of functions as structural controls. They guide initial flow paths; promote the development of blind solution pockets, joint spurs, fissures, or anastomoses along the perimeters of existing conduits; and promote *in situ* disaggregation of bedrock. By providing zones of structural weakness, fractures also promote collapse, which modifies passage morphology and leads to enlargement of conduits if breakdown is removed.

In northern Greenbrier and southern Pocahontas counties, West Virginia, bedding plane partings guide the majority of early flow paths. Most prominent guiding bed partings have anastomoses and are at contacts between argillaceous lower units and purer overlying units. N 60-75 E set joints, generally in *en echelon* zones, guide early flow paths in the Union Limestone, particularly in settings subject to floodwater influences, where the joints also guide blind joint spurs and promote distinctive breakdown features. N 30-45 E set joints locally guide early flow paths, mostly in the lower Greenbrier Group, but the N 60-75 E set joints are much more important where both are present. N 30-45 E set joints guide abundant blind pockets and joint spurs or bells in the lower Greenbrier Group, especially in base level conduits near the Greenbrier River. Thrust-fault ramps from argillaceous units or prominent bed partings, are common, and guide many early flow paths. Thrust faults provide fracture zones subject to extensive collapse, resulting in large collapse chambers and lengthy breakdown-filled canyons that readily transmit water but thwart cave exploration and mapping. Distinctive patterns of conduit growth associated with thrust faults include inclined anastomotic mazes, isolated vadose trenches, and isolated retreating canyons. Wedge-shaped breakdown is associated with many thrust fault zones. Crests of anticlines have normal faults and high angle extension joints that locally guide early flow paths, have associated solution pockets, and produce slab-shaped breakdown.

Surface and Subsurface Drainage Basin Asymmetry: Ramifications for Karst Development in the Appalachian Plateaus

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ABSTRACT

Basin asymmetry is a geomorphic parameter that quantifies disparate size and growth of a drainage basin. For surficial drainage, the Asymmetry Factor (AF) is determined by delineating the master stream, measuring the area to the right of it (AR) and the total basin area (AT). AF is then given by: AF = 100 (AR/AT). An AF value > 50 indicates excessive area in the sub-basin on the right.

This study introduces two related geomorphic parameters useful in fluviokarst settings, Subsurface Drainage Density (D_{sd}) and Subsurface Asymmetry Factor (SAF). D_{sd} = Cave length / Basin area, for either the whole or a part of the basin. SAF = 100(CLR/CLT) where CLR and CLT are the total length of known caves for the right side of the basin and the total basin, respectively.

A strongly karsted sub-basin of the East Fork Obey River, Fentress County, Tennessee was chosen to explore the relationship between these parameters and karst development. This basin is ideally suited for this analysis, having a strong asymmetry and a large number of known caves (99). The basin is developed on the Cumberland Plateau, and is underlain by undeformed sedimentary rocks of mixed lithologies. Local relief is 340 meters, with sharply incised valleys dissecting the flat plateau surface. Regional dip is 0.5 degrees to the ESE. The sub-basin has a total surface area of 184 km^2 , and an AF of 29; strongly biased to the left. The SAF is opposite of this, 87, indicating that most cave passage occurs on the right.

In this basin, AF is structurally determined. Dip slope (consequent?) streams on the left side initiated within larger sub-basins, and were able to maintain/expand them, while right-sided sub-basins descended the free face, and were limited in their catchment. Subsurface karstification, as indicated by the SAF, has been promoted on the right side by subsurface capture of the master drainage, probably under structural control. AF and SAF may serve as time-integrated proxies for surface and subsurface erosion rates.

When adjusted to include only those caves related to tributary drainages, the SAF is 22, close to the AF (29) This suggests a possible relationship between the length of caves developed and catchment area. Such a relationship would be expected from mass balance considerations.

Introduction

The Appalachian Plateaus are home to many well known fluviokarsts such as the Elk River and Greenbrier County karst terranes of West Virginia, the Carter Caves and Cave Creek areas of Kentucky, and Rye Cove, Virginia. In these areas, streams have incised through a mixture of lithologies exposing speleoliferous carbonates. Abandoned caves are left in the valley walls, and the valley bottom frequently has an active karst. There is usually some integration between the valley-wall and the valley-bottom karst systems.

Lithology and structure of the Appalachian Plateaus are similar along the length of the orogen. Cap rocks are generally Pennsylvanian conglomerates and sandstones. Beneath these are Lower Pennsylvanian and Upper Mississippian clastics, underlain by Mississippian mixed lith-

ologies, including thick limestones. Although appearing flat and featureless (Figure 1), broad folds occur with regional dips of up to several degrees. Faulting is present in some areas, in association with the folds and the eastern boun-dary with the Valley and Ridge.

Basin asymmetry is a geomorphic parameter that quantifies disparate size (and possibly growth) of contributing areas in a drainage basin. It is simply the percentage of the total basin area which is present on the right-hand side of the basin (Figure 2). For surficial drainage, the asymmetry factor (AF) is determined by delineating the master stream in the basin, measuring the area to the right of this (as viewed downstream), and the total basin area. AF is then given by: AF = 100 (AR/AT), where AR is the area on the right, and AT is the total area. AF < 50 indicates excessive area on the left side of the basin. An alternative method is to measure total stream lengths on either side, but this is more difficult and unnecessary, as area generally serves well as a proxy for stream length.

AF quantifies disparate growth, and can be used to pinpoint the reasons for this disparity. Hare and Gardner (1985), for example, used AF to discriminate actively uplifting fault blocks in Costa Rica (Figure 3). Because it is dimensionless, it can be used to compare basins of different sizes.

Two related geomorphic parameters that are applicable in fluviokarst settings are introduced in this paper: Subsurface Drainage Density (D_{sd}), and Subsurface Asymmetry Factor (SAF; *see* Figure 4). D_{sd} is the known length of caves in a basin or sub-basin, divided by the area of the basin. It is similar to D_d (surficial drainage density), which is given by the total stream length divided by the surface area. D_{sd} is designed to quantify the "karstifica-tion" of the basin in the same way that D_d quantifies surfi-cial drainage development.

SAF is essentially a comparison of D_{sd} 's, similar to AF in that we normalize it to the total length (total area in the case of AF) in the basin. SAF = 100 (CLR/CLT), where CLR (cave length right) is the total length of known cave passage to the right of the master stream, and CLT (cave length total) is the total length of caves known in the basin. The formula used to calculate SAF can be modified for special purposes by normalizing for surface area, rock type, etc., as is demonstrated later in this paper.

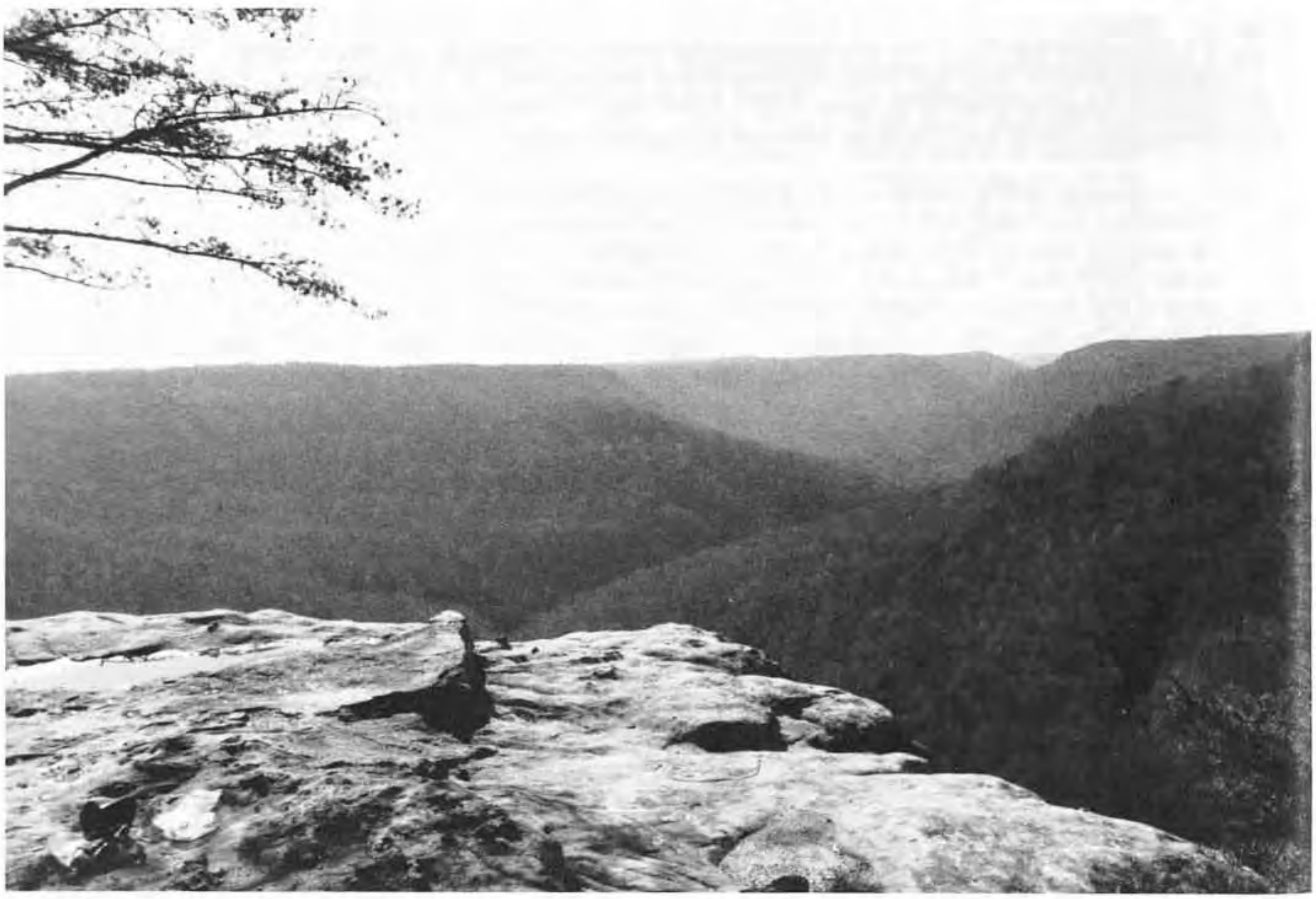

Figure 1: Level upland surface of the Cumberland Plateau, Tennessee. Horizon and outcrop in foreground are the Pennsylvanian Rockcastle Conglomerate.

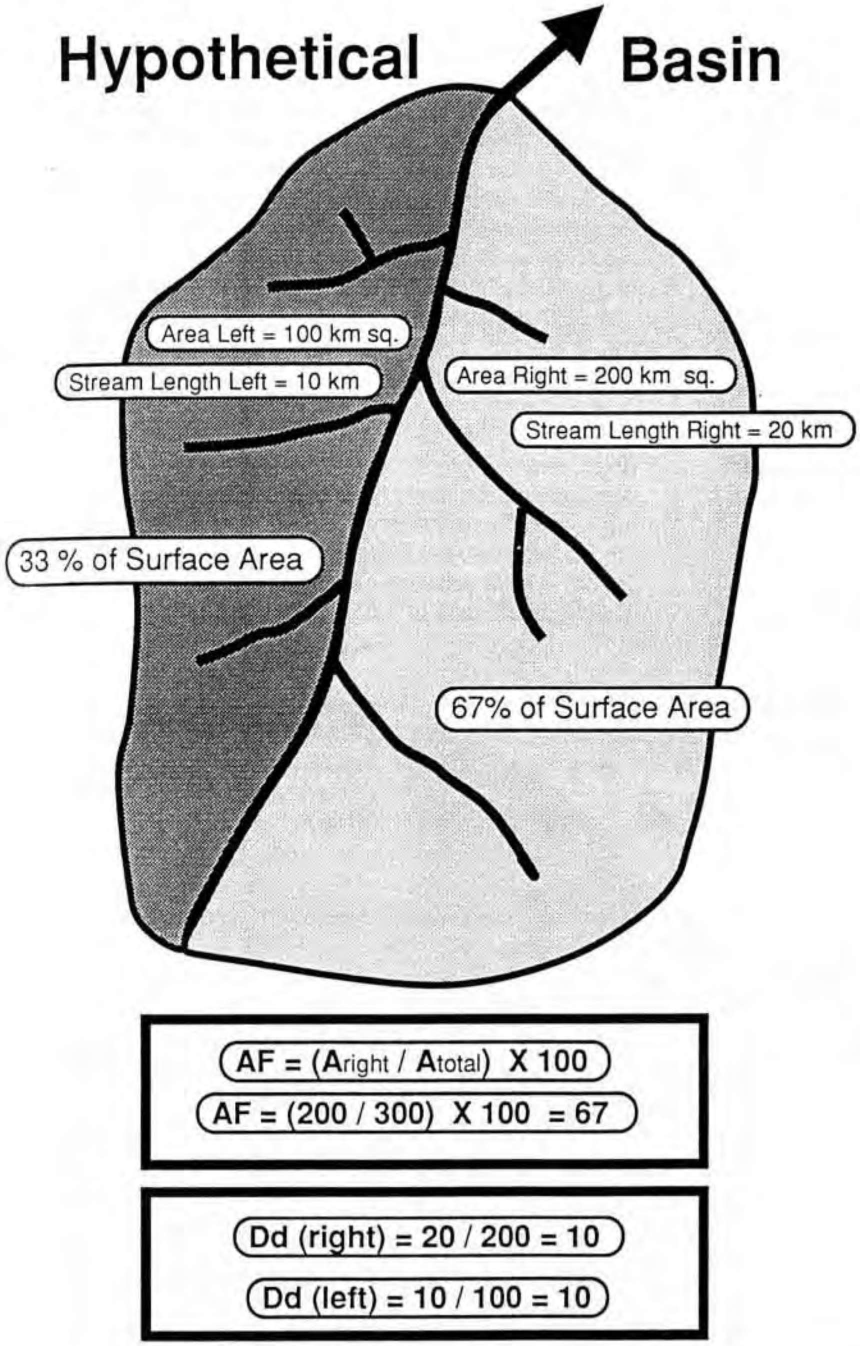

Figure 2: Determination of Asymmetry Factor (AF) and D_d for a hypothetical basin.

Factors Controlling AF

At an instant in time, AF expresses as one number a ratio of basin areas; what does this mean in terms of processes? It can indicate many things, and any further interpretation of the phenomena must be made with additional geologic and hydrologic data.

In all but the largest of basins, it is reasonable to assume that erosion of the "initial" surface began at the same time throughout the basin. In the case of the Appalachian Plateaus, this would be sometime in the Late Permian or Triassic. In the present, if we find that one side of the basin is larger than the other, it implies one of two situations: 1) the larger side began with a larger catchment through fortuitous location of the master stream and has maintained this dominance; or 2) the larger side has been more successful in expanding its catchment for some reason. At the basin scale, reason 2 is attributed to headward growth of tributaries or piracy, and the master stream is considered to be fixed in position.

Primary controlling variables of AF are:

1) Structure - Attitude of the rocks affects initial and subsequent drainage basin enlargement. Streams flowing on dip slopes are given more basin area at the outset.

2) Stratigraphy - In an area of mixed lithology, sub-basins with more erodable substrate may grow more rapidly than others.

3) Climate / Vegetation - Variable rainfall and types of vegetation can stabilize parts of the basin, protecting them from erosion.

4) Tectonism - Currently active uplift can serve as a differential driving force for erosion and migration of divides.

Variables 3 and 4 are not dominant factors in the Appalachian Plateaus.

Factors Controlling D_{sd} and SAF

To be at all meaningful, the area of interest must have been thoroughly searched for caves, and the caves must have been surveyed. Given this condition, for both D_{sd} and SAF the assumption is that both the number and length of undiscovered caves on both sides of the basin are proportionately equal and evenly distributed (*i.e.*, if we do not have a complete sample, it is at least a representative sample).

Primary controlling variables of D_{sd} and SAF are:

1) Structure - Controlling movement of groundwater. Areas with more flow will develop more caves.

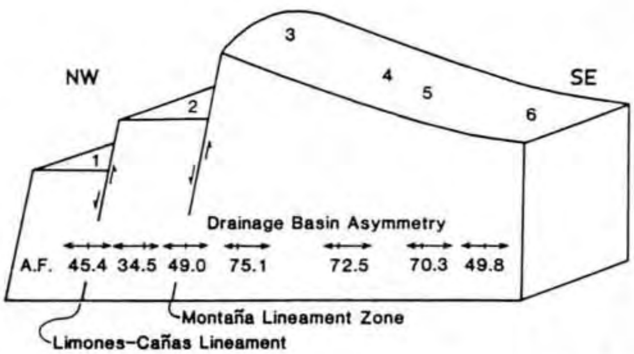

Figure 3: Asymmetry of several adjacent basins on the Nicoya Peninsula, Costa Rica. The researchers were able to delineate between uplifted fault blocks on this basis (from Hare and Gardner, 1985).

2) Stratigraphy - Location of soluble beds and their relation to surface- and ground-water routes of flow.

3) Surficial erosion rates - Caves may form but subsequently be destroyed or disjointed by surficial erosion (unroofing), thereby biasing the calculated values.

Utility of AF and SAF

The utility of AF's and SAF's is two-fold. First, they provide a normalized quantitative measure allowing comparison of two or more basins. Second, in quantifying these discrepancies, we are forced to explain them, and so think about speleogenesis and basin growth in fluviokarsts. If SAF exists, why does it? Destruction of caves or lack of development? If lack of development, why do caves form preferentially on the one side? What does all this indicate about basin growth as a whole?

An Example - East Fork Obey River

A strongly karsted sub-basin of the East Fork of the Obey River (EFO), Fentress County, Tennessee was chosen to explore the relationship between asymmetry factors and karst development. This basin is well suited for such analysis; its surface has been extensively walked by cavers, it possesses a strong asymmetry, and it has 99 known caves with a total length of 296,021 feet.

The drainage is incised into the Cumberland Plateau, in undeformed rocks with a regional dip of 0.5 degrees ESE. The present tilt is associated with the uplift of the Nashville Dome, which occurred throughout the Paleozoic Era. The Plateau surface and upper reaches of the streams are underlain by Pennsylvanian and Upper Mississippian clastics. The lower reaches and valley bottoms are floored by Mississippian carbonates. Local relief is 340 meters, with sharply incised valleys dissecting the Plateau surface. The basin has a total area of 184 square kilometers, and an AF of 29, strongly biased to the left side. The gross SAF is 87, meaning that most cave-passage length is on the right side.

Within this basin, AF is structurally determined (Figures 5 and 6). The sub-basin on the left (West) side is composed of dip slope (consequent?) streams. Due to their initial large catchment, these streams have successfully eroded back into the Plateau. Right-side streams, descending the structural free-face, have been limited in their growth due to lack of initial catchment.

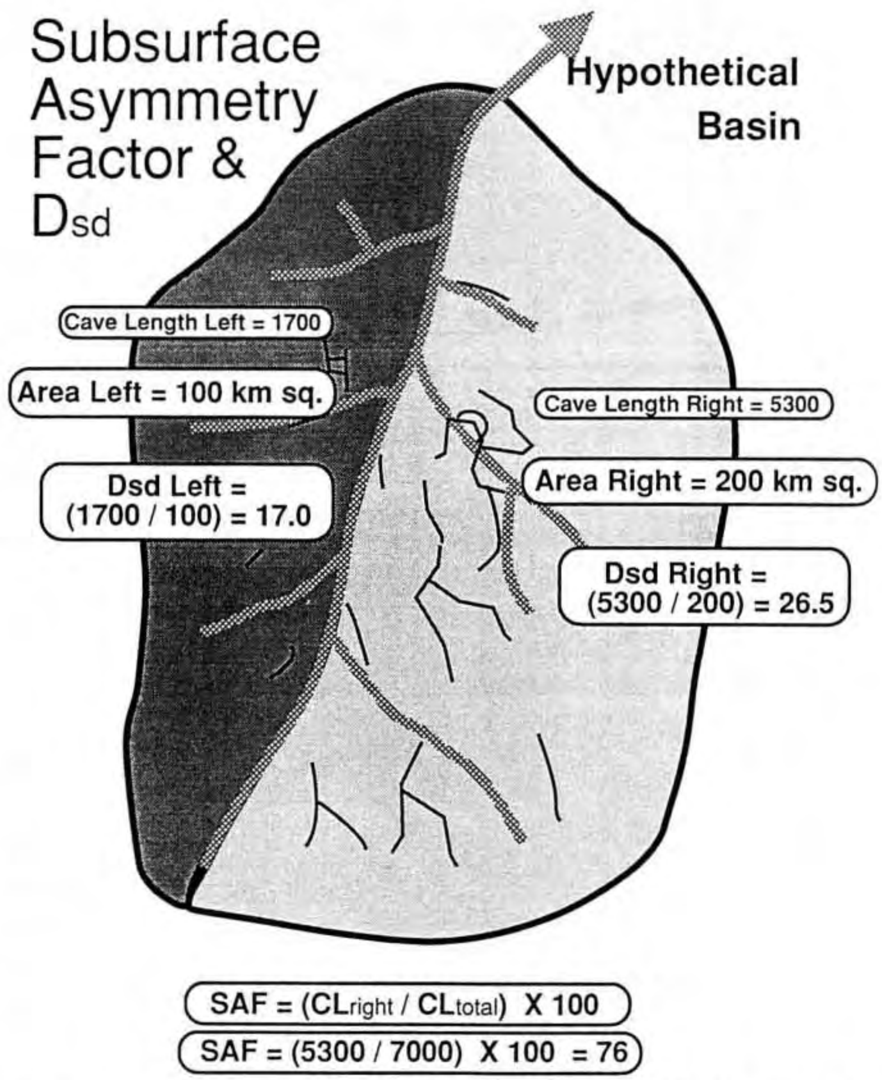

Figure 4: Determination of SAF and D_{sd} for a hypothetical basin, *see* text for discussion.

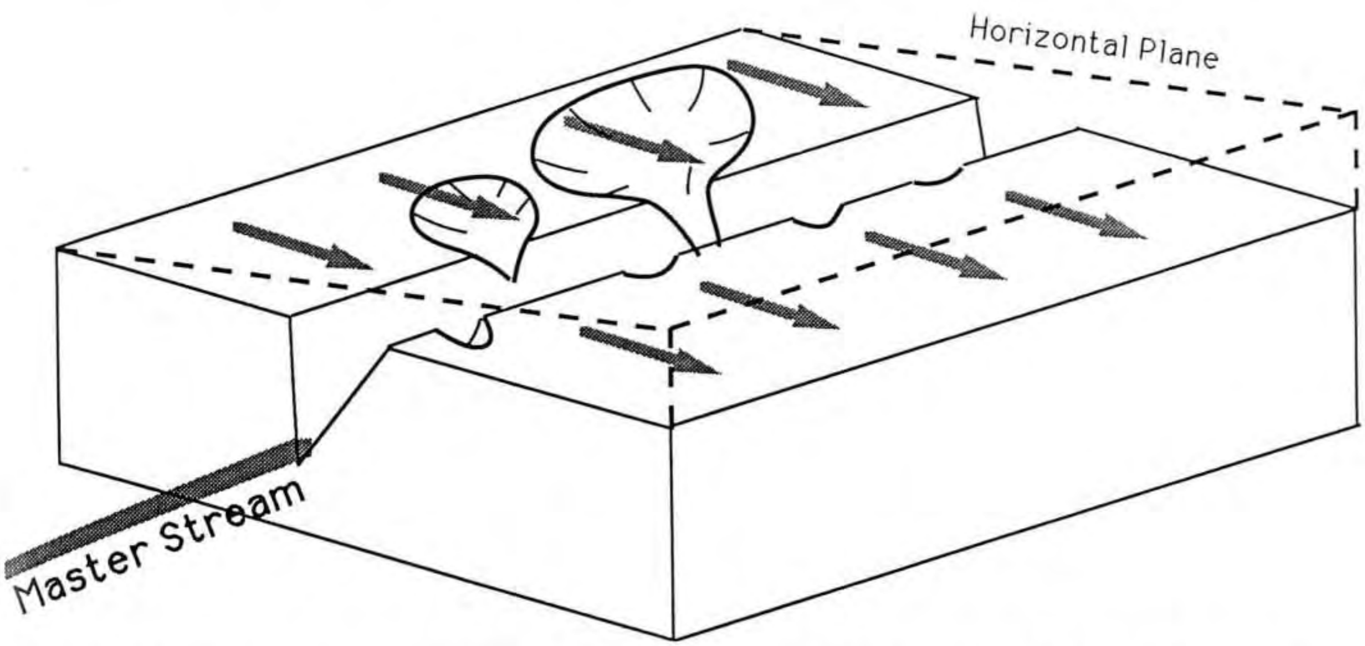

Figure 5: Scheme for development of surficial asymmetry in the sub-basin of the East Fork Obey River at Fentress County, Tennessee. Broad arrows indicate consequent streams on plateau surface. Master stream is incised transverse to this. Sub-basins on the left side have larger initial catchment, and rapidly incise. Sub-basins on the right have small initial catchment, and form only small, steep drainages.

SAF, too, appears structurally determined, but closer analysis of the asymmetry leads to some interesting problems (Figure 6). Of the 99 caves in the basin, 58 are on the left side (total length = 39,624 feet), and 41 are on the right (total length = 256,397 feet). On the right side, however, three caves of great length (Xanadu, Zarathustra's, and Mountain Eye caves) are genetically different from those in the rest in the basin, and account for 96 % (245,067 feet) of the length on the right side. These 3 caves are all found on the East side of the main stream, and appear to have been formed through capture of flow of the EFO, whereas the others in the basin are related to tributary drainages. These caves skew the whole basin value. If they are excluded from the sample, leaving only 11,321 ft of cave on the right, the adjusted SAF is 22. The effect of the shallow (0.5 degree) regional dip has had importance in the location of the three largest caves.

Where does the adjusted SAF point? It brings theAF (29) and SAF (22) values quite close to one another, and points to a possible (and expected from mass-balance considerations) relation between surface drainage and karst development - that the length of caves formed is proportional to the drainage area.

Conclusions

SAF and D_{sd} are parameters that are useful in conjunc-tion with other morphometric parameters in understanding a basin. At the EFO site, there is an apparent relation be-tween surface catchment size and length of caves. It would be useful to see if the relation holds true in other areas in the Appalachian Plateaus. There is no lack of potential test sites, but finding areas that meet the criteria of tho-rough exploration and large size may be difficult.

Acknowledgments

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References

Hare, P.W. and Gardner, T.W., 1985, Geomorphic indicators of vertical tectonism along converging plate margins, Nicoya Peninsula, Costa Rica, *in* Hack, J.T. and Morisawa, M.E. (editors), *Tectonic Geomorphology: Proceedings of the 15th Geomorphology Symposia Series, Binghamton*, p. 76-104.

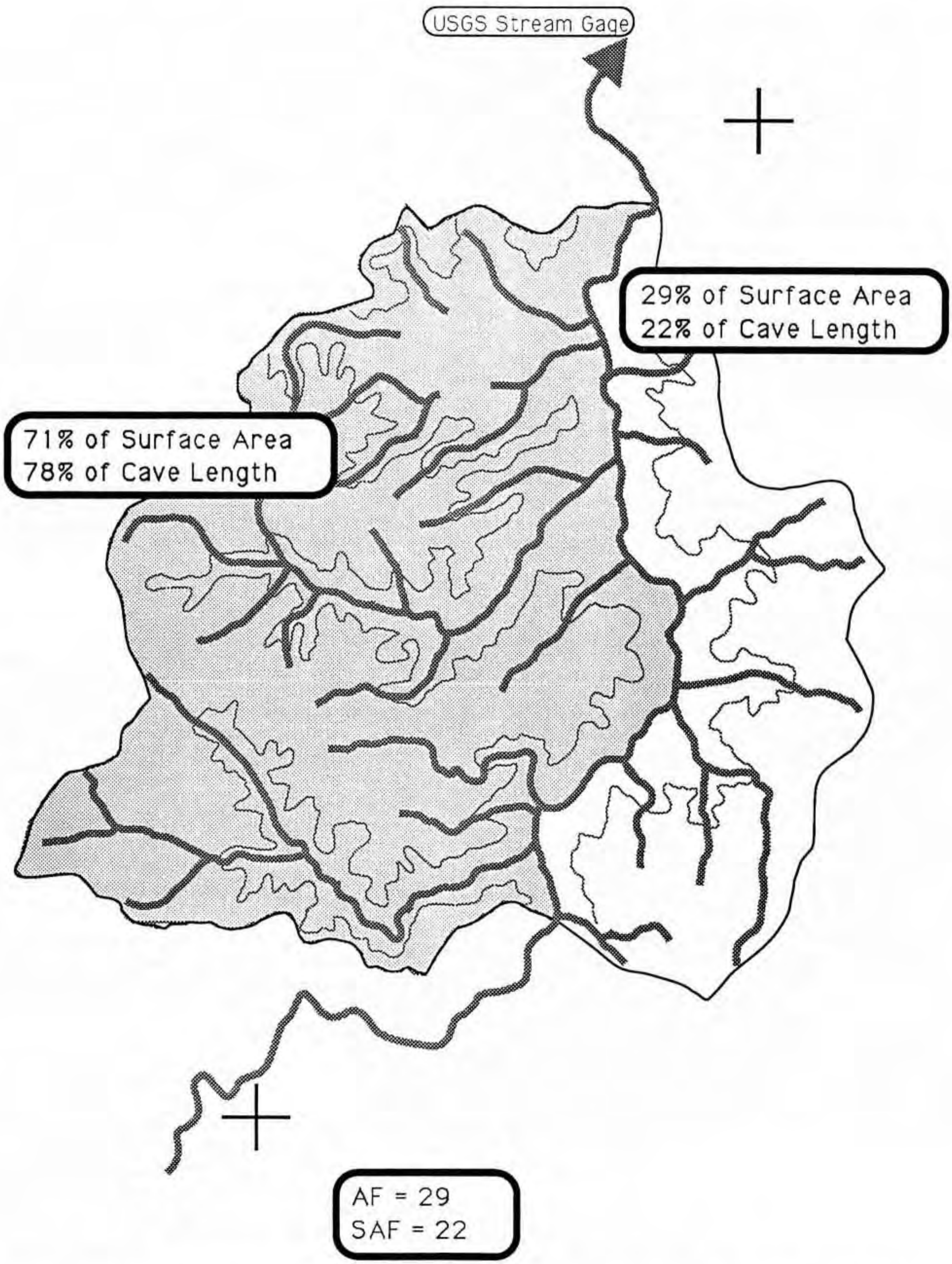

Figure 6: Morphometric relations for the karsted sub-basin of the East Fork of the Obey River. When the 3 longest caves are removed from the sample, surface area correlates well with known cave length. Values shown have been adjusted in this manner.

Environmental Education Regarding Karst Processes in the Appalachian Region

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ABSTRACT

The intracacies of karstic systems make it difficult for a non-specialist to adequately understand what karst is or how it works. Yet, increasingly, it is important that individuals in industry and commerce, in local and regional government, and in the private sector become aware of unique problems associated with karstic terranes. The Appalachian region contains some of the finest karstlands in the United States. However, this region is experiencing one of the highest rates of economic growth in the country. Sections of Appalachian karst are being developed for housing, agriculture, industry, commerce, and transportation.

Karst results from interactive geologic and hydrologic processes that operate both on the surface and in the subsurface. A basic working knowledge of these processes needs to be provided to those who must address environmental problems in the Appalachian region. To do so, the complex character of karst must be reduced to a few fundamental geologic and hydrologic concepts that can be easily visualized and understood. There are five *fundamental requisites for dissolution* to produce karst in any locale: (1) The bedrock must be relatively soluble in natural water. (2) Karstic surface- or groundwater must be chemically aggressive. (3) The bedrock must be both porous and permeable in order for groundwater to move and transmit the solution and solute. (4) The region must possess topographic relief to induce and sustain water flow. (5) Sufficient time must elapse to produce karstic landforms. On a site-specific basis, five significant *modifying factors* govern the local character of karst: (1) Variations in lithostratigraphy determine which rock units are most susceptible to karstification. (2) Geologic structure governs the degree to which rocks are fractured and along what orientations flow systems will evolve. (3) Evolution of the regional topography ultimately establishes inputs, outputs, and paths of flow through the bedrock sequence. (4) The hydrodynamic character of the flow system affects the rapidity of flow and rate of dissolution. (5) Local climate adjusts the rates and intensity of physical and chemical processes that excavate the rock. Knowledge of these attributes leads to a better understanding of local karst settings and potential environmental problems. Specialists in karst must convey these concepts to the non-specialist in a clear manner using easily understood dialog and visually effective graphics. Education in karst processes could and should occur at all levels, including public schools, youth groups, institutions of higher education, civic groups, and governmental agencies.

In recent years, various methods have been employed toward the goal of educating the public about the sensitive nature of karstic environments. These efforts have ranged from interaction between local cavers and the general public to cooperation between speleologists and governmental agencies. The cumulative effect of these endeavors has been encouraging as some local communities strive to 1) *clean up* from past abuses, 2) *monitor* their present use, and 3) *protect the future resources* of their karstlands. Caving groups or other environmental organizations may be able to incorporate into their educational outreach activities some of the materials that were previously generated and are readily available as aids.

Introduction

Landforms developed on or within carbonate rock (limestone and dolostone) through dissolving are collectively known as *karst* (*see* Monroe, 1970, for definitions of karst terms). Much of the karstic landscape of the Appalachian region lies within the Valley and Ridge and Appalachian Plateau physiographic provinces and is characterized by sinkholes, caves, sinking streams, springs, and solution valleys. This karst region is among the largest and

finest in North America (Herak and Stringfield, 1972; Kastning, 1986). Carbonate-rock terranes pose environmental problems that are unique with respect to the wide spectrum of bedrock types, and karstic landscapes are particularly sensitive to environmental degradation (LeGrand, 1973; White, 1988, p. 355-405). Stresses induced by mankind in karstic terrane result in environmental problems that are much more acute than those that would occur in terranes underlain by either crystalline (metamorphic or igneous) or clastic (other sedimentary) rock. Problems such as supply and contamination of groundwater and land instability abound in the Appalachian region, as they do in most populated karst regions, worldwide.

The study of karst is a relatively new science that draws largely on the principles of geology and geography. A thorough understanding of the processes that occur both at the surface and underground and an appreciation for the total hydrologic system necessitates a global familiarity with scientific karst studies. The level and scope of modern karst studies is demonstrated by the recent proliferation of textbooks on the subject (Kastning, 1989a). Recent texts in English include those of Bögli (1978), Dreybrodt (1988), Ford and Cullingford (1976), Ford and Williams (1989), Jennings (1985), Sweeting (1973), Trudgill (1985), and White (1988). Moreover, the number of scientific journal articles and graduate theses on karst is expanding at a phenomenal rate (*see for example* the bibliographies of Lamoreaux and others (1970, 1975, 1986), White and White (1984), and Huppert (1988)).

Despite accelerating research on the subject, karst is not an easy concept for the lay person to understand at the outset. Generally speaking, most people know very little of what exists and what happens below the surface of the Earth. Yet, these same people may have a much greater perception of surficial processes. It is easy to understand why. Surficial phenomena are easily observed and therefore understood; phenomena in the subsurface are not typically seen, much less comprehended ("out of sight, out of mind"). Perhaps most importantly, the average person views the aboveground and the underground as *two distinct zones*, separated by a boundary, namely the surface of the Earth. The most formidable challenge in successful karstland management is breaking down this dichotomy. Accordingly, there are three fundamental concepts that must be stressed: (1) karst is a *single unified system* that integrates the surface and subsurface, (2) karstlands are environmentally very sensitive terranes, and (3) the physical and chemical processes that create karst are the very processes that easily lead to environmental problems.

Karst as an Environmental Liability

Much of the karstic terrane of the United States lies in rural regions where environmental impacts are generally limited to those imposed by agricultural practices and highways (Davies, 1970). In some cases, karst lies within the confines of public land (parks, forests, and the like). However, urbanization is rapidly encroaching in many karst areas and economic development is resulting in severe karst-related environmental problems.

An appreciation for the extent and complexity of environmental problems in karst is best gained by consulting proceedings volumes of recent conferences that have specifically addressed these issues (*e.g.* Beck, 1984, 1989; Beck and Wilson, 1987; Dougherty, 1983; and this volume). The interaction among various natural elements of the karst setting and man's role in the system is illustrated conceptually in Figure 1. The system is far from simplistic and consists of a series of nested loops with feedback that represent direct and indirect causes and effects. Making changes in any one element of the system will have consequential impact on other elements. It is not the intent of this paper to address all of the possible impacts that man and karst have on each other. Rather, it is instructive to select a few situations where karst may be important as an environmental concern in the Appalachian region and to consider methods to interpret pertinent geologic and environmental principles to the public at large.

Mismanagement of karstlands, whether through unsupervised economic development, poor farming practices, improper waste disposal, or other means will often damage groundwater supplies, cave ecosystems or man-made structures built on karst. Among the most severe and immediate environmental problems associated with karst include

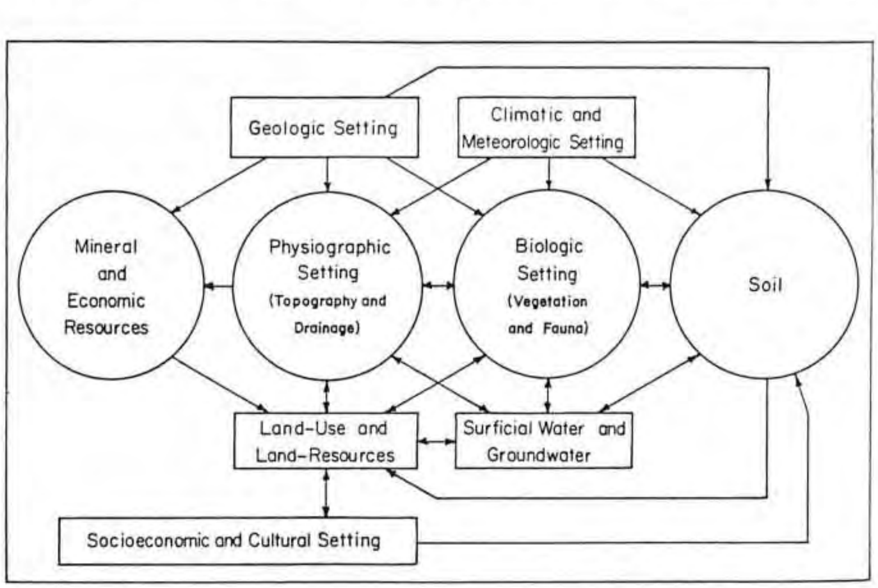


Figure 1: Cause-and-effect relationships among geologic, biologic, geographic, and human factors in karst terranes. Arrows portray directions of effects. From Kastning (1989c).

groundwater supply, groundwater quality, and land instability. Each of these problems is briefly discussed below (*see also* succeeding papers in this volume).

Groundwater Supply

In karst regions, unlike other types of terrane, groundwater is channelized within a natural underground system of interconnected 'pipes' (*see* Plate B, page 84). These natural pipelines collectively transmit water from input points (*recharge zones*) to output points (*discharge points*).

Recharge in karst terrane occurs as water percolates through the soil and into fractures in the carbonate rock over large areas of the countryside (*diffuse recharge*) or as surface streams sink in their entirety where they flow into caves or sinkholes (*discrete recharge*). In most karst regions, both mechanisms operate simultaneously.

Discharge from karstic aquifers occurs where water seeps from the ground over a wide area (*diffuse springs*) or where underground rivers or cave streams exit from large openings (*discrete springs*). Springs may issue anywhere from a few to thousands of gallons per minute. A significant quantity of water is also discharged through man-made wells drilled to obtain water for domestic, commercial, agricultural, or industrial use.

Obtaining usable amounts of water from karstic aquifers may be a 'hit-or-miss" operation. Water is highly localized because it is flowing through solutionally enlarged fractures, within partings between beds of rock (*see for example*, Zewe and Rauch, this volume), or along locally highly permeable beds (Kastning, this volume). In sandstone and other porous-media aquifers, flow is diffuse throughout. In karstlands, however, wells may not yield sufficient water unless a solutional conduit is intersected.

Springs and wells in karst are also highly sensitive to variable weather patterns such as a draught or wet periods, and respond rapidly to weather changes as water is quickly conveyed along solutional conduits. Karstic groundwater supplies, therefore, are flashy and allowances must be made for this erratic behavior in the allocation of water derived from springs or wells.

Another common misconception by the general public relates to the relative roles of sinkholes and groundwater flow. Water does not enter the subsurface at sinkholes just because they are there - but, rather, *sinkholes exist because water enters the subsurface at that locality* and has done so for sufficient time to dissolve and remove the overlying materials. Infilling of a sinkhole will not alter that inherent relationship to the subsurface. Man-made changes to the surficial drainage and to sinkholes may, however, easily alter the *rate* at which the underlying aquifer receives its normal recharge. Vegetation and soil cover slow runoff and absorb some moisture, thereby providing less "flashy" recharge than would impermeable materials (*e.g.* cement drains, asphalt roads or parking lots, and roofs of structures). Sinkholes that have been infilled are less efficient inputs and may cause surficial water to pond or backflood, unless it is diverted away from its natural sinkpoint (thereby altering the recharge at yet another sinkpoint) (*see* Mills and others, this volume). These activities, increasing the rate of runoff and/or blocking the input points, may cause castotrophic flooding or collapse and drastically alter the quantity of groundwater available for use in the immediate vicinity.

Groundwater Quality

If there is one single environmental issue that stands out in the karst of the Appalachians, it would have to be the sensitivity of the karstic aquifers to groundwater contamination. Man's impact is most severe in cases where polluted surface waters enter karstic aquifers. This problem is universal among all karst regions in the United States that underlie areas of economic growth. On the positive side, most karst in the Appalachians lies in rural areas. On the negative side, the region's karstic groundwater problems are increasing with the advent of (1) expanding urbanization, (2) increased production of environmentally unacceptable artificial chemicals, (3) shortage of repositories for hazardous wastes (both household and industrial), and (4) ineffective public education concerning waste disposal and the sensitivity of the karstic groundwater system.

There is a general lack of public understanding of groundwater behavior, particularly in karst. A common perception is that underground waters, such as those flowing from springs, are filtered and nearly pure. On the contrary, karst aquifers *can not filter* contaminated groundwater sufficiently to render it potable at the discharge sites; and because recharge points are directly connected to discharge points, conveyance of contamination is highly efficient.

Sinkholes are natural holes in the ground surface and to the general public appear to be natural sites for dumping of trash (*see* Plate K, page 204). The presence of a sinkhole "eliminates the need" to dig a pit into which refuse can be dumped. The number of active and inactive sinkhole dumps in karst regions is staggering. For example, over 260 illegal dumps have been inventoried for Rockbridge and Botetourt counties, Virginia alone (Slifer, 1987; Slifer and Erchul, 1989; Erchul, this volume). It is estimated that each county with karst in the Valley and Ridge Province has hundreds of sinkhole dumps. The profusion of these dumps is the result of (1) a lack of a refuse-removal service in rural areas and the expense and inconvenience of trash haulage on the part of the landowner, (2) the convenient proximity of sinkholes, and (3) ignorance of the karstic groundwater system on the part of the landowner.

Sinkholes as natural funnels convey toxic substances directly into the karstic plumbing system (Kastning and Kastning, 1990). In many cases, chemicals may be trans-

mitted directly to domestic wells in a matter of a few hours and without filtration. A farmer who places a carcass of a deceased farm animal into a sinkhole (an all too common procedure) may very well be drinking water that passed through that sinkhole! Or, his neighbors may be.

Unfortunately, sinkhole dumping is only one way of contaminating a karstic groundwater supply (Aley, 1972; Aley and others, 1972). Chemical fertilizers and pesticides applied to fields overlying carbonate rock will enter the aquifer through diffuse infiltration and contaminate springs and wells. Runoff from feed lots may also (*see* Brown and Ewers; Ogden, Hamilton, and others, both in this volume). Improper siting of municipal landfills on or near karst causes leakage or runoff from these landfills to easily contaminate karst waters. Corroded underground storage tanks, such as at active or abandoned service stations, may release hydrocarbons directly into karstic aquifers. Chemicals introduced in these ways may include many of the most hazardous, including hydrocarbons, heavy metals, and others. Additionally, leaky septic systems, sewage lines, or effluent from faulty sewage-treatment facilities introduce coliforms and other disease-bearing organisms into the karst system.

Many streams, including those in rural areas, are polluted and most surficial streams in karst terranes readily lose water into the ground through their stream beds. Contaminated waters from the surface that enter carbonate rocks introduce their toxic substances into subsurficial streams. The only difference between surface and underground waters in karst is that the latter is out of sight (and out of mind)! Chemically they may be identical. Accidental spills from overturned tanker trucks and runoff from highways salted in winter to prevent freezing are just two examples of possible contamination along transportation corridors (Werner, 1983; Lovegrove, 1988). Effluent from commercial and industrial operations along such corridors may also be a problem. However, industrial and governmental leaders are often unaware of the sensitivy and inherent complexity of groundwater flow in the karstic subsurface.

Caves contain fragile organisms that have evolved in the natural cave environment. Bats are the most commonly recognized creature of caves, but there are actually an amazing variety of cave-dwelling organisms. Because these animals are highly adapted to their constant ecological surroundings, they are particularly sensitive to disturbances. Foremost of these is the introduction of foreign substances into the groundwater that flows through the caves. Even "clean" fill, such as brush, hay, sawdust, or dirt, may lead to chemical imbalances in karstic groundwater as the vegetative matter decays rapidly and consumes oxygen. In order to conserve species endemic to caves, rare or otherwise, man must maintain clean groundwater in karst regions. Indeed, these vulnerable ecosystems and the organisms that reside in them can serve as indicators of the quality of the groundwater. The only effective means to monitor and protect these organisms is through detailed biogeographic inventories and educating the public.

Many who visit caves on commercial tours are intrigued by the myriad of cave formations. There are, in fact, many unusual, fragile, and often rare formations in caves (Hill and Forti, 1986). These formations, that take centuries or millenia to form, are highly susceptible to contamination and derangement of groundwater flow. The fragility and rarity of these deposits can be easily conveyed to visitors of commercial caves. Important environmental messages can be provided in show caves by properly trained tour guides through entertaining, but educational, interpretation.

Subsidence and Ground Instability

The potential for the surface in karst regions to give way in collapse is brought home from time to time in the media. Massive collapses in which homes or businesses are swallowed by newly formed sinkholes make exciting news. In some states, such as Florida, Alabama, Texas, and Pennsylvania, such occurrences are somewhat frequent (*see* papers by Dougherty and Beck, this volume). Most of these events are triggered by man's intervention with the karstic environment (Waltham, 1989). The most common cause for catastrophic sinkhole collapse is an overpumping of groundwater from karstic aquifers, resulting in a relatively sudden loss of bouyant forces that uphold roofs of cavernous openings. A second cause of collapse occurs in response to changes in the position of the water table due to modifications to surficial runoff and infiltration to the karstic groundwater system.

In areas undergoing development, sinkholes are often viewed as unwanted holes in the ground. If they are filled in to produce level land, the potential for ensuing environmental problems is twofold: First, naturally developed paths of infiltration are often blocked, leading to ponding or flooding on the fill (*see* Plate G, page 158). Secondly, over the long run, fill materials will be sapped into the subsurface and settling may occur. These disturbances easily impact any structures built on the fill. Additionally, the increased weight of water, fill, and structures upon the cavernous bedrock could cause catastrophic collapse in the future (*see* Plate H, page 158).

Fundamentals of Karst Processes

The origin of karstic landforms including caves and other avenues of groundwater flow in soluble rock is ultimately tied to several factors. To have a basic understanding of what karst is and how it operates as a system requires that the fundamental mechanisms responsible for the origin of karst be clearly recognized. Basic education in karst processes should include knowledge of the requisites for karstification and local factors that control and modify these processes at specific sites.

The following discussion comprises an overview of the nature of karst in the Appalachians. The humid-temperate karst in the Appalachian foldbelt of the eastern United States is characteristically different than karst regions elsewhere, for example in arid or tropical regions or in areas underlain by relatively horizontal strata. Where appropriate, karst in Virginia is used as an example because, in many ways, the karst of Virginia is typical of karst in the Appalachian region. Also, the authors have had recent experience with conveying the concept of karst to the lay public in Virginia.

Requisites for Karstification

Soluble Bedrock: The principal method by which caves and similar avenues of water flow are excavated from rock is by *dissolution*, the chemical corrosion of the bedrock by water flowing through it. Many rocks, such as sandstone, granite, shale, and gneiss, are relatively insoluble and very little material is removed by the dissolving action of groundwater. In contrast, calcite, the main ingredient of limestone and dolostone, is highly soluble. It follows that development of dissolutional openings will occur in sedimentary rocks that have a high content of calcite. Limestones of the Valley and Ridge and Appalachian Plateau provinces, for example, are relatively high in calcite content. In places, beds of high solubility are interbedded with those less soluble. Where this occurs, but where other factors remain relatively similar, the greatest development of solutional voids (cave passages and the like) conforms to limestone beds of high purity (*see* Kastning, this volume). Comparisons of maps showing the distribution of caves and other karst features in Virginia (*e.g.* Hubbard, 1983, 1984, 1988; Miller and Hubbard, 1986) and inventories of caves (*e.g.* Douglas, 1964; Holsinger, 1975, 1985; Virginia Speleological Survey, 1987-present) with published geologic maps clearly show that the areal distribution of caves and sinkholes conforms to outcrops of carbonate rock.

Porosity and Permeability: Water is unable to flow through bedrock unless there are openings capable of holding water (*porosity*) and unless those pore spaces are interconnected. The ability of a rock to transmit water along avenues of porosity is referred to as *permeability*. Well integrated and mature karstic drainage networks develop where porosity and permeability are high. Such conditions often include fractures such as joint sets and faults (*see* Plate C, page 100). Fractures form as a result of deformation of bedrock in response to crustal stresses (such as those that formed the Appalachian Mountain range). Intense folding, uplift, and other events result in a high density of joints and faults as brittle rocks break (Kastning, 1977, 1984). Careful perusal of maps of caves shows that many linear passages and chambers closely follow the orientation of principal joint sets. Some significant caves in Virginia are closely associated with faults (Krinitzsky, 1947) and sinkholes are commonly aligned along fracture traces, indicating that infiltration of surficial waters into karstic aquifers occurs along joints and faults (Kastning, 1989b). In the Appalachians, this is best illustrated where broad, low lying valleys are floored with nearly horizontal strata (Kastning, 1988, 1989b).

Chemically Aggressive Water: In order for groundwater to remove limestone by dissolution it must be acidic and not yet saturated with dissolved calcite. In limestone, freely circulating water is typically charged with dissolved carbon dioxide (much like a carbonated beverage) that was picked up as the water percolated through soil. This weak solution of *carbonic acid* seeps into fractures in the rock. Over time slow, but persistent, dissolving of limestone by this solution can create voids in the bedrock ranging from inches in diameter to hundreds of feet in width and height. Flow paths are often long, in many cases producing caves with tens of miles of passages. The extent of soil cover over most areas of limestone in the Appalachians ensures continual enlargement of flowpaths by chemically aggressive karst waters.

Topographic Relief: Groundwater will not flow unless water entering the ground at one elevation can exit at a lower elevation. Differences in elevation, known as *relief*, are necessary to maintain an exchange of water along the flowpath. Without the removal of water saturated with respect to calcite (from the dissolving of carbonate rock) and the input of fresh, chemically aggressive water continually coming into contact with the walls of fractures and incipient cave passages, no further dissolution will occur. The limestone areas of the Valley and Ridge Province have considerable relief and the maintenance of groundwater flow is ensured. Moreover, flow rates are kept reasonably high by steep hydraulic gradients in parts of many drainage networks. Runoff from intense rainstorms and snowmelt in the Appalachians is efficiently conveyed within the karstic aquifers from the points of recharge to springs. As indicated above, this is an important factor in the rapid transmittal of pollutants through karstic terrane.

Time: The *evolution* of cave systems into well-integrated karstic drainage networks is slow by human standards. In geologic context, however, the process is quite rapid. Nearly all caves in the Appalachian region, regardless of their ultimate size, developed during the Quaternary Period (the last two million or so years), even though the rocks in which they have formed are considerably older, typically dating from the early to middle Paleozoic (525 to 320 million years ago). Within our lifetimes, most flow networks will undergo very little structural change. On the other hand, the environment of caves and karstic flow systems can be rapidly altered by human intervention.

Local Modifying Factors

Lithostratigraphy: As a rule, carbonate rocks are not lithologically uniform. Sequences of sedimentary deposits, such as those of the Appalachian region, show considerable stratigraphic variation. Layers of sandstone, shale,

or chert often separate beds of limestone. Flowpaths may be constrained by relatively impermeable beds, thereby channeling groundwater within carbonate beds or perching karstic groundwater above shale deposits (*see* papers by Saunders and Balfour and Heller, this volume).

Bedrock Structure: Beds of carbonate rocks are not perfectly horizontal; in fact, most beds in the Appalachian region are steeply inclined due to intense folding. Even the slightest tilt (dip) of the strata can exert considerable influence on developing flowpaths. Flow within dipping carbonate beds primarily occurs either along strike (horizontally and parallel to the trend of the bed) or down the dip (descending parallel to the slope of the bed). Which of these may occur in any particular case depends on the configuration of the surface topography, the location of baselevel springs, and the hydraulic gradients imposed by lithologic constraints and fracture porosity. Most groundwater in the Appalachian region of Virginia flows along strike. Because the folded rocks and mountain ranges trend northeast-southwest in this region, it is expected that karstic groundwater flow will generally be in these directions. (Of course, flow down the dip may occur locally; only careful mapping will determine this for sure.) As a result, extensive flow of karstic groundwater typically occurs beneath and parallel to the lower flanks of mountains and hills. Flow may also be concentrated where there is a higher density of fractures and thus a greater permeability.

Topographic Evolution: Subsurficial karstic drainage systems develop in response to an evolving surficial landscape. Local baselevels are lowered as major surface streams cut downward through the landscape by erosion. Contemporaneously, the position of the potentiometric surface (water table) in the aquifer drops as baselevel springs migrate to lower elevations. Previously formed conduits and caves are abandoned by groundwater that now flows through the rock at lower elevations. Relict, water-free caves are no longer a significant component of the drainage system (except during occasional high runoff). Younger, lower conduits have taken their place.

Hydrodynamics: The rate of flow and level of turbulence in karstic groundwater flow is highly variable from place to place. High hydraulic gradients and constricted conduits can increase the pressure, forcing water through the system. This in turn increases turbulence and the rate of dissolution. Steep hydraulic gradients promote greater discharge and rapid transmittal of pollutants.

Climate: Climate has been shown to severely affect the rate of karst development. Karst of tropical regions, for example, forms at a much greater rate than does that of an arid, cold region. Karstification in the Appalachian region has been at a moderate rate. Because climates are relatively constant during the course of human activity, this variable has little effect either spatially or temporally with respect to present-day environmental problems.

Analyzing a Karstic System

A complete analysis of a karst region would require careful documentation of each of the aforementioned ten variables (Kastning, 1990). An appreciation of environmental problems in the Appalachian karstlands requires a familiarity with the origin and configuration of subsurface drainage. Accurate determination of paths of groundwater flow necessitates detailed geologic mapping. The path of flow from the points of recharge (infiltration) to points of discharge (springs) is rarely a straight line and may in reality be quite convoluted. Flowpaths of karst water migration are often contrary to the direction of flow in nearby surface streams and water may be pirated through the subsurface from one surficial drainage basin to another.

Helping the Public Understand Karst

Appropriate karst management must include an assessment of the vulnerability of integrated karst systems to changes incurred on the surface. The most effective way to protect the karst environment is to develop an awareness and understanding of potential problems on the part of the local residents. This translates into education; not simply limited to traditional schooling, but by any means whereby the public is exposed to the problems and solutions. These include developing data for karstland management, promoting clean-up and restoration activities, working with civic organizations, youth groups, museums and science centers, developing displays and presentations for local and regional events, and involving the media (newspapers, television, radio, etc.). Examples of activities in Virginia and West Virginia are included in the discussion that follows. However, the environmental problems are typical of the Appalachian region and the educational solutions used in the Virginias may be easily applied in any of the other states of the Appalachians.

Inventorying Karst

Many karstic landforms, such as large sinkholes and sinking streams, are readily identifiable on standard U.S. Geological Survey 7.5-minute topographic maps (scale 1:24,000). However, not all sinkholes, caves, and lesser karstic landforms appear on topographic maps; for example, many sinkholes are simply too shallow to be represented within the contour interval used on a particular map, or in some cases karst features have simply been overlooked in the surveying or cartographic process. Precise inventory of karst necessitates additional work, including use of low-altitude aerial photography and surface reconnaisance by vehicle or on foot. To date there are very few areas of the country where such systematic inventories have been made. A noteable effort in accomplishing such a task includes recent surveying both on a state-wide and county-wide level in the Commonwealth of Virginia (*see* Hubbard, this volume).

The karst of Virginia and West Virginia is characterized by a high density of sinkholes (Herak and Stringfield, 1972; Hubbard, 1984; Kastning, 1986, 1988, 1991). A series of three maps showing exposures of soluble rock and the distribution of sinkholes and caves is being published for the state of Virginia (Hubbard, 1983, 1988). The data is derived from topographic maps, aerial photography, soil survey maps, and the speleological literature. Karst maps for two of the counties in this karst region have recently been published (Miller and Hubbard, 1986; Hubbard, 1990). Karst terrane occasionally appears as a mapped environmental unit in local geologic mapping as well (*e.g.* Schultz, 1981).

Delineation of sinkholes on a map may readily indicate potential subsurficial flowpaths (Kastning, 1984, 1989b). In many situations, sinkholes are topographically aligned. This indicates a structural or stratigraphic control in the hydrogeologic setting wherein groundwater moves along linear flowpaths formed along bedding planes and fractures. The implication is that infiltration entering an aquifer through such sinkholes contributes water to an integrated flow system. The surface arrangement of sinkholes thereby provides a hint of the configuration of the underground drainage. This concept is easily related to the public through diagrams and maps of the areas in question. However, an experienced geologist or hydrologist must compile this data and reduce it to an easily visualized graphical format showing how surficial and underground processes are related. The graphics can be incorporated in virtually all types of informational material, published or otherwise, including pamphlets, posters, newspaper and magazine articles, handouts at meetings, classroom study aids, and many others.

Flowpaths of groundwater through carbonate rock can be determined by injection of harmless fluorescent dyes into subsurface streams and detecting where the dye emerges. The techniques of water tracing in karst are well documented in the textbooks referred to earlier and in handbooks. Water tracing in the Appalachian karst has been underway since the 1950's. A few recent studies in the Virginias include those of Jones (1973, 1983, and this volume, p. 217), Saunders and others (1981), Ogden (1976), Werner (1981), and Quinlan (this volume, p. 168). As flow networks are delineated through exploration and mapping of caves, groundwater tracing, and other means, they become part of the knowledge of the terrane, allowing assessment of environmental problems that may occur if man modifies the landscape or water chemistry. This, in turn, allows local communities to make educated decisions on landuse and economic development.

Inventories of caves and other significant karst features are being maintained by privately operated speleological surveys in both states. These surveys have been in operation for some time (Davies, 1958; Douglas, 1964; Holsinger, 1975; Virginia Speleological Survey, 1987-present). Recent efforts by the Virginia Cave Board and the Virginia Speleological Survey have identified caves considered to be highly significant based on geologic, biologic, hydrologic, archeologic, and historic criteria (Holsinger, 1985; Gulden, 1989).

Organisms inhabiting caves in West Virginia and Virginia have been investigated and documented (*e.g.* Holsinger and others, 1976; Holsinger and Culver, 1988). Additionally, newly discovered species are being added at a regular rate. Research and publication on cave habitats and ecosystems should be part of the inventory in the conservation and management of karst regions. Caves serving as habitats for sensitive and rare organisms are increasingly brought to light, especially where development is threatening their existence. The Virginia Cave Board in conjunction with the Natural Heritage Program regularly contends with potential threats to cave ecosystems in the state.

Cave and Karst Protection Laws

Fortunately steps are being taken to protect the karstic environment in the Appalachian region. For example, both West Virginia and Virginia have enacted state laws that protect caves and their natural contents from vandalism and contamination. The public at large is not aware of these laws and therefore their existence must be made known. Chapters of the National Speleological Society in the Virginias have placed special metallic signs inside many cave entrances informing the visitor of the laws and penalties for violations.

The Commonwealth of Virginia has established the Virginia Cave Board as part of the Department of Conservation and Recreation to take up matters relating to caves and karst in the Commonwealth, to advise other agencies, and to participate in education related to caves, cave science, and cave conservation. The board, composed of eleven members appointed by the Governor, meets three times a year and regularly considers environmental problems emerging at specific sites. Where necessary, it becomes actively involved in mitigating potential threats to caves, karst, and spelean biota.

Cave and Sinkhole Restoration

Cave restoration projects have become increasingly popular among concerned cavers and other volunteers. Restoration of sinkholes has also been attempted in recent years (*see* Plate D, page 146). One of the largest and most successful efforts was the removal of tons of refuse from the entrance area of Stillhouse Cave in Randolph County, West Virginia (Harler, 1991; Vlchek, 1991). Stillhouse Cave is within a highly scenic and popular karst region and lies just a few hundred feet from the well known Sinks of Gandy (*see* photograph on the cover of this volume). This clean-up was a cooperative effort between the speleological community and the West Virginia Department of Natural Resources and Department of Highways. It resulted in the first site in the official West Virginia "Adopt

a Dump" program for volunteer clean-ups of illegal dumps.

Removal of trash and restoration of original contours around cave entrances have been very successful, but such efforts require considerable time and expense and in many states the shear number of sinkhole dumps is staggering. However, clean-up and restoration projects catch the interest of local newspapers and television stations. News stories about these projects are read by many people who live and work directly on the karst surface. Whether conciously or not, these readers note that if volunteers are expending time and energy in cleaning trash from a hole in the ground, there must be some reason. Of course, it is useful if the reasons are given to the reporters as part of the story.

Local chapters of the National Speleological Society regularly clean caves and sinkholes, leading to favorable publicity in the press. The New River Valley Grotto (Chapter) of the National Speleological Society in conjunction with other grottos and a local Boy Scout troop has been cleaning a large trash-filled sinkhole in Pulaski County, Virginia. This was a former entrance to James Cave, the longest (over 7000 feet) cave in the county. The entrance has recently been reopened (after being buried for 27 years under tons of farm and household trash) and water quality in the cave stream is largely restored. About six clean-up events have taken place at the site to date and, in some cases, newspaper and video coverage has been a planned part of the event (Farrar, 1989; Kittredge, 1989).

Figure 2: Cave Conservation Awareness Award earned by Boy Scouts in Virginia and West Virginia. Requirements for the award include (1) a short course about caves and cave conservation, (2) participation in a clean-up or restoration project involving a cave, sinkhole, or other karst terrane, and (3) a caving trip led by experienced cavers.

This clean-up was featured in a recent videotape produced by the Virginia Military Institute Research Laboratory and funded by the Virginia Environmental Endowment for use on public television and in classrooms throughout the karst region of Virginia (*see* Erchul, this volume).

Involvement of youth groups in cleanups is both sensible and desirable. Environmental awareness is currently at a high level among teachers and students in public schools, and many younger people are willing to volunteer time and energy toward cleaning up their community. Some local Boy Scout troops in the New River Valley of Virginia and West Virginia have established a continuing award program (Figure 2) in cave conservation that includes scheduled clean-ups at caves and sinkholes (*see* Plate E, page 146).

Karst and Public Education

It is impossible for those concerned with preserving karst to confront all of the environmental problems through remedial action, including cleanups of caves and sinkholes, legal action to prevent development or to seek restitution from violators of environmental law, or other reactionary measures. Although these efforts will help on a case-by-case basis, they will not keep pace with the impact of *progress*.

Perhaps the single, most effective program to prevent the abuse of karst and promote sound environmental awareness is within the context of primary and secondary education. How the characteristics and mechanisms of karst differ from those of other terranes must be made graphically clear in the classroom, particularly in counties or cities that lie within karst areas or are in close proximity to them. Educational contact with this age group may also be made through youth programs including scouts, 4-H clubs, high school science clubs, and other outdoor-oriented organizations.

Organized cavers and karst researchers are often approached to lead workshops or fieldtrips for youth groups, science teachers, community leaders, symposia participants, or others. An understanding of karst, including the subsurface, can be enhanced by viewing surficial features (*e.g.* sinking creeks, sinkholes, seeps, and springs) without necessarily venturing underground (*see* Plate A, page 58). Therefore, the public at large, including the news media, could learn a great deal from a fieldtrip on karst terrane.

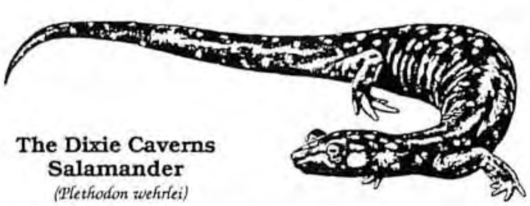
The news media can then effectively join in carrying environmental messages to the general public. Graphic explanations of active karst processes in layman's terms can go a long way toward conveying the need to preserve fragile karst features, water supplies, and cave ecosystems. The use of photography, video-recordings, graphic arts, and writing, especially in conjunction with case histories, has been shown to be effective in reaching citizens living on

karst. Distribution of this information can be in various forms, including presentations of papers or multimedia programs at local, regional, or national meetings, posters, local clean-up and fund-raising events (Figure 3) with attendant publicity in the media, exhibits at commercial caves, museums (*see* Plates I and J, page 186), scout shows, and other community events, and literature for distribution to the public and to landowners.

Some educational materials are readily available from cave conservation organizations, such as the National Speleological Society and the American Cave Conservation Association. Likewise, the Virginia Division of Mineral Resources has published materials on karst designed to educate the public and provide basic data for local communities (Hubbard, 1988, 1989, 1990, 1991; Miller and Hubbard, 1986).

Additionally, the Commonwealth of Virginia, through the Virginia Cave Board, recently published a 22 inch by 28-inch, full-color poster on karst groundwater protection (*see* Plate F, page 152). This poster, depicting the problems created by pollution of karst waters through sinkhole dumping, was distributed free of charge to ninth-grade Earth-Science teachers throughout the Commonwealth, to other educational facilities, and to selected governmental and environmental agencies and groups (Kastning and Kastning, 1990). This poster is also available by request to similar groups in other states.

Thank You For Your Contribution !



The Dixie Caverns Salamander
(Plethodon wehrlei)

The black Dixie Caverns Salamander is a rare phase of the Slimy Salamander (*Plethodon g. glutinosus*) and is found in Dixie Caverns, New Dixie Cave, and Blankenship Cave, all in Roanoke County. This amphibian is one of many species of animals, both vertebrate and invertebrate, that inhabit caves of this area. In protecting caves and the groundwater that circulates through them, we are ensuring that sensitive creatures such as this salamander continue to have the habitat they need for survival. (Drawing by John Schoenherr, card designed by Ernst and Karen Kastning and Steve Lenhart, Radford University.)

Your visit to the Annual
Dixie Caverns Haunted Cave
is contributing to the conservation and protection of caves and their ecosystems.

The **New River Valley Grotto** and **Blue Ridge Grotto**, chapters of the National Speleological Society, use proceeds from the Haunted Cave for local and statewide cave conservation projects and to support national cave protection programs. Caves are a valuable, yet neglected, natural resource in the Appalachian region of Virginia and West Virginia. They are an integral part of the groundwater system in limestone terrane (karst). Conservation of caves, sinkholes, and other karst features is important in the protection of groundwater.

The **National Speleological Society** is dedicated to the safe exploration, study, and conservation of caves. If you are interested in caves and caving, please contact either of these chapters:

New River Valley Grotto, P.O. Box 1027, Dublin, Virginia 24084

Blue Ridge Grotto, c/o Al Stewart, 2528 Montgomery Avenue S.W., Roanoke, Virginia 24015

October 1990

Figure 3: "Thank-you" card presented to visitors to the annual Haunted Cave fund-raising event at Dixie Caverns, Virginia. Proceeds are used for conservation projects. The card is a means to inform the public about caves, karst, and their environmental sensitivity.

Conclusions

One of the best measures of emerging environmental problems in karst regions is the incidence and frequency of news stories on such topics as sinkhole collapses (*see* papers by Dougherty and Beck in this volume), contaminated springs and wells, or accidental spills. This coverage alerts the populace to the extent and distribution of the problem, as it should. Yet, positive environmental education about karst processes is imperative in karst regions such as the Appalachians. Many potentially effective resources and methods are available to concerned citizens who wish to further karst conservation.

Among them are:

- cleanup and restoration projects
- preparation and dissemination of literature on karst and speleology
- inventorying and mapping karst areas
- presentations and exhibits for the public at community events
- working with youth groups, civic organization, and governmental agencies and committees
- volunteer work with show caves, parks, museums, and outdoor organizations
- participation in public awareness activities (*e.g.* Earth Day, litterthons)
- involving the media wherever possible

Educating the public about the environmental liability of karst should be viewed as an opportunity whereby potential problems may be averted or minimized. Instilling an awareness of karst, through remedial, monitoring, and protective actions, coupled with positive educational programs and media involvement, is the most effective means toward managing this sensitive terrane.

References

Aley, T.J., 1972, Groundwater contamination from sinkhole dumps: *Caves and Karst*, v. 14, p. 17-23.

Aley, T.J.; Williams, J.H.; and Massello, J.W., 1972, Groundwater contamination and sinkhole collapse

induced by leaky impoundments in soluble rock terrain: *Missouri Geological Survey and Water Resources, Engineering Geology Series* No. 5, 32 p.

Beck, B.F. (editor), 1984, *Sinkholes: Their Geology, Engineering and Environmental Impact: Proceedings of the First Multidisciplinary Conference on Sinkholes, Orlando, Florida, 15-17 October 1984*: A.A. Balkema, Rotterdam and Boston, 429 p.

Beck, B.F. (editor), 1989, *Engineering and Environmental Impacts of Sinkholes and Karst: Proceedings of the Third Multidisciplinary Conference on Sinkholes and the Engineering and Environmental Impacts of Karst, St. Petersburg Beach, Florida, 2-4 October 1989:* A.A. Balkema, Rotterdam and Boston, 384 p.

Beck, B.F. and Wilson, W.L. (editors), 1987, *Karst Hydrogeology: Engineering and Environmental Applications: Proceedings of the Second Multidisciplinary Conference on Sinkholes and the Environmental Impacts of Karst, Orlando, Florida, 9-11 February 1987*: A.A. Balkema, Rotterdam and Boston, 467 p.

Bögli, A., 1978, *Karst Hydrology and Physical Speleology* (translated from the German by J.C. Schmid): Springer-Verlag, New York, 284 p.

Davies, W.E., 1958, Caverns of West Virginia: *West Virginia Geological and Economic Survey*, volume 19, 350 p. (reprinted in 1965 with 72-page supplement as volume 19A).

Davies, W.E., 1970, Karstlands, *in* United States Geological Survey, *National Atlas of the United States of America*, sheet 77.

Dougherty, P.H. (editor), 1983, *Environmental Karst* (papers from karst symposium at the Association of American Geographers meeting, Louisville, Kentucky, April 1980): GeoSpeleo Publications, Cincinnati, Ohio, 167 p.

Douglas, H.H., 1964, *Caves of Virginia*: Virginia Cave Survey, Falls Church, Virginia, 761 p.

Dreybrodt, W., 1988, *Processes in Karst Systems Physics, Chemistry and Geology:* (Springer Series in Physical Environment 4): Springer-Verlag, New York, 288 p.

Farrar, B., 1989, Cave becomes an area time capsule: *The Southwest Times* (Pulaski, Virginia), v. 86, no. 73 (27 March 1989), p. 1-2.

Ford, T.D. and Cullingford , C.H.D. (editors), 1976, *The Science of Speleology*: Academic Press, New York, 593 p.

Ford, D.C. and Williams, P., 1989, *Karst Geomorphology and Hydrology*: Unwin Hyman, Winchester, Massachusetts, 320 p.

Gulden, R., 1989, Lists of long caves of the United States and the world (unpublished): National Speleological Society Committee on Long and Deep Caves.

Harler, C., 1991, Stillhouse Cave: Cleveland Grotto clean-up: Nearby grottos help make project a success: *NSS News* (National Speleological Society), v. 49, p. 172-175.

Herak, M. and Stringfield, V.T. (editors), 1972, *Karst: Important Karst Regions of the Northern Hemisphere*: Elsevier Publishing Company, New York, 551 p.

Hill, C.A. and Forti, P., 1986, *Cave Minerals of the World*: National Speleological Society, Huntsville, Alabama, 238 p.

Holsinger, J.R., 1975, Descriptions of Virginia caves: *Virginia Division of Mineral Resources Bulletin 85*, 450 p. plus 7 plates.

Holsinger, J.R., 1985, *Annotated List of Significant Caves and Karst Areas in Virginia*: Virginia Speleological Survey (limited distribution document, revised April 1985), 251 p.

Holsinger, J.R. and Culver, D.C., 1988, The invertebrate cave fauna of Virginia and a part of eastern Tennessee: Zoogeography and ecology: *Brimleyana* (The Journal of the North Carolina State Museum of Natural Sciences), no. 14 (June 1988), p. 1-162.

Holsinger, J.R.; Baroody, R.A.; and Culver, D.C., 1976, The invertebrate cave fauna of West Virginia: *West Virginia Speleological Survey Bulletin*, no. 7, 82 p.

Hubbard, D.A., Jr., 1983, Selected karst features of the northern Valley and Ridge province, Virginia: *Virginia Division of Mineral Resources Publication*, no. 44, one sheet (scale 1:250,000).

Hubbard, D.A., Jr., 1984, Sinkhole distribution in the central and northern Valley and Ridge province, Virginia, *in* Beck, B.F. (editor), *Sinkholes: Their Geology, Engineering and Environmental Impact: Proceedings of the First Multidisciplinary Conference on Sinkholes, Orlando, Florida, 15-17 October 1984:* A.A. Balkema, Rotterdam and Boston, p. 75-78.

Hubbard, D.A., Jr., 1988, Selected karst features of the central Valley and Ridge province, Virginia: *Virginia Division of Mineral Resources Publication*, no. 83, one sheet (scale 1:250,000).

Hubbard, D.A., Jr., 1989, *Sinkholes:* Virginia Division of Mineral Resources brochure, 2 p.

Hubbard, D.A., Jr., 1990, Geologic map of Clarke Coun-

ty, Virginia - Plate 1; Map of hydrogeologic components for Clarke County, Virginia - Plate 2: *Virginia Division of Mineral Resources Publication*, no. 102, 2 map sheets (scale 1:50,000).

Hubbard, D.A., Jr., 1991, *Caves and Caving in Virginia:* Virginia Division of Mineral Resources brochure, 2 p.

Huppert, G.N., 1988, Cave and karst related theses in United States and Canadian universities: 1899-1988: *Cave Geology*, v. 2, no. 1, 82 p.

Jennings, J.N., 1985, *Karst Geomorphology* (revised and expanded edition of Jennings, 1971): Basil Blackwell, Oxford and New York, 293 p.

Jones, W.K., 1973, Hydrology of limestone karst in Greenbrier County, West Virginia: *West Virginia Geological and Economic Survey Bulletin*, no. 36, 49 p. plus 2 plates.

Jones, W.K., 1983, Karst hydrology in West Virginia (1983), *in* Medville, D.M.; Dasher, G.R.; and Werner, E. (editors), An introduction to the caves of east-central West Virginia: *National Speleological Society Guidebook Series*, no. 23, p. 25-34.

Kastning, E.H., 1977, Faults as positive and negative influences on groundwater flow and conduit enlargement, *in* Dilamarter, R.R. and Csallany, S.C. (editors), *Hydrologic Problems in Karst Regions*: Western Kentucky University, Bowling Green, Kentucky, p. 193-201.

Kastning, E.H., 1984, Hydrogeomorphic evolution of karsted plateaus in response to regional tectonism, *in* LaFleur, R.G. (editor), *Ground Water as a Geomorphic Agent: International Series*, no. 13, Allen and Unwin, Inc., Boston (Proceedings of the 13th Annual ("Binghamton") Geomorphology Symposium, Rensselaer Polytechnic Institute, Troy, New York, September, 1982), p. 351-382.

Kastning, E.H., 1986, Cave regions of the United States of America, *in* Middleton, J. and Waltham, A., *The Underground Atlas: A Gazetteer of the World's Cave Regions:* Robert Hale, Limited, London, p. 203-220.

Kastning, E.H., 1988, Karst of the New River Drainage basin, *in* Kardos, A.R. (editor), *Proceedings, Seventh New River Symposium, Oak Hill, West Virginia, April 7-9, 1988:* New River Gorge National River, Oak Hill, West Virginia, p. 39-49.

Kastning, E.H., 1989a, *Karst Geomorphology and Hydrogeology: A Bibliography of Principal References*: Limited private printing, 8 p (available from the author).

Kastning, E.H., 1989b, Surficial karst patterns: Recognition and interpretation, *in* Beck, B.F. (editor), *Engineering and Environmental Impacts of Sinkholes and Karst: Proceedings of the Third Multidisciplinary Conference on Sinkholes and the Engineering and Environmental Impacts of Karst, St. Petersburg Beach, Florida, 2-4 October 1989:* A.A. Balkema, Rotterdam and Boston, p. 11-16.

Kastning, E.H., 1989c, Environmental sesnsitivity of karst in the New River drainage basin, *in* Kardos, A.R. (editor), *Proceedings, Eighth New River Symposium, Radford, Virginia, 20-22 1989:* New River Gorge National River, Oak Hill, West Virginia, p. 103-112.

Kastning, E.H., 1990, Virginia karst terrains: Unique problems associated with waste management and groundwater protection, *in* Erchul, R.A. (editor), *Proceedings of the Symposium for Virginia Localities on Waste Management and Groundwater Protection, April 3-4, 1990 at the Virginia Military Institute, Lexington, Virginia*: VMI Research Laboratories, Inc., Lexington, p. 82-93.

Kastning, E.H., 1991, *Caves, Karst, and Environmental Impact in the New River Drainage Basin of Virginia and West Virginia: Guidebook for a Geologic Fieldtrip, Appalachian Karst Symposium, Radford, Virginia, 23 March 1991:* Radford University, Department of Geology, Radford, Virginia, 36 p.

Kastning, K.M. and Kastning, E.H., 1990, *In Karstlands... What Goes Down Must Come Up!:* Virginia Cave Board, Department of Conservation and Recreation, poster, 22" by 28".

Kittredge, K., 1989, No place for trash: *New River Current*, v. 1, no. 279, (13 January 1989), p. 1-2 (section of the *Roanoke* (Virginia) *Times and World-News*).

Krinitzsky, E.L., 1947, A fault-plane cavern: *Journal of Geology*, v. 55, p. 107-119.

LaMoreaux, P.E. (editor-in-chief); Tanner, J.M.; and ShoreDavis, P., 1986, Hydrology of limestone terranes: Annotated bibliography of carbonate rocks (volume 3): *International Association of Hydrogeologists, International Contributions to Hydrogeology*, v. 2, 341 p.

Lamoreaux, P.E., Warren, W.M., and others, 1970 and 1975, Annotated bibliography of carbonate rocks and hydrology of limestone terranes: *Geological Survey of Alabama Bulletin* no. 94, parts A and E, 242 p. and 168 p.

LeGrand, H.E., 1973, Hydrological and ecological problems of karst regions: *Science*, v. 179, no. 4076 (2 March 1973), p. 859-864.

Lovegrove, R., 1988, Tanker spills 3,000 gallons of fuel: *Roanoke* (Virginia) *Times and World-News*, v. 25, no.

21, (21 July 1988), p. B6.

Miller, E.V. and Hubbard, D.A., Jr., 1986, Selected slope categories and karst features map of Giles County, Virginia: *Virginia Division of Mineral Resources Publication 70*, map, scale 1:50,000, 1 sheet.

Monroe, W.H., 1970, A glossary of karst terminology: *United States Geological Survey Water-Supply Paper* 1899-K, 26 p.

Ogden, A.E., 1976, *The hydrogeology of the central Monroe County karst, West Virginia:* Ph.D. dissertation (unpublished), West Virginia University, 263 p.

Saunders, J.W.; Ortiz, R.K.; and Koerschner III, W.F., 1981, Major groundwater flow directions in the Sinking Creek and Meadow Creek drainage basins of Giles and Craig Counties, Virginia, U.S.A., *in* Beck, B.F. (editor), *Proceedings of the Eighth International Congress of Speleology, Bowling Green, Kentucky, July 18-24, 1981*: National Speleological Society, Huntsville, Alabama, v. 1, p. 398-400 (reprinted in *The Tech Troglodyte*, 1985, v. 24, no. 2, p. 54-60).

Schultz, A.P., 1981, *Geology of the City of Radford*: unpublished map (scale 1:7200), 2 sheets (copies available from the City Engineer's office, City of Radford).

Slifer, D.W., 1987, Rockbridge's illegal dumps: *Focus on Water* (Virginia Water Resources Research Center), no. 2, 8 p.

Slifer, D.W. and Erchul, R.A., 1989, Sinkhole dumps and the risk to ground water in Virginia's karst areas, *in* Beck, B.F. (editor), *Engineering and Environmental Impacts of Sinkholes and Karst: Proceedings of the Third Multidisciplinary Conference on Sinkholes and the Engineering and Environmental Impacts of Karst, St. Petersburg Beach, Florida, 2-4 October 1989:* A.A. Balkema, Rotterdam and Boston, p. 207-212.

Sweeting, M.M., 1973, *Karst Landforms*, Columbia University Press, New York, 362 p.

Trudgill, S., 1985, *Limestone Geomorphology* (Geomorphology Texts, No. 8), Longman, London and New York, 196 p.

Virginia Speleological Survey, 1987-present: *Virginia Cellars*, v. 1-.

Vlchek, F., 1991, More than garbage: *NSS News* (National Speleological Society), v. 49, p. 176.

Waltham, A.C., 1989, *Ground Subsidence*: Blackie and Son, London, 224 p.

Werner, E., 1981, *Guidebook to the karst of the central Appalachians: Prepared for the Eighth International Congress of Speleology, Bowling Green, Kentucky, U.S.A., July 18 to 24, 1981:* National Speleological Society, Huntsville, Alabama, 51 p.

Werner, E., 1983, Effects of highways on karst springs - An example from Pocahontas County, West Virginia, *in* Doughterty, P.H. (editor), *Environmental Karst:* GeoSpeleo Publications, Cincinnati, p. 3-13.

White, W.B., 1988, *Geomorphology and Hydrology of Karst Terrains*, Oxford University Press, New York, 464 p.

White, W.B. and White, E.L., 1984, Cave and karst-related papers in the mainstream scientific literature: A bibliography: *Cave Geology*, v. 1 no. 9, p. 291-392.

Regional Karst Studies: Who Needs Them?©

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ABSTRACT

Regional karst studies delineating areas of carbonate bedrock, sinkholes, caves, and permanent stream segments serve to indicate the degree to which areas, underlain by carbonate bedrock, have undergone karstification. There are three potential hazards associated with karst: subsidence, groundwater pollution, and flooding. To the inhabitants of karst areas, the loss of life and property to sinkhole collapse is probably the most commonly perceived potential karst hazard. Karst is an unstable terrain typified by differential subsidence with respect to sinkholes or the highly irregular (pinnacled) soil-bedrock interface. Because of the solutional origin of karst topography and its characteristic sinkhole, cave, and pinnacle features, the close relationship between karst and groundwater should be apparent. The linkage of sinkholes and groundwater by caves or smaller conduits, the relatively rapid movement of groundwater in conduit components of the aquifers, and the directional, but commonly unpredictable, nature of karst aquifers contribute to the high susceptibility of karst aquifers to pollution. Sinkhole-flooding hazards are most common in karst areas near base level; however, the use of sinkholes for drainage outfalls and poor management of siltation during construction can result in sinkhole ponding and flooding at any elevation.

There is an increasing awareness of a decline of groundwater quality in Virginia, especially in karst areas. Planners and local governmental officials are indicating concerns about the causes of groundwater pollution and how it can be arrested. Regional karst studies, used in combination with geologic and hydrologic information, provide the key to protecting groundwater resources from further degradation in the Valley and Ridge province of Virginia. Karst terrains are highly susceptible to pollution from waste disposal, leaks and spills of chemicals and fuels, and the improper application of agricultural chemicals. The containment and transportation of chemicals, fuels, and wastes poses additional risk to karst aquifers because of the instability inherent to karst terrains.

Introduction

Thick sequences of various types of carbonate rock are exposed to weathering in the Valley and Ridge physiographic province of Virginia. Virtually all exposed carbonate rock in Virginia has undergone some degree of karstification. The purpose of regional karst mapping is the delineation of karst areas and a relative characterization of the degree of karstification in each area. Regional karst mapping has been completed over approximately two thirds of the Valley and Ridge physiographic province at a scale of 1:250,000. The three types of hazards associated with karst are subsidence, flooding, and pollution of groundwater. Subsidence hazards have traditionally dominated the fears of residents of karst areas. Recently, concerns over the pollution of groundwater have prompted local governments to seek planning and land-use restriction as solutions to protect their karstic groundwater resources.

Distribution and Mapping of Karst

The significant karst in Virginia is developed on the folded and faulted sedimentary carbonate rock types (Figure 1, L's) in parts of 27 counties in the Valley and Ridge physiographic province. Sinkholes, caves, pinnacles, and subterranean drainage are indicative of the solutional origin of this karst topography; however, travertine-marl deposits are a common accretionary feature. Other karst areas include narrow belts of marble (Figure 1, M's) in both the Blue Ridge and Piedmont physiographic provinces and partially indurated shelly sands (Figure 1, S's) of the Yorktown Formation in the Coastal Plain physiographic province. Karst features associated with the marbles include sinkholes, pinnacles, subsurface drainage, and possibly a few caves. Sinkhole development in the Coastal Plain

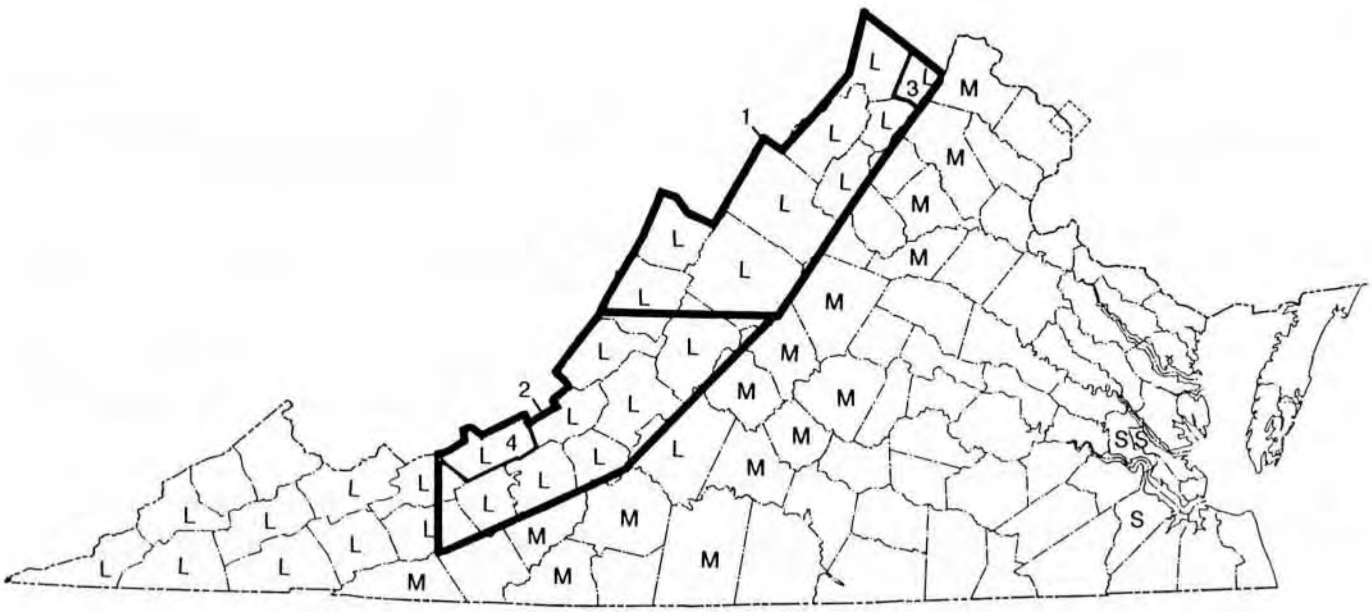

Figure 1. Distribution of karst in Virginia: L, M, and S indicate counties in which limestone and dolomite, marble, and shelly sands, respectively, have undergone some degree of karstification. Regional karst maps have been published for areas 1 (Hubbard, 1983) and 2 (Hubbard, 1988), and county karst maps published for Clarke (3; Hubbard, 1990) and Giles (4; Miller and Hubbard, 1986) counties.

province is not limited to the outcrop extent of the Yorktown Formation; sinkholes are also found in the Bacons Castle, Windsor, and Charles City formations overlying the Yorktown carbonate-rich sands (Johnson and others, 1987). Subsurface drainage, pinnacles, and at least one travertine deposit are also known, but no hard evidence of extensive lateral solutional conduit development has been documented.

Regional mapping of karst in Virginia has characterized karst intensity by the location and distribution of sinkholes and caves in areas underlain by carbonate rocks in two thirds of the Valley and Ridge province of Virginia (Figure 1, areas 1 and 2). Two published 1:250,000-scale maps (Hubbard, 1983, 1988) indicate the relative degree of karstification. Geologic information and cave locations were compiled from published and unpublished sources. Sinkholes were located by stereoscopic examination of aerial photography previously used for photoinspection and revision of 1:24,000-scale (7.5-minute) topographic maps. This method yields significantly more sinkholes than simply using those indicated on the topographic maps because the contour intervals on the maps are too large to provide sufficient resolution. Thorough ground surveys, such as soil surveys, provide the most accurate representation of sinkholes, but accurate positioning of sinkholes onto base maps can be a problem, and pseudosinkholes such as old excavations and sag ponds may not be distinguishable from true sinkholes. A comparison of techniques used in Clarke and Giles counties reveals that only 19% of the sinkholes identified from photography are located on the topographic maps and only 19% of the sinkholes identified in the soil survey are identifiable on the photography or only 3% of the soil-survey sinkholes can be found on the topographic maps of Clarke County. Similarly, only 34% of the sinkholes identified from photography are located on the topographic maps of Giles County.

County karst maps, depicting sinkholes and caves, for Clarke (Figure 1, area 3; Hubbard, 1990) and Giles (Figure 1, area 4; Miller and Hubbard, 1986) counties have been published at a 1:50,000 scale. Sinkhole and cave information was determined by the same methods as for the regional karst maps. Additional work with sinkhole locations in Rockbridge (Ron Erchul, 1986, oral communication) and Montgomery (John Flynn, 1991, oral communication) counties has been conducted on 1:24,000-scale topographic maps. Both regional and county karst maps are useful in determining the relative degree of karst development for an area; however, these tools are for preliminary assessment and are not intended to replace site-specific evaluations.

Potential Karst Hazards

There are three potential hazards associated with karst: subsidence, pollution of groundwater, and flooding. Karst terrains are inherently unstable and are typified by differential subsidence, expressed as sinkholes. Man-made structures also may undergo differential subsidence if their foundations are not designed for the highly irregular (pinnacled) soil-bedrock interface common in the Valley and Ridge karst areas. The greatest perceived hazard to inhabitants of karst areas is subsidence. Although sinkhole formation is a natural process in karst, man is not necessarily an innocent victim. Man-induced hydrological alterations com-

monly trigger the mechanisms of sinkhole development (Sowers, 1976; Hubbard, 1989). Typically, surface water (runoff, drainage outfalls, or leaks) is introduced or rerouted and percolates to solutionally enlarged bedrock fractures and creates voids that eventually stope to the surface, or fluctuations in the groundwater levels (water well pumping or extended droughts) drain water-filled voids that stope to the surface to form sinkholes.

Sinkhole flooding is not a major problem in the karst areas of Virginia, but may result from the constriction or plugging of sinkhole drains or by the overwhelming of these natural drains by increases in runoff from artificial surfaces. Inadequate control of erosion during construction can permit sediment-laden runoff to restrict or plug sinkhole drains. The increased runoff from residential, commercial, or industrial surfaces such as from roads, parking lots, and man-made structures is significant (Aley and Thomson, 1981) and can overwhelm the capacity of nearby sinkhole drains and connecting subsurface conduits. Sinkhole flooding occurred after the development of a shopping center and housing in sinkholes in the Fairlawn area of Pulaski County, Virginia in 1980 (R.B. West, 1980, oral communication). Litigious problems arise from flooding associated with housing developments (Quinlan, 1984).

The most extensive hazard associated with the Valley and Ridge karst terrains is groundwater contamination. The karst aquifers are complex reservoirs containing diffuse (slow) and conduit (fast) flow components in a "black box" arrangement. Contaminants introduced (by sinkholes, caves, or thin soil over fractured bedrock) into the karstic aquifer may rapidly appear at springs, cave streams, or water wells or may not appear at all. Contaminant inputs in karst aquifers do not disappear, they just may not appear where they can be readily observed or are expected. The appropriate question in groundwater monitoring in karst terrains is not "can we detect the tracers or contaminants?", but "where are the tracers or contaminants going?" In karst terrains, "into the ground" is synonymous with "into the groundwater."

The three karst hazards of subsidence, flooding, and pollution of groundwater suggest that karst terrains are not well suited to human occupation. Karst areas are unstable with respect to topography as well as groundwater quality and the degree to which this instability is apparent is probably grossly proportional to the level to which man has developed these areas.

Governmental Concerns in Karst

In the early 1980s, there was a subtle shift in the requests by governmental authorities for geologic information in the siting of water wells for public water supply, especially from Augusta, Rockingham, Smyth, and Wise counties. Not only were groundwater resources sought, but concerns were expressed about the integrity of karst-aquifer recharge and if groundwater could be protected against contamination in karst areas. In 1986, county administrators and regional planners sought hydrogeologic information to formulate a groundwater protection plan for Clarke County. Environmental concerns about karstic groundwater quality have since been expressed in Botetourt County. Over the last year, concerns for the protection of karstic groundwater quality have been expressed by Montgomery, Shenandoah, and Page counties.

Repetitive themes of inquiry include: How are karstic groundwater resources polluted?; What can or cannot be put in sinkholes?; What should we do about sinkholes?; How can we protect and insure the quality of our groundwater resources?

Utilizing Regional Karst Mapping

Regional karst maps in the Valley and Ridge physiographic province of Virginia indicate the extent of the carbonate rocks as well as the distribution of sinkholes and caves. A reasonable assumption is that the greater the density of sinkholes and caves the greater the potential for subsidence and groundwater-pollution hazards; however, the absence of sinkholes and caves does not indicate that no subsidence or groundwater-pollution hazards exist. Regional karst maps depict a spectrum of the sinkholes observable on aerial photography and the location of known caves. Unmapped sinkholes exist as well as unreported caves. The on-site absence of sinkholes and caves does not indicate a lack of karst. Herein lies the importance of mapping the extent of carbonate rock, as indicated on the regional karst maps, because all exposed carbonate rock in the Valley and Ridge province of Virginia has undergone some degree of karstification. Land modification, contour smoothing, and filling in sinkholes and caves may alter the appearance of the land, but the potential for subsidence, flooding, and pollution of groundwater still exists and may in fact have been increased.

Karst maps are most effective when utilized in conjunction with geologic and soils maps. Correlations between patterns of karst features and rock or soil units may imply an increased risk of subsidence or groundwater-pollution hazards. Specific rock types may contain relatively higher or lower densities of solutional features (sinkholes and caves); such trends may be especially manifested in folded and faulted geologic structures. Rock units displaying relatively high densities of solutional features may represent important recharge areas for the groundwater aquifer; however, karstic aquifers are quite complex and even detailed hydrological studies may lead to only an elementary understanding of a specific karst area. Soils maps can provide information about the distribution of soils too impermeable for septic-tank drainage fields; unfortunately soils that do not percolate slowly enough to adequately filter septic-tank effluents are not indicated.

Karst areas are poor sites for waste disposal and storage. Leachate and effluents from disposal and storage sites

present a potential groundwater-pollution hazard in karst terrains. Regional karst maps may aid in the preliminary siting of waste-disposal and waste-storage facilities. Certain chemicals, including some defoliants and pesticides should not be used in karst areas; liquid hydrocarbons and other chemicals may pose substantial risks to groundwater resources from improper storage, use, or disposal in carbonate terrains. Storage and pipeline transfer of some liquid chemicals, fuels, and hazardous wastes may represent a considerable risk to groundwater resources in intensely karsted areas, especially where storage and pipelines are not designed with respect to subsidence or containment of leakage. The first key to understanding how to minimize karst hazards is to determine where the karst is.

References

Aley, T. and Thomson, K.C., 1981, *Hydrogeologic Mapping of Unincorporated Greene County, Missouri, to Identify Areas Where Sinkhole Flooding and Serious Groundwater Contamination Could Result from Land Development*: Project Summary prepared for Greene County Sewer District, by Ozark Underground Laboratory, Protem, Missouri, under contract with Missouri Department of Natural Resources, 11 p.

Hubbard, D.A., Jr., 1983, Selected karst features of the northern Valley and Ridge province, Virginia: *Virginia Division of Mineral Resources Publication 44*, one sheet.

Hubbard, D.A., Jr., 1988, Selected karst features of the central Valley and Ridge province, Virginia: *Virginia Division of Mineral Resources Publication 83*, one sheet.

Hubbard, D.A., Jr., 1989, Sinkholes: *Virginia Division of Mineral Resources Brochure*, 2 p.

Hubbard, D.A., Jr., 1990, Map of hydrogeologic components for Clarke County, Virginia: *Virginia Division of Mineral Resources Publication 102*, plate 2.

Johnson, G.H.; Burmester, J.L.; McKay, W.A.; and Wallister, W.G., 1987, Distribution and origin of sinkholes in the Atlantic Coastal Plain of southeastern Virginia (abstract): *Geological Society of America, Abstracts with Programs*, v. 19, no. 2, p. 91.

Miller, E.V. and Hubbard, D.A., Jr., 1986, Selected slope categories and karst features map of Giles County, Virginia: *Virginia Division of Mineral Resources Publication 70*, one sheet.

Quinlan, J.F., 1984, Litigious problems associated with sinkholes, emphasizing recent Kentucky cases alleging liability when sinkholes were flooded, *in* Beck, B.F. (editor), *Sinkholes: Their geology, Engineering and Environmental Impact: Proceedings of the First Multidisciplinary Conference on Sinkholes, Orlando, Florida, 15-17 October 1984*, A.A. Balkema, Rotterdam, and Boston, p. 293-296.

Sowers, G.F., 1976, Mechanisms of subsidence due to underground openings, *in* Subsidence over mines and caverns, moisture and frost actions, and classification: *Transportation Research Board Record 612*, National Academy of Science, Washington D.C., p. 2-8.

Spatial–Temporal Characteristics of Karst Subsidence in the Lehigh Valley of Pennsylvania

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ABSTRACT

Causes of karst subsidence and the geographical distribution of this phenomena are analyzed for the Lehigh Valley, the local name in eastern Pennsylvania for the Great Valley of the Appalachians. This is one of the most densely populated areas in the Appalachians and it is one of the areas most prone to karst damage in the United States with over $1,000,000 damage occurring yearly. Four karst-related deaths have occurred this century, the latest as recently as 1990. This study is a necessary first step in understanding the processes responsible for karst subsidence in order that mitigation techniques can be initiated.

An examination of data bases supplied by local utilities and municipal agencies, supplemented by field inspection and analysis of newspaper reports, topographic maps, and air photos, has ascertained that the Richenback Formation has the highest density of sinkhole development. The Leithsville Formation has the lowest density of sinkholes per square kilometer, owing to its more dolomitic nature. The most damaging subsidences also occur on the Leithsville Formation because it has a deep colluvial cover that, in places, exceeds 30 meters and allows the formation of suffosion dolines. Causes of local subsidence are highly correlated with human activity and include: road construction, general construction, faulty utility-line installation, groundwater drawdown from wells, reactivation of covered dolines, faulty design of detention basins and runoff swales, and leaking water and sewer lines. There is a definite seasonal correlation between doline formation and the early "spring thaw" of late January and February, followed by a secondary period of subsidence in summer due to the decline in groundwater level.

Introduction

Newspaper headlines and stories in other news media in the Lehigh Valley make numerous references to "sinkholes" and karst collapse. Over $1,000,000 in karst-related damage occurs yearly in the Lehigh Valley, that part of the Great Valley of the Appalachians that cuts diagonally from northeast to southwest across Pennsylvania. In the United States, this amount of damage is second only to the much larger karst-prone area of central Florida. Some collapse episodes within the past five years have resulted in loses in excess of $500,000 each: the Macungie sinkhole (Dougherty and Perlow, 1988), the Vera Cruz road collapse (Bonaparte and Berg, 1987), and the Allentown church disaster (Clark and Reaman, 1988). Because of the high density of population, there is also a danger to human life. Three lives were lost in a 1925 collapse in the City of Allentown (Wittman,1988), and another death and seven injuries resulted from the collapse of two townhouses and an accompanying gas explosion on August 29, 1990 (Cassler, 1990).

It is not unusual to see headlines in local newspapers like "Residents flee street–gobbling Macungie sinkhole" (Buzgon, 1986), "Another day in the Valley, another sinkhole" (Whelan, 1986), "30-foot sinkhole opens in shopping center" (Morning Call, August 4, 1986), "Emergency work at Upper Saucon sinkhole complete" (Morning Call, November 4, 1986), "PennDOT says it's not to blame in latest sinkhole in Upper Saucon" (Darrah, 1987), "Another U. Saucon road is affected by sinkhole" (Morning Call, March 7, 1987), "Sinkhole threatens to undermine Northampton Borough home" (Berton, 1987), "Muhlenberg dormitory gets that sinking feeling" (Youngwood, 1988), "City church collapses into sinkhole" (Clark and Reaman, 1988), "City firm hired to fill sinkholes at ABE Airport" (Cowen, 1988), "Lower Nazareth woman files lawsuit over sinkholes" (Morning Call, July 27, 1989), and many more. The news reports only indicate the largest and most disastrous sinkhole occurrences in the area because most collapses are not reported in the news media. Perusal of roadmaster records in suburban and rural townships show the problem to be much greater than indicated by the news

media. Local residents are also eager to share accounts of their favorite neighborhood sinkhole collapse, telling stories about missing dogs and disappearing back yards, and even a humorous account of a football coach at a high school football game who, while pacing the sidelines, was engulfed waist deep in a sinkhole. The above referenced and personal accounts show that there are numerous collapses in this region and that the problem bears investigation because of its economic and life threatening impact.

In order to minimize the loss of life and the destruction of property, local government officials must know what causes subsidence-collapse. Clues to the formation of collapses can be found in an analysis of their spatial and temporal distribution. It is important from a planning perspective to know what areas are the most prone to collapse so that zoning and subdivision ordinances can be written in such a way as to minimize the danger from subsidence. The temporal aspect is also important so that emergency-service organizations can plan for the possibility of a period of greater collapse activity. Therefore, it is the purpose of the current research to investigate the causes of karst collapse, the spatial distribution of occurrences, and the temporal aspects in the Lehigh Valley. The Lehigh Valley is representative of the Great Valley of the Appalachians, and information from this study can be applied to the Lebanon Valley, East Penn Valley, Shenandoah Valley and other similar areas in the Great Valley of the Appalachians. This is especially true of the Reading, Harrisburg, and Hershey urban areas of Pennsylvania where the stratigraphic profile is similar.

Study Area

The Lehigh Valley is generally considered to be that part of the Great Valley of the Appalachians extending from the Delaware River on the east to the Schuylkill River on the west (Figure 1). This encompasses Lehigh and Northampton counties and includes the Allentown/Bethlehem/Easton metropolitan area with a population of nearly three–quarters of a million (Joint Planning Commission, 1991). If nearby Reading is included, the urban area exceeds one million people, most of whom live on the limestone of Lehigh Valley. Contrary to the negative publicity the area received from the song "Allentown" by Billy Joel, the area is not depressed and withering on the vine. It is a dynamic urban area that has been stimulated by the recent completion of Interstate 78. The area already contains the Northeast Extension of the Pennsylvania Turnpike and other major highways such as routes 22, 100, and 309. With easy access to New York City and Philadelphia, the region has experienced substantial growth as a warehousing center. Inexpensive office space has resulted in an influx of tertiary activities that have replaced jobs lost in the shrinking heavy industrial base. Expansion has resulted in an 8.1% increase in population over the past ten years (Joint Planning Commission, 1991). That increase, along with movement from the core cities to the suburbs, is resulting in increased urban sprawl. Figure 2 shows the urban land use that is concentrated on the limestone lowlands, a use which may not be compatible with the karst landscape.

The Lehigh Valley is a distinct physiographic region located between South Mountain of the Blue Ridge Province, composed of Pre–Cambrian and Cambrian–Ordovician granitic gneiss, quartzite and sandstone, and Blue Mountain, the first ridge of the Appalachians, composed of

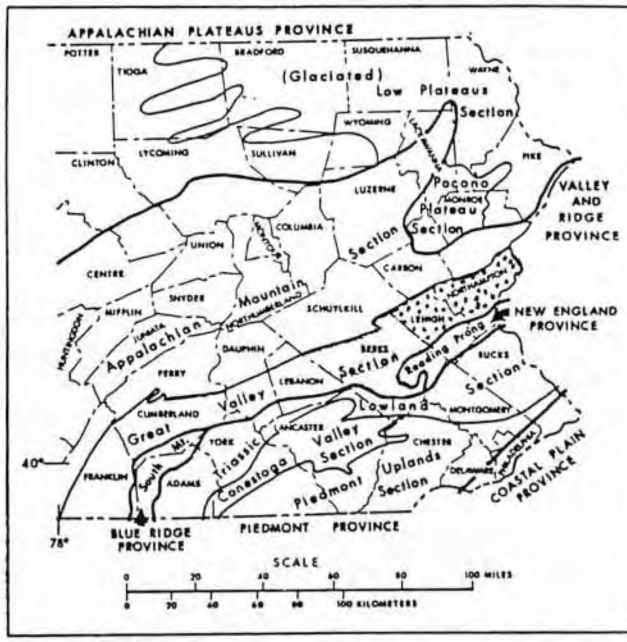

Figure 1: Location of the Lehigh Valley study area, Pennsylvania (shown shaded). Map from Wood and others, 1972.

Figure 2: Urban land use in the Lehigh Valley, Lehigh and Northampton counties, Pennsylvania. Map from Joint Planning Commission, 1991.

Silurian sandstone and conglomerate, partially metamorphosed to quartzite. Between the two resistant ridges lies a 25-km-wide valley having over 400 m in relief. The valley floor is composed of the Martinsburg Formation shale forming a high structural bench of undulating topography on the northwestern side of the valley, and limestone on the southeastern side of the valley forming a flat agricultural plain stretching to the base of South Mountain (Miller and others, 1942). Figure 3 shows the geologic formations of the Lehigh Valley west of Allentown, extending from Kutztown in the south to Slatington in the north (Lash and others, 1984).

Spatial Attributes of Karst Collapse

Karst collapse in the Lehigh Valley is restricted to the limestone belt on which most of the urban development is located. There are six limestone formations within this zone. The shaly limestone of the Jacksonburg Formation in the northwest has several Portland Cement quarries. This is followed in sequence by the progressively older Epler, Rickenbach, Allentown, and Leithsville formations. Table 1 shows the thickness and characteristics of the formations in the Allentown area (Myers and Perlow, 1986). From a cursory examination, one should expect to find many sinkholes in the Allentown Formation because of its great thickness and large aerial extent. Fewer sinkholes should occur on the shaly Jacksonburg Formation because it is thinly bedded and impure. In addition, there should be few sinkholes on the Leithsville Formation because it is highly dolomitized and is covered by an extensive South Mountain colluvium (exceeding 30 meters in places). The

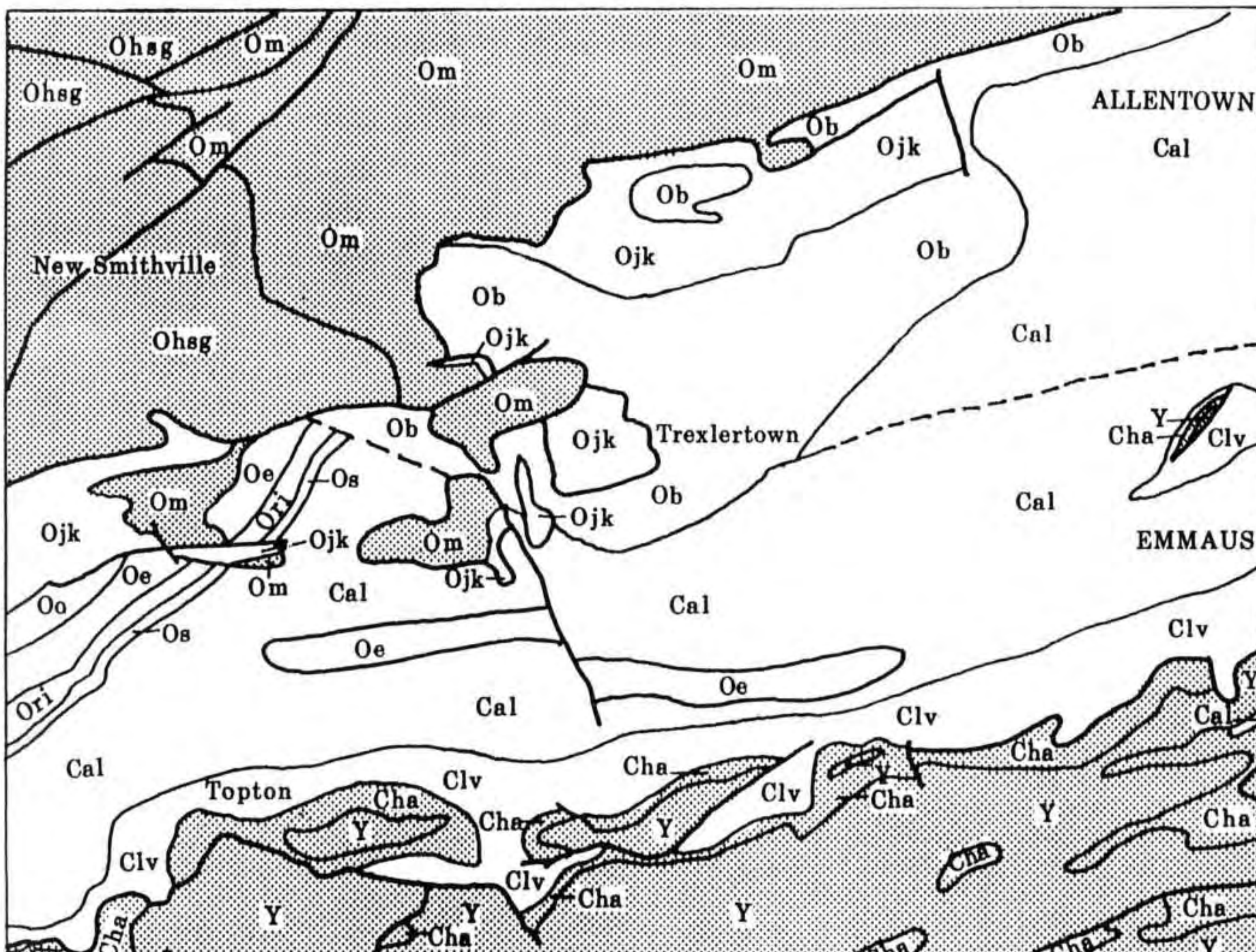

Figure 3: Geologic formations of the Lehigh Valley west of Allentown, PA. Shaded areas to the northwest are the shale hills composed of the Hamburg sequence (Ohsg) and the Martinsburg Formation (Om). The limestone lowland of the Lehigh Valley is in white and is composed of the Jacksonburg Formation (Ojk), Ontelaunee Formation (Oo), Epler Formation (Oe), Rickenbach Formation (Ori), Stonehenge Formation (0s), Allentown Formation (Cal), and the Leithsville Formation (Clv). The shaded areas to the southeast are the Hardyston Formation (Cha) and the undifferentiated gneisses of South Mountain. (Source: Berg and Dodge, 1981.)

Ontelaunee is a minor formation in aerial extent and therefore has few sinkholes.

Table 2 shows the results of a study of sinkhole occurrence in the Lehigh valley (Myers & Perlow, 1986). Data for the study was taken from topographic maps, aerial photographs, utility company records, and records of local government and engineering offices. A total of 1574 sinkholes were identified. It is not surprising to find the largest number of sinkholes on the Epler and Allentown formations, covering the greatest geographic area, and the least number on the Jacksonburg and Ontelaunee formations covering a small area. Note that the Rickenback Formation has the highest density of sinkholes per unit area (9.3 per square mile), followed by the Allentown Formation with 8.6, and the Epler Formation with 6.9. An attempt to classify the sinkholes into categories based on their origin shows the Richenback and Epler formations have the highest density of naturally occurring sinkholes, whereas the Allentown Formation has a high utility-related component. Several recent sinkhole episodes, not reported by the media, have occurred on the Allentown Formation after it was stripped during construction of a housing development . This means that human action aggravates sinkhole formation in an area that otherwise does not have a significant number of naturally occurring sinkholes. Therefore planners should pay more attention to the Rickenback, Allentown, and Epler formations because of the great likelihood of sinkholes developing in them.

Investigation of newspaper clippings of sinkhole formation and field visits to sinkhole sites add further information to the previous study that is not apparent from the table. Although the Allentown Formation has a larger number of sinkholes than most other formations, the individual sinkholes are small because of the thin overburden and the small size of the joint-controlled points of recharge. Sinkhole "eyes" are close to the surface and are

Table 1: Characteristics of Lehigh Valley Carbonate Rocks. Source: Myers and Perlow, 1986.

Formation (Age)	Thickness (m)	Formation Description and Weathering Characteristics
Jacksonburg Formation (M.Ord.)	170-460	Dark-gray shaley limestone grading downward into crystalline, high-calcium limestone. Low to moderate porosity and permeability; thin soil mantle; relatively few solution features.
Ontelaunee Formation (L. Ord.)	0-200	Medium-gray, finely crystalline dolomite; cherty at base; missing at many locations. Solution-enhanced porosity and bedrock pinnacles characteristic. Moderate to thick soil mantle.
Epler Formation (L. Ord.)	270±	Interbedded very fine grained, medium-gray limestone and gray dolomite. Solution-enhanced porosity; few bedrock pinnacles; very thick soil mantle.
Richenbach Formation (L. Ord.)	220	Gray, fine to coarse dolostones, thin bedded at top to thick bedded toward base. Solution-enhanced porosity and bedrock pinnacles characteristic; moderately thick soil mantle.
Allentown Dolomite (U. Camb.)	575	Alternating bed of light- and dark-gray weathering dolomite; stromatolites and oolites common; some orthoquartzite beds. Solution-enhanced porosity and bedrock pinnacles characteristic; soil mantle generally thin.
Leithsville Formation (Uppermost L. M. Camb.)	350	Interbedded fine- to coarse-grained dolostones and tan phyllite; few thin sandstone beds. Solution-enhanced porosity; bedrock and pinnacles common; commonly covered with thick colluvium near uplands.

Table 2: Average Sinkhole Density for various geologic formations and sinkhole types. Source: Myers and Perlow, 1986.

Formation	Total Area (mi^2)	Total No. of Sinks	Average Sinkhole Density (No./mi^2) (all occurrences)	Average Sinkhole Density (sinks/square mile) Naturally Occurring	Construction Related	Utility Related	Structure Related
Jacksonburg	24.0	054	2.2	1.8	0.3	0.1	–
Ontelaunee	06.4	028	4.2	1.4	1.4	1.4	–
Rickenbach	18.7	174	9.3	4.5	4.0	0.8	0.05
Epler	74.5	518	6.9	4.0	2.5	2.5	–
Allentown	85.0	731	8.6	2.2	2.2	4.2	–
Leithsville	32.0	069	2.1	1.0	0.2	0.9	–

easily repaired. The Leithsville Formation on the other hand, has the lowest density of sinkholes (2.1 per square mile), but it is the site of some of the most disastrous collapses in the area. The Macungie sinkhole formed on June 24, 1986 with a resulting hole that was 40 meters across and nearly 20 meters deep and cost in excess of $700,000 to repair (Dougherty & Perlow, 1988). The very large size of the sinkhole is related to the deep colluvial cover of the formation allowing the development of suffosion sinkholes. This is also the location where allogenic waters from the Hardyston Sandstone and undifferentiated gneisses of South Mountain come in contact with the limestone of the Lehigh Valley. Large, damaging sinkholes form at this location, although the overall density of sinkholes is lower in the Leithsville than in any other formation in the Lehigh Valley.

Temporal Characteristics of Subsidence

In this context, temporal characteristics refer to the months of the year when karst subsidence is most likely to occur in the Lehigh Valley. A crude surrogate value to assess this property was needed because it is difficult to get access to records that list dates of sinkhole collapse. Articles about sinkholes in the Allentown *Morning Call*, the largest-circulation newspaper in the Lehigh Valley, were counted over the five year period from 1986–1990. Eighty-nine articles have sinkhole in their title or are directly related to subsidence or sinkholes. Although many articles are not related to sinkhole subsidence specifically, they may have been prompted by a recent sinkhole collapse or a period of sinkhole-collapse. During the same period, another 22 subsidences were documented by fieldwork in Lower Macungie Township in suburban Allentown. The months in which the 111 sinkholes formed are shown on the graph in Figure 4.

There is a definite seasonal distribution of sinkhole occurrence with a peak in the summer and early fall, and a secondary peak in January and February. The summer peak corresponds to the time of the year when the groundwater is at its lowest and there is little water in the

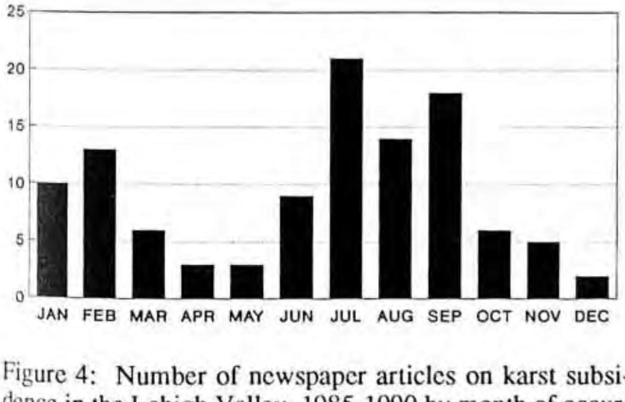

Figure 4: Number of newspaper articles on karst subsidence in the Lehigh Valley, 1985-1990 by month of occurrence. Source: *The Morning Call*, Allentown, PA.

soil or rock to help bear the weight of the overburden. This is when suffosion sinkholes are most likely to occur (Dougherty and Perlow, 1988; Miller, 1987; Buzgon, 1986). When the ground is saturated there is little chance for surficial runoff to enter joints and areas of incipient sinkholes; but, during the summer, surface runoff is directed to these points of access and this preferential flow can reactivate a sinkhole. The problem is aggravated by excessive well pumpage due to a higher seasonal demand from municipal and agricultural wells. This results in a lowering of the groundwater table and a lessening of the overburden support. There is also the possibility that excessive lawn watering can cause reactivation of some of the smaller sinks that appear in domestic yards at this time of the year. Summer is the peak construction period, and during this time large areas of earth are disturbed, resulting in new drainage patterns and the development and/or reactivation of sinkholes.

A January and February secondary maximum of sinkhole activity is more difficult to explain. This follows a period of quiescence with December being the lowest month of sinkhole activity. The ground may be frozen and there is a higher likelihood of snow than rain during December. Precipitation falling on frozen ground is more likely to end up as surface runoff than as infiltration in the incipient sinkholes. By late January and February, an early "spring thaw" occurs resulting in the liberation of large amounts of water that flow into the sinkhole eye and can reactivate it.

Although not statistically significant because of the small size of the sample, the pattern of karst collapses in the past five years indicates that the most prevalent mechanism is a type of subjacent urban karst collapse. Areas that have been developed are covered by buildings, streets, and parking lots, forming an impermeable surface. Leaking water and sewer lines may provide some of the water that flows under the urban caprock and seeks a drain. Under these conditions, loose unconsolidated sediments are washed below, leaving gaping voids under the urban landscape, and resulting in eventual collapse (Sanchez, 1988; Wittman, 1988; Clark and Reaman, 1988). Other collapses have occurred owing to indiscriminate filling of sinkholes and later development of the area (Harris, 1986; Lowry, 1987). Once the organic matter in the fill has decayed, the sinkhole may rejuvenate and cause a collapse. There are also several incidences of sinkholes repeatedly forming on the same site, such as the Macungie sinkhole (Dougherty and Perlow, 1988) and the Vera Cruz sinkhole (Darrah, 1987). Other episodes have been caused by development (Myers and Perlow, 1986), runoff from highways (Leffler, 1988), summer drawdown of the water table (Miller, 1987; Dougherty and Perlow, 1988; Wittman, 1988), faulty installation of water and sewer lines (Nixon, 1988; Sanchez, 1988), and faulty design of storm-water detention basins and surface drainage routes (Leffler, 1988). All major episodes of collapse mentioned in the news media during the five-year study period had a threshold that was most likely exceeded by human intervention.

Conclusions

This is only a preliminary analysis based on a small sample. Records from the roadmasters and highway departments in the area may eventually yield a larger sample on which to base a more detailed analysis. Spatially, the Allentown and Leithsville formations appear to be the most dangerous areas of the Lehigh Valley; the former because of a higher density of sinkhole development and the latter because of the large suffosion dolines that form in the deep colluvial cover. Temporally, there is a summer maximum of sinkhole activity and a late winter/early spring secondary peak. Further work is needed to substantiate both the spatial and temporal patterns identified above.

References

30–foot sinkhole opens in shopping center: *The Morning Call*, Allentown, PA, August 4, 1986.

Another U. Saucon road is affected by sinkhole: *The Morning Call*, Allentown, PA, March 7, 1987.

Berg, T.M. and Dodge, C.M., 1981, *Atlas of Preliminary Geologic Quadrangle Maps of Pennsylvania:* Pennsylvania Topographic and Geologic Survey, Harrisburg, PA.

Berton, V., 1987, Sinkhole threatens to undermine Northampton Borough home: *The Morning Call*, Allentown, PA, September 12, 1987.

Bieber, S., 1986, Homes evacuated as hole opens in Macungie street: *The Morning Call*, Allentown, PA, June 24, 1986.

Bonaparte, R. and Berg, R.R., 1987, The use of geosynthetics to support roadways over sinkhole prone areas, *in* Beck, B.F. and Wilson, W.L. (editors), *Karst Hydrogeology: Engineering and Environmental Applications: Proceedings of the Second Multidisciplinary Conference on Sinkholes and the Environmental Impacts of Karst, Orlando, Florida, 9-11 February 1987*: A.A. Balkema, Rotterdam and Boston, p. 437-445.

Buzgon, M., 1986, Residents flee street–gobbling Macungie sinkhole: *The Morning Call*, Allentown, PA, June 24, 1986.

Cassler, K., 1991, Gas explosion kills woman, levels Allentown row homes: *The Morning Call*, Allentown, PA, August 30, 1991.

Clark, J. and Reaman, D., 1988, City church collapses into sinkhole: *The Morning Call*, Allentown, PA, February 18, 1988.

Cowen, D., 1988, City firm awarded contract to fill sinkholes at ABE (airport): *The Morning Call*, Allentown, PA, July 27, 1988.

Darrah, T., 1986, PennDOT says its not to blame for latest sinkhole in Upper Saucon: *The Morning Call*, Allentown, PA, September 25, 1986.

Darrah, T., 1987, Upper Saucon horror: Return of the Sinkhole: *The Morning Call*, Allentown, PA, January 10, 1987.

Dougherty, P.H. and Perlow, M., Jr., 1988, The Macungie Sinkhole, Lehigh Valley, Pennsylvania: Cause and Repair: *Environmental Geology and Water Science*, v. 12, no. 2, p. 89–98.

Dougherty, P.H., 1989, Land use regulations in the Lehigh Valley: Zoning and subdivision ordinances in an environmentally sensitive karst region, *in* Beck, B.F. (editor), *Engineering and Environmental Impacts of Sinkholes and Karst: Proceedings of the Third Multidisciplinary Conference on Sinkholes and the Engineering and Environmental Impacts of Karst, St. Petersburg Beach, Florida, 2-4 October 1989*: A.A. Balkema, Rotterdam and Brookfield, p. 341-348.

Emergency work at Upper Saucon sinkhole complete: *The Morning Call*, Allentown, PA, November 4, 1986.

Harris, K., 1986, Sinkhole site once a pond, geologist reports: *The Morning Call*, Allentown, PA, July 1, 1986.

Joint Planning Commission, 1991, *Population Summary for Lehigh and Northampton Counties*: Allentown, PA.

Kochanov, W., 1987, Sinkholes and Karst Related Features of Lehigh County, Pennsylvania: *Pennsylvania Topographic and Geologic Survey*, Open-File Report.

L. Nazareth woman files lawsuit over sinkholes: *The Morning Call*, Allentown, PA, July 27, 1989.

Lash, G.G.; Lyttle, P.T.; and Epstein, J.B., 1984, Geology of an accreted terrain: The eastern Hamburg Klippe and surrounding rocks, eastern Pennsylvania: *49th Annual Field Conference of Pennsylvania Geologists*, Harrisburg, PA.

Leffler, P., 1988, Farmers say runoff floods their fields: *The Morning Call*, Allentown, PA, June 20, 1988.

Lowry, T., 1987, Upper Saucon sinkhole repair questioned: *The Morning Call*, Allentown, PA, January 16, 1987.

Miller, B.L. and others, 1942, Lehigh County, Pennsylvania: Geology and Geography: *Pennsylvania Topographic and Geologic Survey, County Report 39*, 492 p.

Miller, T., 1987, Sinkhole prone Saucon Valley worries

planners: *The Morning Call*, Allentown, PA, December 28, 1987.

Myers, P.B. and Perlow, M., Jr., 1986, Development, occurrence, and triggering mechanisms of sinkholes in the carbonate rocks of the Lehigh Valley, eastern Pennsylvania, *in* Beck, B.F. (editor), *Sinkholes: Their Geology, Engineering, and Environmental Impact: Proceedings of the First Multidisciplinary Conference on Sinkholes, Orlando, Florida, 15-17 October 1984*: A.A. Balkema, Rotterdam and Boston, p. 111-115.

Nixon, J., 1988, Broken water main caused Easton sinkhole: *The Morning Call*, Allentown, PA, December 15, 1988.

Sanchez, L., 1988, City crews to repair washout below street: *The Morning Call*, Allentown, PA, February 20, 1988.

Whelan, F., 1986, Another day in the Valley, another sinkhole: *The Morning Call*, Allentown, PA, June 26, 1986.

Wittman,B., 1988, Deep history of sinkholes in Allentown: *The Morning Call*, Allentown, PA, April 10, 1988.

Wood, C.R.; Flippo, H.N.; Lescinsky, J.B.; and Barker, J.L., 1972, Water resources of Lehigh County, Pennsylvania: *Pennsylvania Geological Survey Water Resources Report W31*, 263 p.

Youngwood, S., 1988, Muhlenberg dormitory gets that sinking feeling: *The Morning Call*, Allentown, PA, January 30, 1988.

Zoning Ordinance of Lower Macungie Township: Lower Macungie, PA: Lower Macungie Township Board of Supervisors, 1989.

Plate D (Above): Clean-up of large dump site at entrance to Stillhouse Cave, Randolph County, West Virginia, August 1990. This event was organized by the Cleveland Grotto of the National Speleological Society with the cooperation of the West Virginia Department of Natural Resources and Department of Highways. Cavers from many grottos in the Appalachian region devoted two days to the effort and removed 15 truckloads, or 22,500 pounds, of debris. The site is the first "Adopt-A-Dump" Project in West Virginia. *See* in this volume Kastning and Kastning, p. 123, and Erchul, p. 147, for discussions of clean-up projects and illegal dumping in sinkholes. *Photograph by Karen M. Kastning.*

Plate E (Left): Members of Boy Scout Troop 46 of Radford, Virginia at entrance of New River Cave, Giles County, Virginia, following a trip to clean trash from this heavily visited cave. A new cave conservation program for scouting groups has been initiated in the New River Valley of Virginia and West Virginia. Participation in a cave or karst clean-up is a requirement for the Cave Conservation Awareness Award (*see* Figure 2 of Kastning and Kastning, this volume, p. 130). *Photograph by Karen M. Kastning.*

Illegal Disposal in Sinkholes: The Threat and the Solution

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ABSTRACT

In 1988 the Virginia Military Institute conducted a study supported by the Virginia Environmental Endowment on sinkhole dumping and the risk to groundwater in Virginia's karst areas. The karst area of Virginia is located in the Valley and Ridge Province and consists of 24 counties on the western edge of the Commonwealth. Due to the karst/limestone geology that characterizes the region, these counties share a common vulnerability to their groundwater resources. Karst aquifers are among the most sensitive to disturbance and are readily contaminated. One of the most visible and obvious sources of contamination is the dumping of trash and waste into sinkholes. Botetourt and Rockbridge counties, the areas that were the focus of the research, have thousands of sinkholes within their boundaries. The possibility of contamination to local groundwater is great due to the presence of household hazardous waste in dumps. A total of 260 illegal dump sites were documented in the study area, with 75 percent of these existing in karst areas. Approximately 23 percent of the illegal dumps were in sinkholes. More than 90 percent of the population of the study area take their water from wells and springs. The research has concentrated on an effective methodology of locating and assessing potentially dangerous sinkhole-dump sites. The project has documented these techniques, allowing other localities in karst terrain to obtain similar information. A slide show and video have been produced to show local schools and citizen organizations the hazards associated with sinkhole dumping. Also, a survey cost analysis for identification, evaluation, and documentation of these illegal disposal sites has been developed.

The task of compiling a listing of all known open dumps within a regional boundary can be effectively realized using the methodology proposed in this study. In addition, the extent (numbers, nature, and sizes) of each site should be described to aid in proper remediation and elimination.

Introduction

Illegal disposal in sinkholes has become so widespread that it has resusted in one of the most comprehensive studies ever conducted in the Commonwealth of Virginia to identify, evaluate and document this threat to the environment. The study was conducted by personnel of the VMI Research Laboratory (VMIRL) in 1988 and 1989 and supported by a grant from the Virginia Environmental Endowment. A previous independent study conducted in Rockbridge County, Virginia by Slifer highlighted the problem and alluded to the scope and magnitude of such a study. Slifer's study was published by the Virginia Water Resources Research Center (Slifer, 1987) and established the guidelines and impetus to obtain more data to better define the threat to groundwater in the karst areas of Virginia caused by illegal disposal in sinkholes. Fortunately for the second study, Slifer was able to contribute as a member of the VMIRL research team. He was active in coordinating, analyzing and documenting the data collected in this study and comparing it with previous data he collected in Rockbridge County. This paper reviews the effectiveness and methodology of the Rockbridge and VMIRL studies, provides a cost analysis, and discusses the draft and promulgated regulations relating to illegal disposal in local solid-waste management plans.

The Study

As previously mentioned, a case study of illegal disposal sites in Rockbridge County, Virginia was conducted and documented by Slifer in 1986 and 1987. In 1988 a similar but more comprehensive survey commenced in neighboring Botetourt County (Slifer and Erchul, 1989; Slifer, 1990). Botetourt and Rockbridge are similar in

size, geology, and population. Both counties are primarily rural and each operates a single sanitary landfill. Whereas Rockbridge maintains a solid-waste collection system of dumpster boxes, Botetourt relies on private commercial haulers. These counties differ in their legal approach to the problem. Rockbridge lacks an ordinance that even recognizes or defines illegal disposal and has made little effort at enforcement. Botetourt has an adequate local ordinance banning illegal disposal, but enforcement efforts have been random rather than from a systematic program to identify and eliminate illegal disposal sites.

For both surveys on illegal disposal sites, information was collected by mailing a letter and county map to approximately two dozen persons who are knowledgeable about the counties. Reported sites were then field checked and evaluated. In both cases the rate of response exceeded 50 percent. This technique is inexpensive, simple, and produces reasonably good data for those sites that are the most visible and that are usually located along road sides. However, this questionnaire technique does not approximate the actual number of disposal sites in a county owing to the large number of sites located on private property and not visible from roads. Consequently, in the Botetourt study, it was decided to supplement the questionnaire data by conducting aerial investigations of the county. A small plane was flown at altitudes low enough to identify most illegal disposal sites. Two flights were made at altitudes ranging from 600 to 1,000 feet during periods when leaves were off the trees. A total flight time of approximately 10 hours was required to cover and recheck the county. A set of topographic quadrangle maps was used by a spotter to help guide the pilot to fly a series of parallel flight lines across the county, and to record locations of disposal sites as they were observed. It was found that the most efficient way to record site locations was to note the latitude and longitude coordinates from a digital LORAN-C display as the plane was directly above a site. These coordinates, when plotted on topographic maps afterward, were usually accurate to within several hundred feet. A total of 106 disposal sites were recorded during these two flights, and many of them were photographed. The aerial data was then verified by field checking each site. The total number of confirmed sites in Botetourt was 168. An update of the 1986 survey of disposal sites in Rockbridge County led to documentation of 92 sites. Nearly twice as many sites are known in Botetourt, primarily because that county was inventoried by aerial survey. A similar aerial survey for Rockbridge would probably double the number of disposal sites known there.

During the field check each site was evaluated and many were photographed. A field-check form was developed to ensure a standardized evaluation. A score was developed for each site by assigning numerical values to factors such as disposal-site size, site geology, topographic setting, public access, frequency of use, waste types, and proximity to water. The ranked scores were assigned to categories that describe the relative threat to groundwater at each site.

For both counties illegal disposal sites ranking in the most serious categories comprise approximately 29 percent of the total. These illegal sites are believed to be actual or imminent hazards in terms of groundwater contamination. An average of approximately 17 percent of the total number of illegal disposal sites fell within the low-impact category.

The number of illegal disposal sites located in the two karst areas is nearly identical - 73 percent for Botetourt and 74 percent for Rockbridge. Rockbridge has a greater share of illegal-disposal sites in sinkholes (29 percent) than does Botetourt (16 percent). This is probably due to a greater density of sinkholes in the Rockbridge area. Because groundwater is inherently at risk from illegal disposal in sinkholes, the weighing of ranking factors tends to place sinkhole-disposal sites into the more serious categories. In Rockbridge 58 percent of the illegal disposal sites have public access, compared to only 40 percent in Botetourt. This is a reflection of the survey methodologies; because the Rockbridge survey was not aerial, it tends to favor the more visible roadside sites that, by definition, have public access. An important, apparent difference between Rockbridge and Botetourt illegal disposal is the prevalence of 55-gallon drums observed in Botetourt sites. Drums are rare in Rockbridge sites but were counted in 21 percent of the Botetourt sites. Drums are of course often used to contain hazardous liquid chemicals or waste. Many of the drums are empty but some still contain unknown substances. Because personnel evaluating the Botetourt sites were not trained to sample for hazardous materials, data is not available to characterize contents of the drums. Perhaps Botetourt's relative proximity to the City of Roanoke and its numerous industrial and commercial facilities may explain the presence of drums in at least 36 sites.

Using the results of the case studies of Rockbridge and Botetourt counties, and extrapolating the data, the total number of illegal disposal sites within the Valley and Ridge Province (the karst region of Virginia) is projected to be approximately 4032. Of that total, about 2943 would be located in karst and about 927 would be in sinkholes. Based on an average of 29 percent, approximately 1169 illegal disposal sites would be in the serious-threat category.

If illegal disposal is not limited to the karst region of Virginia, the statistics projected for the entire Commonwealth could exceed 15,000 illegal disposal sites throughout the state. This alarming statistic would require an active educational program, stringent state regulations, and enforcement by localities if the number of illegal disposal sites are to be reduced and eliminated.

Draft Regulations

Effective July 1, 1989, the Virginia General Assembly directed the Department of Waste Management to prepare regulations governing the preparation of local

solid-waste management plans and to specify the procedures by which local governments must attain certain recycling goals. An advisory committee was formed and a draft document was prepared. The draft regulations were concep-tual in nature and were based on numerous perspectives. These draft regulations were to receive public comments and formal hearings and then would be revised to become the promulgated regulations for the development of solid-waste management plans by localities in the Common-wealth of Virginia.

In the draft regulations, specifically noteworthy to this study, was the requirement for all Virginia localities to provide a listing in their solid-waste management plans regarding all known open dump sites within the regional boundary, to name the current owner of the site, to describe the nature and size of the site, and to determine the extent of illegal disposal and describe actions to eliminate it. The draft version defined an "open dump" as a site on which any solid waste is placed, discharged, deposited, injected, dumped, or spilled so as to create a nuisance or so as to pose, within the determination of the Executive Director, a substantial present or potential hazard to human health or the environment, including the pollution of air, land, surface water, or groundwater. "Illegal disposal" was defined as the disposal which is contrary to applicable law or regulations.

It appeared fortuitous that the regulation requirement just stated could be effectively met by the methodology proposed in our comprehensive study conducted on illegal disposal in Botetourt County. A letter was addressed to all counties in the Commonwealth informing them of the study and that a cost analysis for the identification, evaluation, and documentation of illegal disposal sites was being prepared to allow them to understand the methodology and the associated costs. This would enable them to make necessary budgetary adjustments if they desired to implement such a study for their locality.

Survey Cost Analysis

Some generalization is required in compiling an accurate survey cost analysis for the identification, evaluation, and documentation of illegal disposal sites for a specific locality. An assumption is that the cost would be based on our research study of a western Virginia county. Realizing that flying, labor, and material costs will vary depending on location, size of locality, and extent of evaluation required, the cost analysis presents ranges of costs for each item. Various options are considered in order to provide an estimate for budgetary planning based on individual locality requirements.

A survey cost analysis is divided into three line items; identification, evaluation, and documentation. Table 1 shows the cost breakdown of each of these line items. The identification line-item is divided into maps, inquiry, aerial reconnaissance, and data recording. Most of the identification costs are fixed, one-time purchases. However, even without a survey of knowledgeable citizens, aerial reconnaissance must be conducted if effective and complete data is to be generated.

Various options are offered in the evaluation line-item of the cost analysis that could be selected depending on the extent of detailed information desired. The greater the detailed information desired - the greater the cost, and consequently the evaluation line-item will usually be the most influential factor in the total survey cost. In addition, each of these options have subsets affecting cost. For example, each option could be conducted using available personnel or contracting others to do the work. VMIRL had proposed to use VMI cadets, home on summer furlough, to conduct the evaluation survey near or in their home localities. This would provide summer employment for cadets to conduct the evaluation if county personnel were not available. Cadets would be trained in proper data-collection techniques prior to conducting their survey.

Finally, the documentation line-item provides the finalized data analysis, location of sites on maps, and other detailed information desired in a final report and/or listing on a computerized spread sheet. The cost of the documentation line-item is relatively fixed and the range of cost will only vary as a result of requiring contract support or if the work can be conducted efficiently by available personnel.

A survey for a county the size of Botetourt County (548 square miles) should cost between $4,500 to $8,400 depending on the detail of data desired. This equates to $8 per square mile using available personnel and $15 per square mile if contracted personnel are required. These cost figures should be adequate for budgetary estimates for providing data to determine the location, size, nature and extent of sites of illegal disposal and open dumping within the regional boundaries of any locality. By determining the location of sites, the owner could be ascertained and a course of action could be prescribed to eliminate the site. It is true that new illegal disposal sites could occur at any time and another survey should be planned in three to five years to determine the effects of the current action to eliminate sites discovered on the first survey and to locate any new illegal disposal sites if they exist.

Promulgated Regulations

On May 15, 1990, after numerous public meetings and formal hearings on draft regulations, the Commonwealth of Virginia Department of Waste Management promulgated regulations for the development of solid-waste management plans. These regulations made it mandatory for every city, county, and town in the Commonwealth to develop a solid-waste management plan and submit it to the state for approval. However, nothing was mentioned about requirements to provide a listing of all open dumps within the regional boundary, to determine the extent of

INDENTIFICATION:	Range of Cost
Maps (Topographic and Geologic)	\$ 75-\$ 100
Inquiry (Write letters, send questionnaires and maps, and process and plot responses)	\$ 370-\$ 820
Aerial Reconnaissance (Aircraft rental flying time: 2 trips with one make-up flight scheduled, 10 to 20 hours, film for aerial photographs)	\$1,500-\$2,200 \$ 500-\$ 690
Data Recording (Data plotting on maps and entry into computer)	\$ 570-\$ 870
Identification Total	\$ 3,015-\$4,680
EVALUATION:	
Option 1: (Check all illegal disposal sites in detail. This includes measurements, analysis of contents and photographic data)	\$ 1,500-\$2,500
Option 2: (Check only the largest or most accessible illegal disposal sites)	\$ 750-\$1,250
Option 3: (Evaluate and prioritize according to photographic data, DRASTIC maps or similar information on surface and groundwater vulnerability)	\$ 550-\$1,050
Option 1A, 2A: (Meet with county representatives and train them in field checking and the data-collection routine and allow them to check all sites)	\$ 300-\$ 600
Option 1B, 2B: (Allow VMI cadets to collect data during the summer vacation break)	\$1,000-\$1,500
Evaluation Total	\$ 550-\$2,500
DOCUMENTATION:	
Report and Data Analysis (Assemble all data and analyses. Produce report and computer spread sheets. Meet with county representatives to present results)	\$ 920-\$1,220
Documentation Total	\$ 920-\$1,220
GRAND TOTAL	\$4,485-\$8,400

Table 1: County Survey Cost Analysis for Identification, Evaluation and Documentation of Illegal Disposal Sites

illegal disposal, or to prescribe action to eliminate dumps. In fact, the only pertinent items mentioned in the final version of the regulations were the definitions of an "open dump" and "illegal disposal." The omission of any means to curtail open dumping and illegal disposal in the promulgated regulations was very surprising to this investigator in light of the data generated in the two counties that were studied. On inquiry with the Department of Waste Management, the following reasons were given as to why the illegal-disposal regulations were dropped from any waste-management plan.

a. Information and data of sites is not readily available to the localities or state and thus the time needed to generate a plan without information would require data gathering, prohibiting timely submission.

b. It would be difficult and costly to enforce regulations and determine actions to eliminate illegal disposal sites.

c. There are complex legal and financial aspects associated with clean up of all illegal disposal sites within the Commonwealth.

Conclusions

The Commonwealth of Virginia Department of Waste Management has excluded the draft-regulation clause dealing with open dumps and illegal disposal in the promulgated regulations for the development of solid-waste management plans for localities within the Old Dominion. It is this investigator's opinion that the effectiveness of other forms of waste management, such as source reduction, recycling, landfilling, and combustion, used to control the waste stream will be questionable or misleading if information on illegal disposal is not determined. The research presented in this paper on illegal disposal in sinkholes highlighted an environmental problem in the karst region of Virginia. However, it is this investigator's opinion that illegal disposal in the rest of the Commonwealth, and probably in much of the nation, exists on the same scale as in the two counties studied. Botetourt and Rockbridge counties comprise approximately 1000 square miles, and extrapolation of data from a sampling area of this size to the rest of the state is not unrealistic. Even if we conservatively estimate from our data that 10,000 sites of illegal disposal and open dumping exist in the Commonwealth, this is a considerable amount of the waste stream that is totally unregulated. In addition, from the Botetourt County study, the fact that 21 percent of the sites contained 55-gallon drums could infer that a part of this unregulated waste stream may be hazardous waste.

In summary, this research demonstrated a technically feasible and cost-effective approach and methodology to accurately determine the number, location, size, and threat to the environment of illegal disposal sites. It provided documentation in the form of a report, and video tape, and slides to inform and educate people of environmental problems associated with illegal disposal. It also provided complete data for illegal disposal sites for one locality (Botetourt County) in the Commonwealth. A need exists

to conduct this type of study in other localities. The data obtained would then be available for solid-waste management plans, and a regulated, methodical eradication of illegal-disposal sites could then progress throughout the Commonwealth.

References

Slifer, D.W., 1987, Rockbridge's illegal dumps: *Focus on Water* (Virginia Water Resources Research Center), no. 2, 8 p.

Slifer, D.W., 1990, Groundwater at risk in the common-dump of Virginia, *in* Erchul, R.A. (editor), *Proceedings of the Symposium for Virginia Localities on Waste Management and Groundwater Protection, April 3-4, 1990 at the Virginia Military Institute, Lexington, Virginia*: VMI Research Laboratories, Lexington, p. 96-101.

Slifer, D.W. and Erchul, R.A., 1989, Sinkhole dumps and the risk to ground water in Virginia's karst areas, *in* Beck, B.F. (editor), *Engineering and Environmental Impacts on Sinkholes and Karst: Proceedings of the Third Multidisciplinary Conference on Sinkholes and the Engineering and Environmental Impacts of Karst, St. Petersburg Beach, Florida, 2-4 October 1989*: A.A. Balkema, Rotterdam and Brookfield, p. 207-212.

Plate F: Environmental poster produced by the Virginia Cave Board, Department of Conservation and Recreation. This color poster, measuring 22 by 28 inches, has been distributed to teachers of Earth Science in public schools of Virginia. The poster graphically emphasizes the sensitivity of karst groundwater to contamination from dumping in sinkholes and other sources. Poster was funded by the Virginia Natural Heritage Program of the Department of Conservation and Recreation, Virginia Department of Education, and the Cave Conservancy of the Virginias. *See* Kastning and Kastning, this volume, p. 123 for discussion. *Photograph provided by Jack Jeffers of Radford University.*

Evaluating a Landfill Expansion in Karst Terrain

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ABSTRACT

A landfill expansion is being proposed in Rockingham County, Virginia to meet future waste-disposal needs. The site is underlain by Ordovician-age dolostone of the Beekmantown Group. Unlike several other states, Virginia does not expressly prohibit construction of a landfill in karst. However, the applicant must provide evidence that the landfill is not sited in a geologically unstable area in which geologic or geomorphic features may result in sudden or non-sudden events that could cause subsequent failure of liners. Also, evidence must be given to assert that a monitorable groundwater table exists beneath the site - an often difficult task in karst terrain.

This paper describes the geologic investigation and approach to risk assessment that was conducted to evaluate the site for landfill development. A number of techniques were utilized to collect subsurface data including test borings, air-track probes, air-rotary drilling, downhole geophysics, and slug tests. This data was evaluated in conjunction with geologic reconnaissance data, fracture-trace data, air photos, and sinkhole maps to assess geologic risk factors at the site.

Project-development risk factors that also could affect future sinkhole potential were considered, such as dewatering, final depth of overburden, and changes in surface-water infiltration. Based on the assessment of both geologic and developmental risk factors, it was concluded that the site was favorable for landfill development.

Introduction

Solid-waste regulations for many states lying partially within the Appalachian Valley and Ridge, such as Pennsylvania and West Virginia, prohibit the development of landfills above carbonate rock. Virginia allows consideration of such sites, although the onus is clearly on the owner to demonstrate geologic stability and monitorable groundwater conditions. The major concern with stability is the potential loss of liner integrity should a subsidence sinkhole occur. Subsidence sinkholes develop due to subsurface erosion of overburden soil into pre-existing solution voids or cavities in the rock. Because groundwater flow through fractured media such as carbonate rocks will follow preferential flow paths created by solutional enlargement along joints, fractures, and bedding planes, it is sometimes not possible to install monitoring wells at locations so that the data obtained is representative of the aquifer flow beneath the landfill.

Rockingham County currently operates a 48-acre landfill southwest of Harrisonburg, Virginia. A 40-acre expansion is planned that will meet new regulatory requirements of a doubly lined system. To limit potential negative environmental impacts of the current site, "landfill mining" (removal of existing wastes to newly lined cells) is being recommended. This is considered a desirable improvement over present conditions, because wastes might currently be in close proximity to, or in direct contact with, karst-solution features, allowing unhindered avenues of seepage into the groundwater table. Mined-landfill areas will later be developed in accordance with the double liner requirements.

Geologic Setting

The landfill site is located in the Valley and Ridge Province. Bedrock beneath the site consists of the upper dolostone member of the Beekmantown Group (Brent, 1960), that is made up of five, thickly-bedded Ordovician-age dolostone and limestone formations containing numerous hard, thin beds of chert. The Upper Dolomite member, as well as the limestone members, generally weather into deep, red, elastic silts and lean clays with chert fragments scattered throughout. The Beekmantown Group underlies the entire site, extending well beyond and trending northeasterly.

The Staunton Thrust, a major thrust fault, strikes northeastwardly and dips southeastwardly from its outcrop two miles southeast of the site (Gathright and Frishmann, 1986). Local structure is impacted by the presence of this thrust sheet, with several small northeast-trending folds paralleling the front of the thrust. At the western end of the landfill site, mapped bedrock strikes range from N 25 W to N 63 E, with dips ranging from 2 to 21 degrees to the west. At the eastern end of the site, strikes range from N 20 W to N 70 E with dips ranging from 5 degrees east to 10 degrees to the west. These dips are influenced by the small folds, the Middlebrook Anticline to the southeast, and the Long Glade Syncline to the northwest.

A fracture-trace analysis was used to help define potential monitoring-well sites and identify lineaments along which solutional enlargement might be more prominent. Approximately 33 fracture traces were identified on aerial photos in and around the landfill site. The strongest of the fracture trends are oriented N 0-10 E and N 50-90 W. The first trend parallels the axis of a small north-trending anticline located midway between the Middlebrook Anticline and the Long Glade Syncline. The second trend is normal to the regional strike of the Beekmantown Group.

Approach to Risk Assessment

In order to define the level of risk to the site, various geologic risk factors were identified including sinkhole frequency, fracture trends, and groundwater conditions. The geologic risk factors were defined through a three-step process which included: (1) a preliminary site evaluation, (2) site reconnaissance, and (3) a comprehensive subsurface exploration program. To assess the overall risks to the site, the geologic factors were considered together with project-development risk factors including final overburden thickness following waste-cell excavation, changes in surface-water infiltration, and groundwater pumping (DeStephen and Wargo, 1990). This evaluation resulted in an opinion as to whether the site was either favorable or unfavorable for landfill development.

Preliminary Site Evaluation

The first step in the preliminary site evaluation was identifying the potential for sinkhole development within limestones and dolostones of the Beekmantown Group. A comparison of sinkholes on a regional basis (Hubbard, 1983) was made with respect to the Beekmantown Group as shown on Figure 1. Hubbard identified sinkholes using stereo examination of low-altitude aerial photography, and only sinkholes about 30 ft in diameter or greater are shown. From this data it was concluded that on a regional basis, sinkhole occurrence is not significant within the Beekmantown Group. A higher sinkhole frequency in the Beekmantown Group around Harrisonburg is attributable to the Upper Limestone Member, which predominates in that area.

The region surrounding the site was further investigated for sinkhole occurrence by identifying closed-end depressions from topographic maps and stereographic pairs of aerial photographs. Probable sinkholes were located, and their proximity to the site are indicated on Figure 2. Based on this analysis, a low sinkhole frequency of about 3.2 sinks/square mile was identified for the Beekmantown Group limestones and dolomites. The proposed landfill disposal area is only 0.034 square miles. From this analysis, the probability for future sinkhole occurrence at the site is considered very low.

Site Reconnaissance

A site reconnaissance was conducted and included a thorough observation of surface features at the site. Interviews were also conducted with landfill personnel to develop a history of the site. Site reconnaissance was aimed at identification of subtle features such as wide bowl-like depressions, small swales, changes in vegetation and similar features that might be a result of karstification. No indications of sinkholes were observed within the landfill-expansion area, although a sinkhole is present adjacent to the existing landfill.

Subsurficial Exploration Program

Several subsurficial exploration techniques were employed to obtain information on the overburden soils, character of the underlying rock, groundwater occurrence, and karst features. These techniques included 15 test borings, eight air-track probes, continuous rock coring at one deep monitoring-well location, air-rotary drilling at three monitoring-well locations, borehole-geophysical logging, and geologic logging of a 55-foot-deep borrow trench.

Of the five air-track probes drilled into rock, none showed evidence of voids, that is, lack of dust expelled from the hole or sudden drop in the drill stem. All air-track probes in the rock showed a fairly steady drill rate, with a continuous thick dust with rock chips, indicative of continuous rock.

Within the landfill-expansion area, overburden thicknesses were substantial, varying from 20 to 60+ ft as indicated on Figure 3. General trends in the depth to rock could be estimated, although several rock pinnacles had already been encountered by earth moving equipment during shallow excavation of borrow soils.

Geophysical logging was performed in three of the four wells drilled on site to correlate geological strata encountered in the boreholes and to identify possible water-bearing fractures in order to place the well screens. A suite of four logs consisting of natural gamma, density, resistivity, and caliper, was run in each of the three boreholes. No significant fracture zones were identified on the geophysical logs. The data was used to select well-screen intervals, which ranged from depths of 70 to 140 ft.

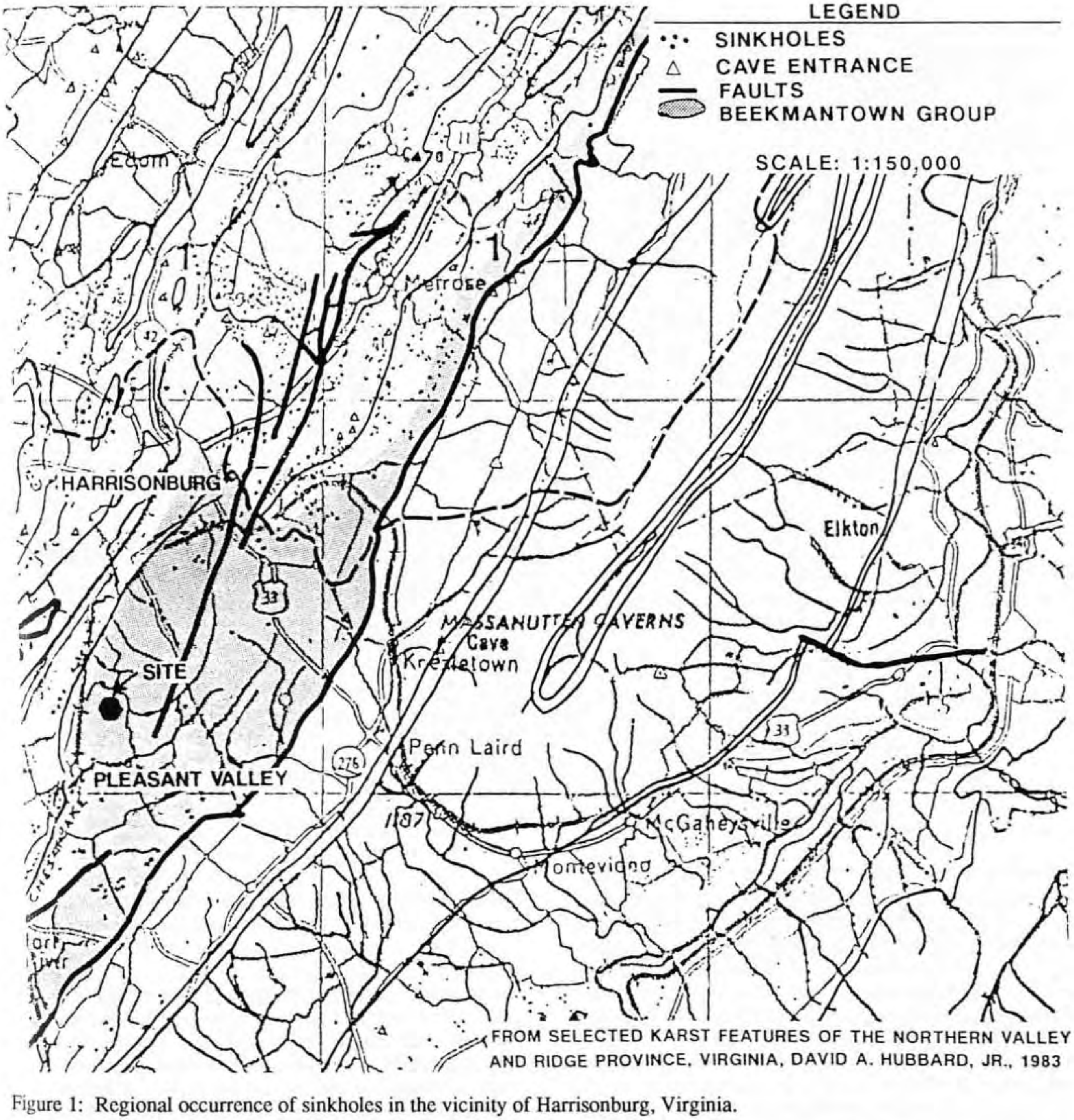

Figure 1: Regional occurrence of sinkholes in the vicinity of Harrisonburg, Virginia.

Groundwater was not encountered within the soil overburden at the site. All of the monitoring wells drilled in rock produced water, but not immediately after drilling. Water levels rose above the screened depths, indicative of a confined aquifer. From groundwater-elevation data, flow is generally westerly at a steep hydraulic gradient of 4%. Low transmissivities of 45 to 200 gpd/ft calculated from falling-head *in situ* permeability (slug) tests in each well suggested a lack of solution features. All three downgradient wells were equally spaced across the site and equidistant from the upgradient well. The hydraulic heads in these wells were very similar, indicating that there were no major interconnected karst solutional openings that might give rise to unpredictable groundwater levels over short distances. This implied that groundwater flow beneath the site could be monitored, with water-quality results representative of groundwater quality beneath the landfill.

Developmental Risk Factors

Developmental risk factors were also considered in the evaluation of the site. Development of a site can either be

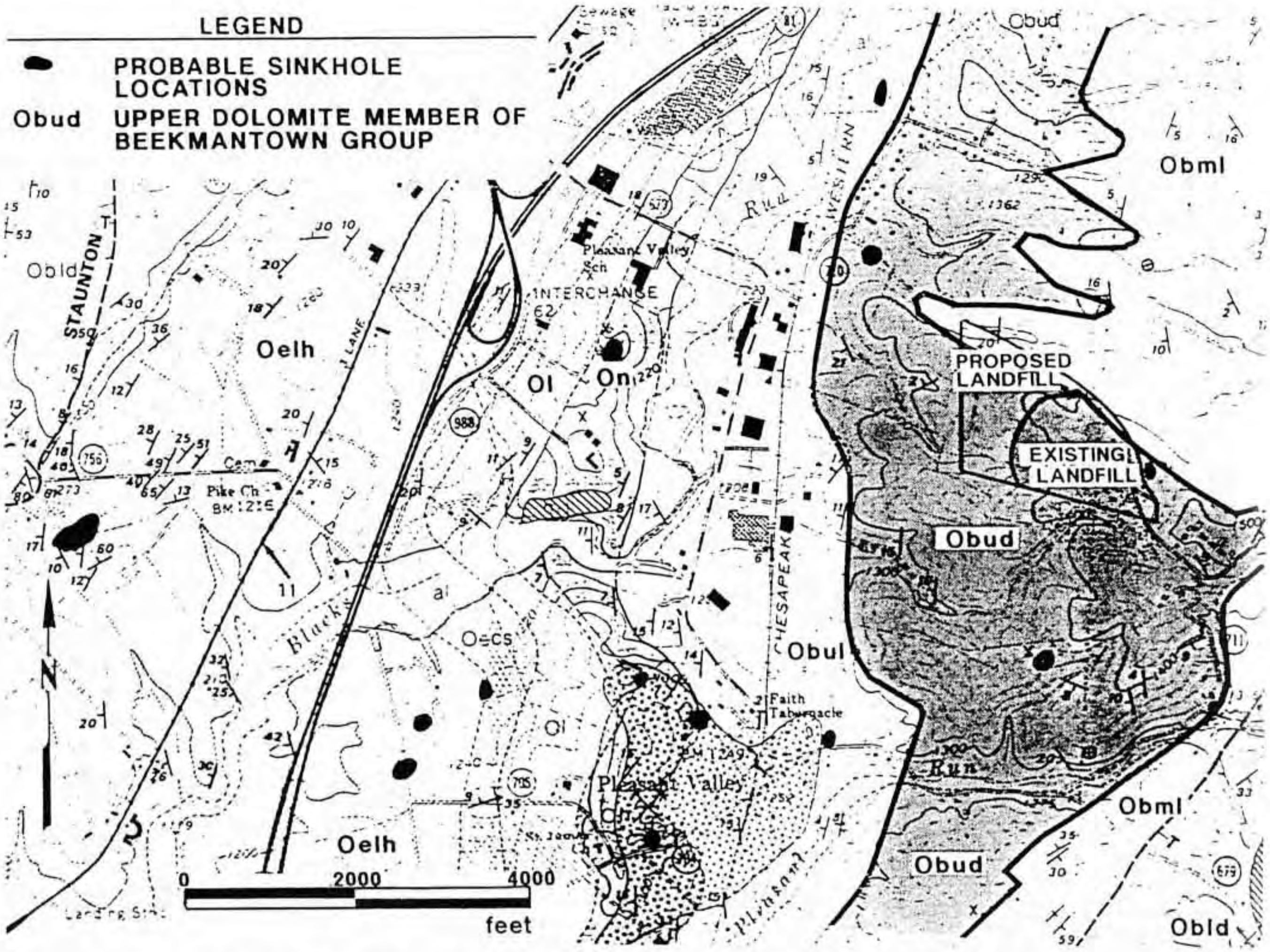

Figure 2: Probable sinkhole locations in vicinity of proposed landfill expansion.

a benefit or a detriment with regard to sinkhole potential. The majority of induced sinkholes are subsidence sinkholes, formed by subsurface erosion into pre-existing voids, induced by allowing increased infiltration of surface water, or by intense pumping of groundwater. The proposed landfill development was not considered to be detrimental with regard to sinkhole development for three reasons: (1) the double-liner system would reduce (eliminate) surface infiltration, a major triggering mechanism in the inducement of subsidence sinkholes, (2) the maximum 15-foot cuts necessary for development would leave a considerable overburden thickness above the rock, and (3) future groundwater pumping adjacent to the landfill will be limited.

Where overburden soils are thick, the potential for soil raveling into any solutional voids is lessened (Newton, 1987). Surface failure usually does not occur, even under wetted conditions, unless overburden thickness is less than about 7 feet (Williams and Vineyard, 1976). Typical overburden thickness remaining at the site following landfill construction will vary from 10 to 50 feet, with an average depth greater than 25 feet.

Subsidence sinkholes can also be triggered by intense groundwater pumping, most notably where the groundwater occurs in the soil overburden. Groundwater pumping can result in a downward migration of overburden soil by way of the following mechanisms: (1) loss of buoyant support to residual soil arching above rock openings, (2) increase in velocity and seepage forces associated with groundwater movement, and (3) movement of water to bedrock openings. However, groundwater conditions at the site are favorable with regard to water-table lowering, as the occurrence of groundwater is below the rock surface. Based on estimated drawdowns calculated from the rock permeability data, singular domestic deep wells with pumping rates less than 10 gpm were not considered to impact the site. Any such wells will be at least 500 feet from the landfill as required by regulations.

Conclusions

All of the geologic risk factors evaluated at the site supported a very low potential for future sinkhole activity in the landfill development area. These include the following:

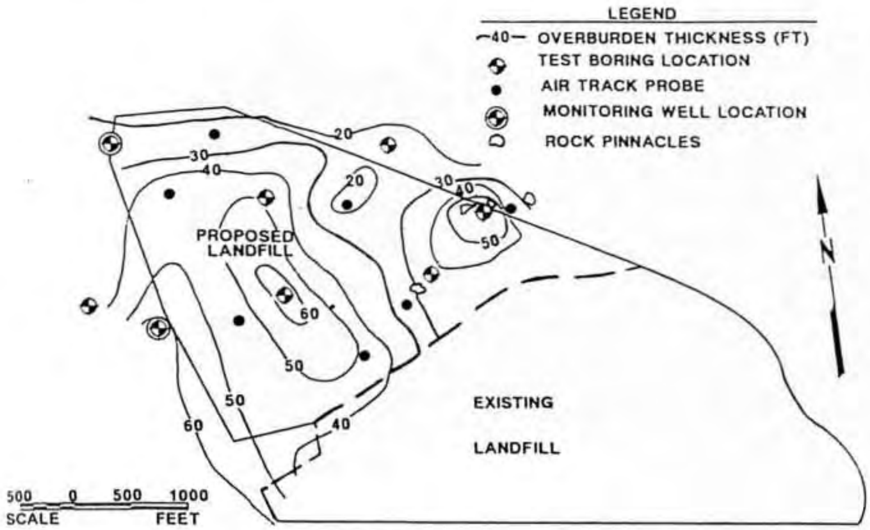

Figure 3: Thickness of overburden at proposed landfill-expansion site prior to construction.

- No sinkholes were observed within the landfill development area based on site reconnaissance, and review of topographic maps and aerial photos.
- The Upper Dolomite member of the Beekmantown Group is not particularly sinkhole prone, with an estimated sinkhole frequency of only about 3.2 sinks/square mile in the vicinity of the site.
- There are no major faults or other structural features occurring on site that would increase the likelihood of solution features.
- Fracture-trace analysis indicated no major lineaments except surficial drainageways at the site.
- No cavities or voids were indicated in any of the test borings or air-track probes that were drilled on site.
- No voids or mudseams were identified in the 78 linear feet of rock coring performed. Likewise, no loss of drilling water was observed, as might be indicative of highly fractured rock or solutional voids.
- No solutional openings or major fracture zones were indicated in over 450 linear feet of geophysical logging at the well locations.
- Transmissivities for the uppermost aquifer, as calculated for each of the wells, were very low at 45 to 200 gpd/ft.

In addition, project-development risk factors related to the landfill construction were considered to be non-detrimental with regard to sinkhole potential. Positive aspects of the design included (1) average final overburden thicknesses in excess of 25 feet, (2) use of a double-liner system to eliminate surface-water infiltration, and (3) favorable groundwater conditions together with regulatory limitations on future groundwater pumping.

Based on a combination of positive geologic and developmental risk factors, the overall sinkhole potential at the site was estimated to be low. The site was considered favorable for landfill design, and a Virginia Department of Waste Management Part A landfill permit was applied for and approved.

References

Brent, W.B., 1960, Geology and mineral resources of Rockingham County: *Virginia Division of Mineral Resources Bulletin 76*, 174 p.

DeStephen, R.A. and Wargo, R.H., 1990, Risk assessment for foundation design in karst terrain: *Annual Meeting of the Association of Engineering Geologists, Pittsburgh, Pennsylvania, October 1990.*

Gathright, T.M., II and Frischmann, P.S., 1986, Geology of the Harrisonburg and Bridgewater Quadrangles, Virginia: *Virginia Division of Mineral Resources Publication 60.*

Hubbard, D.A., Jr., 1983, Selected Karst Features of the Northern Valley and Ridge Province, Virginia: *Virginia Division of Mineral Resources Publication 44*, one sheet.

Newton, J.G., 1987, Development of sinkholes resulting from man's activities in the Eastern United States, *U.S. Geological Survey Circular 968*, 54 p.

Williams, J.H. and Vineyard, J.D., 1976, Geologic indicators of catastrophic collapse in karst terrain in Missouri: *National Academy of Science Transportation Research Record 612*, p. 31-37.

Plate G: "Lake Fairlawn," Pulaski County, Virginia, early 1980s. Infilling and paving of a sinkhole during construction of commercial establishments resulted in blockage of natural drainage and caused periodic flooding following storms. This problem resulted in the installation of a drainage system at considerable expense. Pulaski County has since established guidelines for new construction in karstic terrane. Bus is traveling of U.S Route 11. *See* Mills and others, this volume, p. 159 for a discussion of sinkhole flooding. *Photograph by Stephen D. Hale.*

Plate H: Sinkhole collapse within town of Austinville, Wythe County, Virginia, October 1989. Although sudden collapses such as this are infrequent, they may pose a problem for local residents. Note the porch to a home in the left foreground and the clothesline hanging into the sinkhole. A snow fence was erected to keep people back from the edge of the sinkhole. *See* Dougherty, p. 139, and Beck, p. 231, this volume for discussions on sinkhole collapse. *Photograph by Ernst H. Kastning.*

Predicting Sinkhole Flooding in Cookeville, Tennessee, Using SWMM and GIS

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ABSTRACT

More than 160 sinkholes occur within the city limits of Cookeville, Tennessee, many of which are subject to flooding. To aid in management of future development, the Stormwater Management Model (SWMM) and the ARC/INFO Geographic Information System (GIS) were used to predict the degree of sinkhole flooding for specific rainfall events. First, all sinkholes and their tributary watersheds were delineated on special 1:2400-scale maps with 5-foot contour intervals. Next, the following GIS coverages were prepared: topography, sinkholes, watersheds (of individual sinkholes), and land use (as determined from a 1987 city zoning map). The TIN module of ARC/INFO was used to generate an additional coverage, a slope map. For each watershed, land-use coverage was used to estimate the average impervious area and slope-map coverage was used to determine the average slope. Information provided by these coverages was entered into the SWMM and surface runoff then simulated for the following events: 50-year and 100-year, 3-hour events (3.0 and 3.9 inches, respectively) under dry and saturated antecedent-moisture conditions.

The TIN module was used to calculate the volume of each closed-contour interval for each sinkhole and these volumes were then compared to the volume of runoff predicted by the SWMM. The GIS was programmed to shade the areas between closed contours within each sinkhole according to the incremental volume flooded. Results showed that for the 100-year storm under saturated conditions, 27 sinkholes would overflow and an additional 33 would have their uppermost closed contour interval 50-100% filled. The graphical output allows planners to rapidly determine the effect of proposed development on sinkhole flooding in a watershed.

Introduction

Flooding in small drainage basins poses a problem in densely populated areas. The problem is of particular concern where sinkholes are abundant, for much of the runoff may be retained in the basins for hours or days rather than moving rapidly downstream as it does in nonkarst areas. Urbanization compounds the problem in several ways. First, the increased impervious area increases the proportion of rainfall that runs off, filling sinkholes more rapidly. Second, after more desirable building sites are depleted, construction of homes and businesses often shifts to less desirable lower areas, in and near sinkholes. Third, partial filling of sinkholes often occurs in connection with this construction. This decreases the storage capacity of sinkholes that serve as retention basins, and although it may prevent flooding of structures built on the fill, it results in more rapid rises of water in the remaining part of the sinkhole. Sinkhole capacity may even be exceeded so that water spills over and floods other sinkholes and areas not previously subject to flooding.

For the planning of development it is highly desirable to anticipate these effects. Blanket ordinances prohibiting building in or near sinkholes, or any filling of sinkholes, may unnecessarily stifle development. City officials are likely to come under strong pressure to grant exceptions, some of which may be justified. A more discriminating and probably effective approach can be provided by the ability to predict the specific effects of particular proposed development projects. Demonstrating in a quantitative manner the deleterious effects that a proposed project is likely to have can provide a more convincing case against undesirable development than a blanket ordinance. This paper describes a computer-based method of predicting sinkhole flooding and of evaluating the potential for flooding for any proposed development. This study was part of a project that also involved the evaluation of water quality

and the delineation of subsurface flow paths within the city (George and others, 1990; Pride and others, 1988).

Previous Work

White and Reich (1970) found that in drainage basins ranging from 2 to 200 square miles in area, the mean annual flood is much smaller in basins underlain by carbonate bedrock than in basins underlain by clastic rock, presumably because in the former basins much of the storm water enters groundwater storage and then resurges slowly over the next several days. Whether this result applies to basins on the order of a square mile or less is not certain. Betson (1976) studied runoff yield from four urbanized basins in east Tennessee ranging in area from 0.24 to 1.60 square miles, two of which were underlain by insoluble or moderately soluble rock and two by highly soluble rock. Urbanization produced a much greater increase in runoff yield in the latter two than in the former two. Apparently under rural conditions much of the runoff in basins underlain by highly soluble rock is lost to the subsurface. Therefore, as the land surface becomes increasingly impervious with increasing urbanization, the relative increase in runoff in these basins is much greater than in those underlain by less soluble rock.

Kemmerly (1981) discussed problems of sinkhole flooding and suggested a four-step approach for dealing with this hazard. First, all sinkholes and areas likely to be affected by sinkhole flooding should be determined from maps and field inspection. Second, aerial and ground-based photography should be taken during rainy periods to help determine the volume of runoff collected in sinkholes. This should be supplemented with staff-gauge, water-level-recorder, and precipitation data. Sinkhole floodstage, storage capacity, and subsurface drainage data can then be evaluated using appropriate hydrologic models.

The third step is calculating the volume of runoff that is supplied to a sinkhole by its catchment basin during a 100-yr intensity rainfall. The elevation to which the sinkhole is filled by this volume is then calculated and shown on the map as the "sinkhole floodplain" (100-yr flood-line elevation) analogous to the 100-yr floodplain along streams used for flood insurance purposes. Fourth, recommendations are made by local planning commissions to regulate land use within sinkholes that constitute flooding hazards.

Faulkerson and others (1981) and Mills and others (1982) carried out parts of steps 1 and 2 for the Cookeville area. Forty-two topographic maps (1:2400 scale with a contour interval of 5 feet) were used for the study. They were based on aerial photographs of the city taken in January, 1972. All sinkholes within this area, together with their drainage basins, were delineated on the topographic maps. The morphometry of sinkholes and sinkhole catchments in the Cookeville area has been studied by Mills and Starnes (1983). Extensive field observations were also made of hydrologic conditions and sinkhole flooding. Evidence for flooding, such as the presence of dried mud or flood debris on trees in the sinkholes, was found in 33.9% of the sinkholes. Such evidence tells little about the frequency of flooding but provided a first approximation of flooding hazard.

The only extensive effort to collect sinkhole water-level data in an urban karst area appears to be that by Crawford (1982) and Crawford and Groves (1984) for Bowling Green, Kentucky. Little information of this sort has been obtained in Cookeville, although extensive groundwater tracing has been conducted by Faulkerson and others (1981) and Hannah and others (1989).

Physical Setting

The study area of about 35 square miles lies in parts of the four U.S. Geological Survey 7.5-minute quadrangles shown in Figure 1. The area is located approximately in the center of the eastern Highland Rim physiographic province between the Cumberland Plateau to the east and the Central Basin to the west. Outliers of the western Cumberland Escarpment occur in the eastern part of the study area in the form of mesas capped by sandstones of the Hartselle Formation. These mesas provide the highest elevations, about 1430 ft. The lowest point in the study area is 930 ft, so the maximum relief is about 500 ft. Excluding the mesas, the relief is only about 160 ft, with elevations over much of the area averaging about 1100 ft. All but 4% of sinkholes occur on either the St. Louis Limestone or the Warsaw Formation. Both formations contain large quantities of insoluble material, and as a result, many meters of residuum commonly mantle the surface. At Cookeville the mean annual temperature is about 57 F and the mean annual total precipitation is 56 in for the period 1931 to 1985 (NOAA, 1986; National Weather Service, 1988).

With regard to hydrologic setting, Cookeville is situated on a drainage divide between Roaring River and Caney Fork River, both tributaries of the Cumberland River. As a result, only small streams occur in the study area (Figure 2). Most streams are only a kilometer or so long and terminate in sinkholes. Approximately 7 miles of cave passage have been surveyed within the city limits.

Methods

Geographic Information System

Large-scale maps delineating sinkholes and their catchment areas prepared by Faulkerson and others (1981) (*e.g.*, Figure 3) were digitized on the ARC/INFO Geographic Information System. ARC/INFO consists of two software packages linked together. ARC is the software that handles coordinates and topology, whereas INFO is the software that handles non-spatial attributes of the map. When

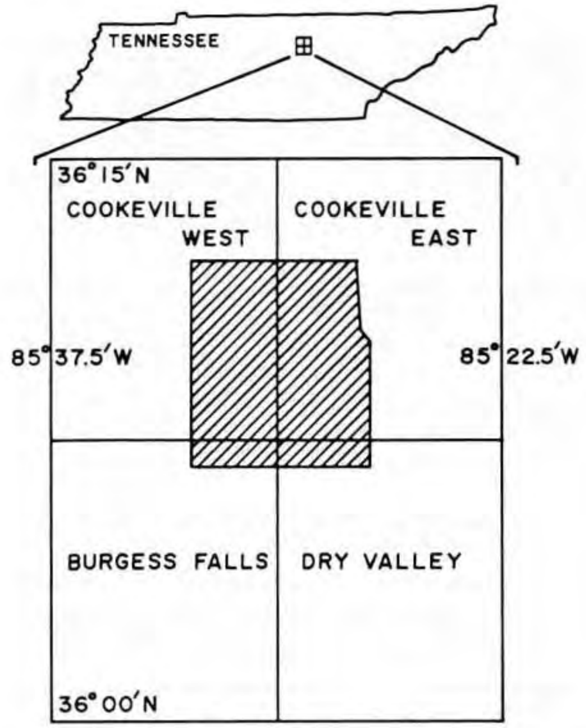

Figure 1: Index map of study area. Shaded area is covered by 42 1:2400-scale topographic maps used for this study.

a map is digitized a number of files are created. This set of files is called a coverage. The following coverages were prepared: sinkhole topography (represented by 5-ft contour lines), watershed topography (represented by 25-ft contour lines), watershed boundaries for each sinkhole, and land use (as determined from a 1987 city zoning map). The TIN module of ARC/INFO was used to generate an additional coverage, a slope map.

The basic approach in this study was to use a runoff model to calculate runoff generated in sinkhole catchments by storm events of particular intensities. Using ARC/INFO, the predicted volume of runoff was then compared with the volume of the sinkhole at various contours and the area of the sinkhole inundated was then computed. For a first approximation, sinkhole floors were assumed to be impervious. The GIS was programmed to shade the area between closed contours within each sinkhole according to the incremental volume flooded.

Stormwater Hydrology Modeling

The Stormwater Management Model (SWMM), developed by the U.S. Environmental Protection Agency (Huber and others, 1987), was used to predict stormwater-runoff quantity. A model of this sophistication was not really needed for the hydrologic simulations carried out herein, but SWMM also has the ability to model water quality, desirable for future research.

Simulations of single-rainfall events by SWMM were used to determine the potential for flooding in each of the

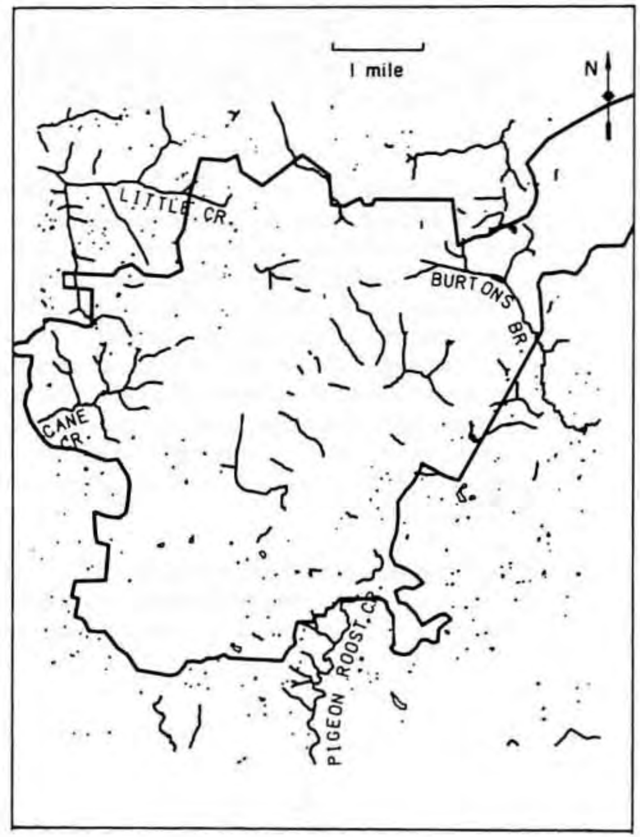

Figure 2: Map of streams and ponds in Cookeville area. Heavy line shows Cookeville city limits.

drainage basins within the study area. The storm duration to use for modeling is a question of some importance. In drainage basins a fraction of a square mile in area, the peak discharge is produced by rainfall durations of only a few minutes. Where sinkhole flooding is the main concern, however, obviously a much longer duration is called for. For a completely impervious sinkhole, the appropriate duration would be a day or longer. Because most sinkholes do drain to a degree, however, some duration between these two extremes seems most appropriate. As a reasonable compromise, a duration of three hours was selected, roughly corresponding to the duration of winter storms in the study area. The 100-year, 3-hour storm has been accepted by the U.S. Department of Housing and Urban Development as the determining storm for sinkhole floodplains in Bowling Green, Kentucky (Crawford and Groves, 1984).

In the present study, simulations were made for both 50- and 100-year, 3-hour storms. As no National Weather Service hourly precipitation data base exists for Cookeville, intensity-frequency duration curves for Nashville, Knoxville, and Chattanooga were used. These may underestimate the intensity by a small amount, as the mean annual precipitation of Cookeville is several inches higher than the other cities. The precipitation intensity is about 1.0 in/hr (3.0 in total) for the 50-yr, 3-hr storm and about 1.3 in/hr (3.9 in total) for the 100-yr, 3-hr storm. Each simulation was performed under two different soil-moisture

conditions: 1) saturated conditions prior to the rainfall event (zero initial moisture deficit), and 2) very dry conditions (initial moisture deficit of 0.32 volume-of-air to volume-of-voids).

The hydrologic input parameters to the SWMM that were needed to characterize the sinkhole watersheds were divided into climatic and morphologic types. Concerning the former, the duration and frequency of the storm event were used to develop a hyetograph. The time step used for the hyetograph was five minutes. The resulting hyetographs for the 50- and 100-year storm intensities were uniform block shapes and not representative of a natural storm. They were, however, appropriate for planning the volumes of runoff water to be transported through each watershed, which was the output datum of chief concern for sinkhole flooding.

Morphologic hydrologic parameters included watershed area, average watershed slope, average percentage of impervious area, roughness factors, depression storage, average infiltration capacity, drainage network, and width of overland flow. The watershed area was obtained from the GIS watershed coverage and the average slope was obtained from the slope-map coverage. The average percentage of impervious area was obtained from Cookeville zoning maps by the following procedure. First, zoning maps were superimposed on aerial photographs of the city. Then, a grid was placed over representative areas on the photographs corresponding to each zoning category and the percentage of grid squares covering paved areas determined. This percentage was then used for other city areas falling into the same zoning category. For a given watershed, then, the percentage impervious area could be computed based on the types and proportions of zones in that watershed. These computations were performed by the GIS.

The roughness coefficients used in the model were 0.014 for the impervious areas and 0.30 for pervious surfaces. These values are standard Manning's roughness coefficients for asphalt and short turf (Huber and others, 1987). Depression storage values used in the model

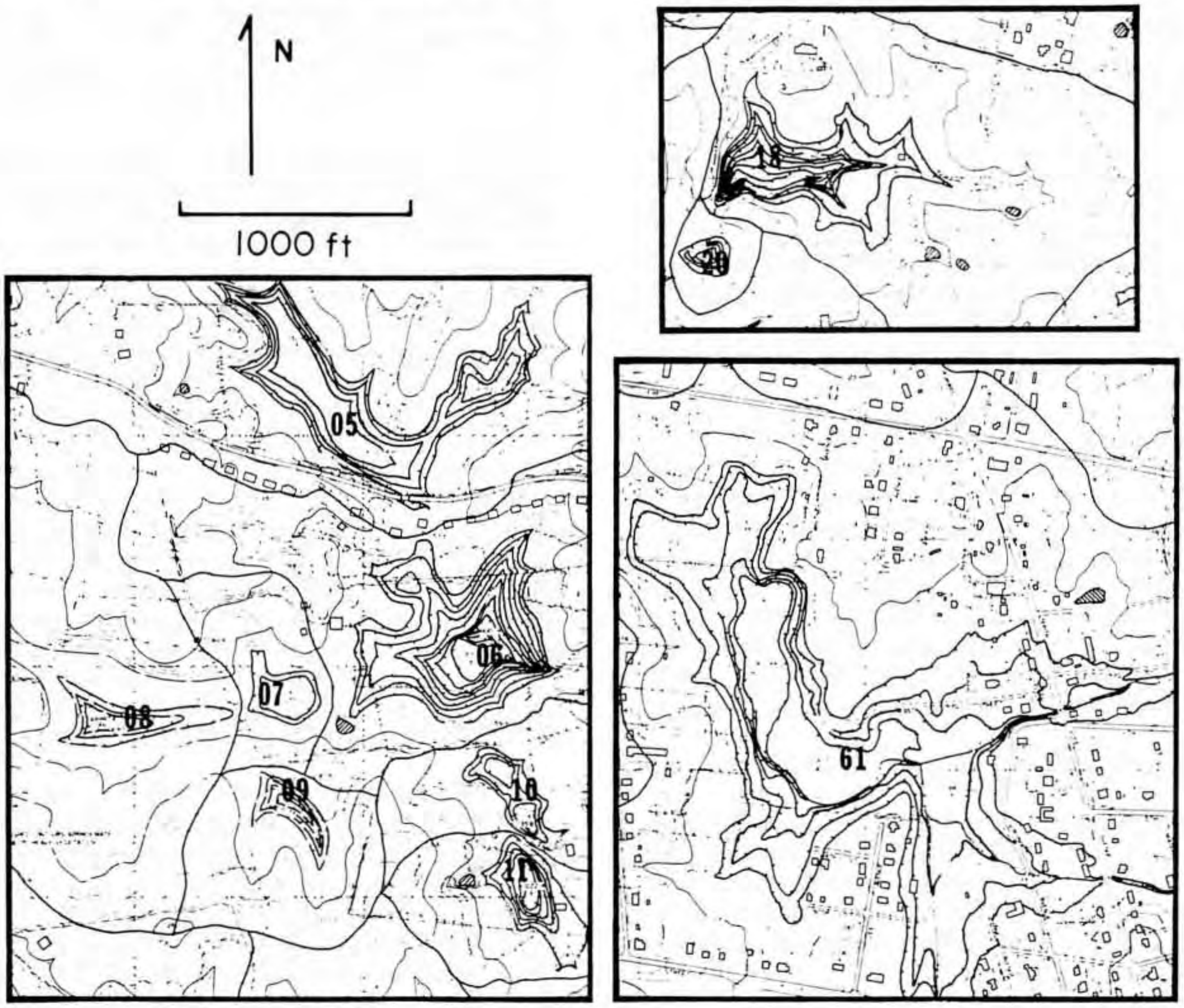

Figure 3: Example of large-scale maps delineating sinkholes and catchment areas. Contour interval is 5 feet. Dark hachured lines show closed depressions (sinkholes), whereas dark lines without hachures show catchment divides. Numbers identify individual sinkholes.

simulations were 0.011 in for impervious surfaces and 0.1 in for pervious surfaces. The average infiltration capacity was determined from the Green-Ampt equation. Values used for particular parameters of this equation were: capillary suction, 12.0 in of water; saturated hydraulic conductivity, 0.3 in/hr; and an initial moisture deficit of 0.0 volume of air/volume of voids (indicative of saturated conditions).

The drainage network was not simulated. The width of overland flow normally is assumed to be twice the length of the main drainage channel. However, during calibration of the SWMM, it was found that this value produced excessively large flow rates. By experimentation it was found that the best prediction of the hydrograph was obtained when the width of overland flow was set equal to the length of that part of the top-most closed contour of the sinkhole that faces the major part of the sinkhole drainage basin.

Calibration of the SWMM required measuring stormwater runoff from runoff events. Four sites were selected for this calibration. They were representative of watersheds in the study area and had well-defined, channelized flow entering the sinkhole. This allowed adequate flow monitoring. A simple staff gauge was used to measure the rise and fall of water in the channel over time. The water level was recorded at five-minute intervals and a velocity meter was used to determine stream discharge at various times during the rainfall event. These measurements, as well as Manning's equation for open-channel flow, were used to develop a flow-rating curve for each location so that hydrographs could be constructed.

Hyetographs for each of the four catchments were obtained by placing rain gauges within each basin and observing them during rainfall events. Rainfall intensities were grossly estimated for five-minute intervals.

A simulated hydrograph was generated by entering the hyetograph into the SWMM. This hydrograph was superimposed on the measured hydrographs and the differences in volumes and peaks of the two hydrographs compared. The width of overland flow was adjusted to calibrate the predicted hydrograph to the measured hydrograph. Having calibrated the SWMM, additional runoff events were simulated using the calibrated model. The degree of match of the predicted and measured hydrographs for each of these events was then evaluated in order to verify the calibration.

Each sinkhole watershed was characterized in the manner described previously and the data entered into a computer file for the particular sinkhole. The overland-flow width for each watershed was entered as the length of the top-most closed contour of the sinkhole. The SWMM was altered to produce an output file, generated by a specific hyetograph, that contained the total volume of surface runoff within each watershed. This output file was then used as an input file for the GIS, that mathematically compared the volume of runoff for a watershed with the volume of the sinkhole. The GIS was then used to graphically display the area inundated by the rainfall input.

Results

Four sinkhole drainage basins were used to calibrate the runoff model. As an example of these results, Figure 4 shows the hyetograph/hydrograph calibration results for sinkhole 62. The measured hydrograph for August 4, 1987, was the result of a short, intense summer storm. After adjusting the length of the overland flow, the predicted-runoff volume was 2% less than the calculated volume. Using this calibration, additional simulations were then run for other observed rainfall events at this and other sinkholes. Table 1 shows the difference between measured-runoff volumes and volumes predicted by the SWMM. Although some simulations were off by as much as 58%, the simulation at least gives the correct order of magnitude for runoff volumes, which is sufficient for the present study. Part of the error may be caused by difference between the actual impervious area within some basins and that predicted by the zoning map.

The SWMM was used to simulate runoff in all 164 sinkhole basins for both 50- and 100-year, 3-hr storms under both dry and wet antecedent-moisture conditions. Only the extremes of these four simulations are discussed here: the 50-yr storm with dry-antecedent conditions and the 100-yr storm with wet-antecedent conditions.

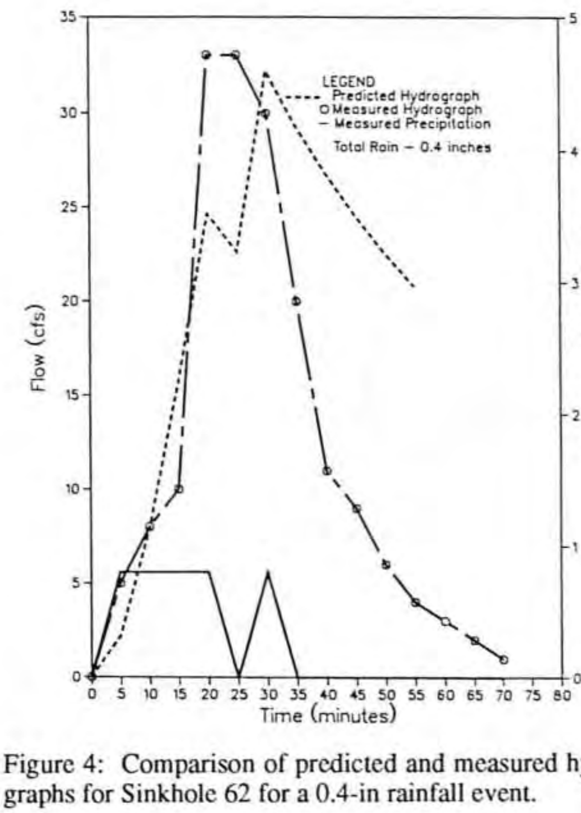

Figure 4: Comparison of predicted and measured hydrographs for Sinkhole 62 for a 0.4-in rainfall event.

TABLE 1: Runoff Calibration Results

Sinkhole number	Date	Volume difference between measured and predicted (%)
62	08/04/87	+ 2.0
62	08/05/87	- 34.0
62	01/20/88	+ 58.0
71	09/29/87	+ 16.0
36	09/29/87	+ 0.5
88	10/10/87	- 7.0

Note: Positive sign (+) indicates an underprediction of total runoff volume. Negative sign (-) indicates an overprediction.

Figures 5 and 6 show the levels of sinkhole inundation that were produced by these simulations for two small areas of Cookeville. Figure 5 shows a group of relatively small sinkholes with highly developed watersheds. As can be seen, all but one of these is filled beyond capacity by the 50-yr, dry-antecedent-condition event, not to mention the 100-yr, wet-antecedent-condition event (not shown). Several of these sinkholes, in fact, are known to flood to substantial levels even during storm events with recurrence intervals of only a few years. The effect of flood waters cascading from basin to basin was not evaluated, but this is obviously a phenomenon to consider. Sinkhole 62, for example, would spill into sinkhole 63, and 63 into 64. Sinkhole 65 also flows to 64. Table 2 shows the volume of runoff predicted to overflow these sinkholes during both events of both magnitudes.

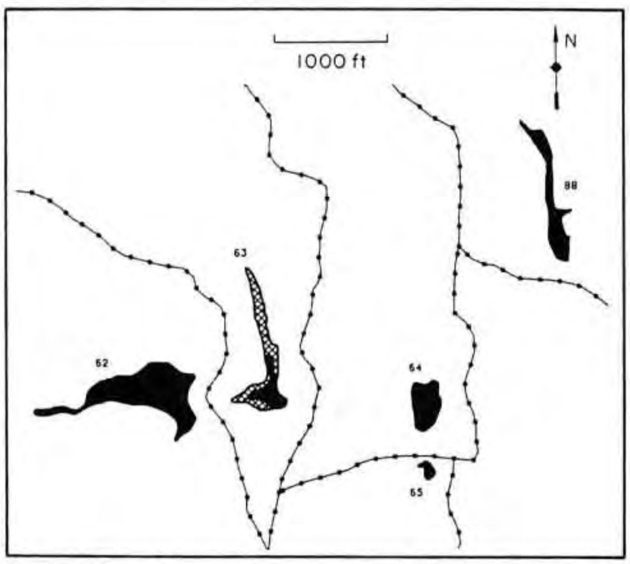

Figure 5: Example of predicted flooding in sinkholes in a commercial area resulting from 50-yr 3-hr rainfall with dry antecedent conditions. Solid shade indicates complete inundation to that level and crosshatch pattern indicates 50-100% inundation of that contour interval. A 50-yr 3-hr rainfall with wet-antecedent conditions results in overflow of all sinkholes shown here. Lines with dots show catchment divides. Numbers identify particular sinkholes.

TABLE 2. Predicted Sinkhole Overflow

Sinkhole number	Predicted overflow from 50-yr event, dry conditions (cubic feet)	Predicted overflow from 100-yr event, wet conditions (cubic feet)
62	368,100	2,126,100
63	*	490,900
64	131,300	483,600
65	76,200	380,300
88	833,500	4,170,800

* Not filled to capacity.

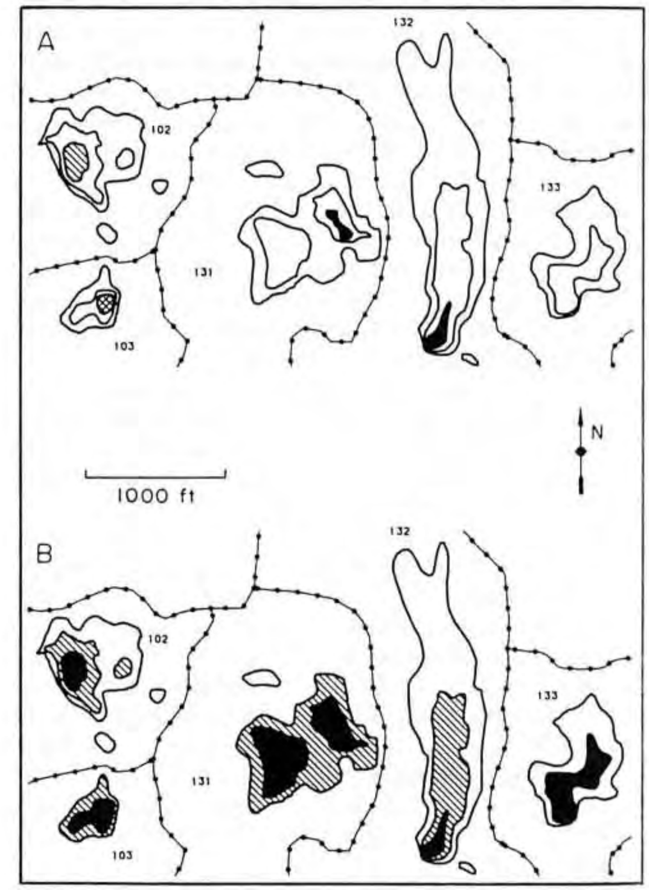

Figure 6: Examples of predicted flooding in sinkholes in a single-family residential area. A. 50-yr 3-hr rainfall with dry-antecedent conditions. B. 100-yr 3-hr rainfall with wet-antecedent conditions. Solid shade indicates complete inundation to that level. Crosshatch pattern indicates 50-100% inundation of that contour interval, and diagonal line pattern, 25-50%. Contour interval is 5 feet. Lines with dots show catchment divides. Numbers identify particular sinkholes.

Figure 6 shows a residential area with somewhat larger sinkholes. In the 50-yr, dry-antecedent-condition event, only the lowest parts of some sinkholes are flooded (Figure 6A). In the 100-yr, wet-antecedent-condition event, flooding is more extensive, with wide areas being inundated (Figure 6B). Only small sinkholes, however, are filled to the highest closed contour.

Simulations for all 164 sinkholes showed that the 50-yr, dry-antecedent-condition event results in 5 sinkholes filled to overflowing and another 6 with their topmost contour interval 50-100% filled. The 100-yr, wet-antecedent-condition event results in 27 sinkholes filled to overflowing and another 33 with their topmost contour interval 50-100% filled.

The main factors affecting the degree of flooding in a given sinkhole should be the proportion of impervious area in the watershed and the ratio of the area of the drainage basin to the volume of the sinkhole. Figure 7 shows a scatter plot of these two parameters for 81 sinkholes. Note that the area:volume ratio is much more important for predicting the degree of flooding than is the proportion of impervious area and, in fact, appears to provide a good preliminary method for assessing the likelihood of flooding.

Other factors not quantified in this study also may

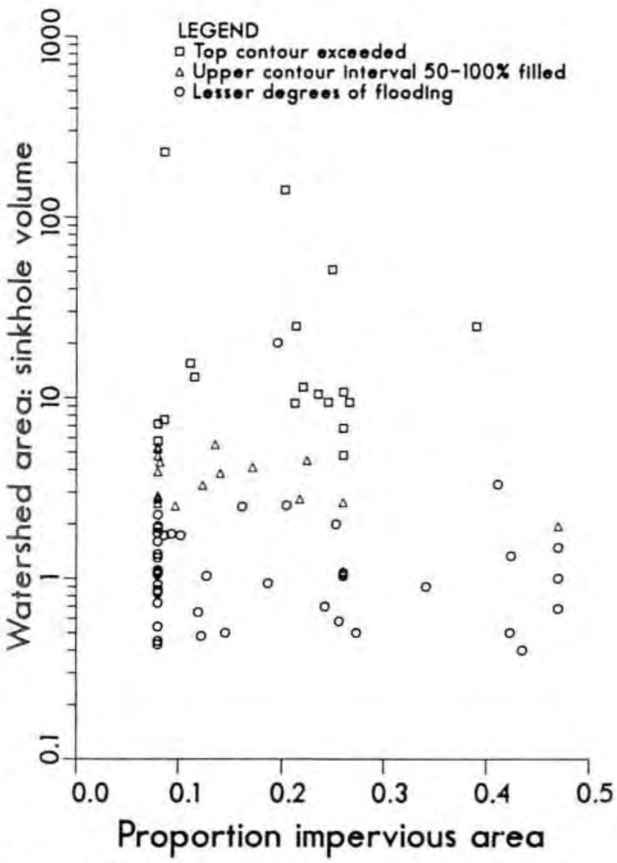

Figure 7. Plot of percent impervious area against basin-area:sinkhole-volume ratio for 81 sinkholes.

bear on whether a sinkhole is likely to represent a flooding problem. For example, the form of a sinkhole influences the likelihood of development. Deep, steep-walled sinkholes are obvious hazards and do not invite development. Shallow, gentle-walled sinkholes, however, often do not seem hazardous, and may well undergo development. Commonly, construction initially occurs within the upper contours of these sinkholes, which may pose no flooding hazard owing to adequate storage in the broad, flat bottoms. As the higher sites are taken, however, eventually development extends to the sinkhole bottoms, resulting in such hazards as ponding over roads, flooded lower levels of buildings, increased infiltration into sewer lines, and collapse due to piping of foundation material. Sinkholes 64 and 65 (Figure 5), for example, are highly developed. The amount of development in locations downgradient from sinkholes may also be important, particularly for sinkholes likely to overflow. Sinkholes 62 and 88 (Figure 5), for example, have intensive commercial development downgradient.

Discussion

The procedure outlined in this paper provides an excellent way to screen a large number of sinkholes for potential flooding hazard. Once the high-priority sinkholes are determined, however, detailed hydrologic data are required for more precise modeling. Few such data have been gathered in Cookeville. The first need is installation of recording rain gauges that can provide rainfall intensity data for five-minute or smaller time intervals. For high-priority sinkholes fed chiefly by one stream, a water-level recorder near the stream mouth would allow the stream discharge to be monitored, thereby allowing an accurate record of water entering the sinkhole.

Still more important, however, is the placement of water-level recorders within sinkholes to measure leakage rates. Leakage rates of sinkholes with regolith-covered floors can be approximated by measuring the hydraulic conductivity of the regolith, but leakage rates of sinkholes drained by swallets probably can only be determined empirically. The rate at which the sinkhole drains, of course, is important for modeling. The appropriate rainfall duration to use in the model probably will be different for fast-draining and slow-draining sinkholes. In addition, the volume of leakage must be subtracted when computing the degree of flooding in a sinkhole.

Another reason that water-level recording is important is that, as Crawford (1982, p. 8) has pointed out, in some cases even the assumption that a sinkhole floor is completely impervious may not be conservative enough. Many sinkholes are interconnected by a common subsurface conduit into which they normally drain. If there is a constriction in this conduit downstream from the sinkhole of concern and if there are upstream sinkholes feeding the conduit, flood water may back up until it actually rises out of the conduit into the sinkhole.

Another effect that can confound modeling is that the groundwater basin that is supplying water to the sinkhole may be larger than the surficial topographic basin, so that the amount of water entering the sinkhole during a heavy rain may be more than predicted. This difficulty can be corrected by using additional dye tracing to better define areas contributing flow to sinkholes.

Resource limitations allow only a small number of sinkholes to be continuously monitored. For sinkholes with lower priorities, however, it is possible at least to measure the maximum depth of flooding with only a small amount of effort. Crest-stage gauges (Leopold, 1968) can easily and inexpensively be installed in a large number of sinkholes. Aerial and ground photography of sinkholes during and after intense storms would be another quick method.

An advantage of the modeling approach discussed here is that, as leakage rates and other hydrologic data are obtained from field measurements, they can readily be incorporated into the model, thereby increasing its sophistication with time.

Conclusions

Surface runoff from 50-yr and 100-yr storms were simulated using the SWMM for 164 sinkholes and their catchment areas. Out of the 164, only 11 sinkholes were predicted to flood beyond the top closed contour or have greater than 50% of their highest contour interval inundated during the 50-yr storm with dry-antecedent soil-moisture conditions. For the 100-yr event with wet-antecedent conditions, however, a total of 60 sinkholes were predicted to similarly flood. A priority list was compiled of the 28 sinkholes most likely to produce flooding problems, based on the simulations, the amount of present development in the sinkholes, and the amount of downstream development.

The modeling procedures developed in this project will assist city planners and developers in minimizing stormwater flooding hazards. To further enhance the proper management of stormwater within urban karst areas, the following recommendations are made:

1. The GIS database should be upgraded to allow fine discretization of the study area into subbasins, including storm drains. In addition, more dye tracing of subterranean flow paths should be conducted in order to more accurately delineate groundwater drainage basins and to better understand the hydraulic connections of sinkholes.

2. Recording precipitation gauges should be installed in the study area to provide precise rainfall-intensity data needed for future modeling of runoff. For the highest-priority sinkholes, water-level recorders should be installed both in inflowing streams (to aid in the calibration of runoff) and in the sinkholes themselves (to establish hydraulic conductivities of sinkhole floors and subsurface conduits). Crest-stage recorders should be placed in other sinkholes of possible concern in order to document depth of flooding.

3. The study area should be modeled to determine the effects of overflowing sinkholes cascading into down-gradient sinkholes. In addition, high-priority sinkhole drainage basins should be modeled for various development and abatement scenarios.

References

Betson, R.P., 1976, The hydrology of karst urban areas, *in* Dilamarter, R.R. and Csallany, S.C. (editors), *Hydrologic Problems in Karst Areas*: Western Kentucky University, Bowling Green, Kentucky, p. 162-175.

Crawford, N.H., 1982, Hydrogeologic problems resulting from development upon karst terrain, Bowling Green, Kentucky: *Guidebook Prepared for Karst Hydrology Workshop, August 31-September 3, 1982*, sponsored by Groundwater Section, Region IV, U.S. Environmental Protection Agency, 34 p.

Crawford, N.H. and Groves, C.G., 1984, *Storm Water Drainage Wells in the Karst Areas of Kentucky and Tennessee*: Report prepared for the U.S. Environmental Protection Agency, Underground Water Source Protection Program Grant No. G004358-83-0, 52 p.

Faulkerson, J.; Burden, D.; Burden, K.; Edwards, C.; Kinley, T.; Lee, T.; Sparks, V.; Starnes, D.; Walls, E.; and Webster, S., 1981, *Karst Hydrology, Morphology, and Water Quality in the Vicinity of Cookeville, Tennessee*: Report to the City of Cookeville: Tennessee Technological University, Cookeville, Tennessee, 67 p.

George, D.B.; Taylor, H.N.; Pride, T.E.; Ogden, A.E.; Robinet-Clark, Y.; Forde, R.; and Mills, H.H., 1990, *Assessment of the Cookeville, Tennessee, Stormwater Management System*: Center for the Management, Utilization, and Protection of Water Resources, Tennessee Technological University, Cookeville, Tennessee, 122 p.

Hannah, E.D.; Pride, T.E.; Ogden, A.E.; and Paylor, R., 1989, Assessing ground water paths from pollution sources in the karst of Putnam County, Tennessee, *in* Beck, B.F. (editor), *Engineering and Environmental Impacts of Sinkholes and Karst: Proceedings of the Third Multidisciplinary Conference on Sinkholes and the Engineering and Environmental Impacts of Karst, St. Petersburg Beach, Florida, 2-4 October 1989*: A.A. Balkema, Rotterdam and Brookfield, p. 183-188.

Huber, W.C.; Heaney, J.P.; Nix, S.J.; Dickinson, R.E.; and Polmann, D.J., 1987, *Storm Water Management Model; Users' Manual, version III. EPA-600/2-84-109a,*

(tenth printing): Storm and Combined Sewer Section, Systems and Engineering Evaluation Branch, Wastewater Research Division, Municipal Environmental Research Laboratory, Cincinnati, Ohio.

Kemmerly, P.R., 1981, The need for recognition and implementation of a sinkhole-floodplain hazard designation in urban karst terrains: *Environmental Geology*, v. 3, p. 281-292.

Leopold, L.B., 1968, Crest stage gage of U.S. Geological Survey, *in* Tricart, J. (editor), *Study of Slope and Fluvial Processes*: Paris, Jacques et Demontrond, p. 181.

Mills, H.H. and Starnes, D.D., 1983, Sinkhole morphometry in a fluviokarst region: Eastern Highland Rim, Tennessee, U.S.A.: *Zeitschrift für Geomorphologie*, v. 27, p. 39-54.

Mills, H.H.; Starnes, D.D.; and Burden, K.D., 1982, Predicting sinkhole flooding in Cookeville: *Tennessee Tech Journal*, v. 17, (Tennessee Technological University), Cookeville, Tennessee, p. 1-20.

National Oceanic and Atmospheric Administration (NOAA), 1986, *Cookeville, Tennessee, Station Precipitation Data for 1931 through 1985 period*: NOAA, Cookeville, Tennessee Station, Cookeville, Tennessee.

National Weather Service, 1988, *Climatological data: Tennessee*: National Oceanic and Atmospheric Administration, v. 93, no. 1-12.

Pride, T.E.; Ogden, A.E.; Harvey, M.J.; and George, D.B., 1988, The effect of urban development on spring water quality in Cookeville, Tennessee: *Proceedings of the National Water Well Association Karst Conference, Dublin, Ohio*, p. 97-120.

White, E.L. and Reich, B.M., 1970, Behavior of annual floods in limestone basins in Pennsylvania: *Journal of Hydrology*, v. 10, p. 193-198.

Application of Dye Tracing to Evaluation of a Landfill Site in a Karst Terrane in the Tennessee Appalachians

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ABSTRACT

Experience has shown that all landfills leak. Therefore, when a landfill is being designed, one must assume that it will leak, assess the consequences accordingly, and prepare to write off part of the underlying aquifer as a source of potable water. Tracing, most easily and most economically done with fluorescent dyes, is the *only* reliable technique available for verifying one's interpretation of the hydrogeology of an area underlain by carbonate or other fractured rocks, determining the sites to which leakage will flow, designing a reliable monitoring system, and evaluating whether the system, once installed, is germane to the monitoring problem.

Some of these conclusions are illustrated by a site proposed for landfilling 30 miles east of Knoxville, Tennessee, in Knox Group dolomites. A different dye was simultaneously put into each of four wells drilled to the soil-bedrock contact, as much as ninety feet deep. Flow was downdip and bi-directional along the strike. Dye was recovered at more than ten wells and thirty springs, many of which are used as public and domestic water supplies, and at some sites for more than nine months.

Pre-test mapping of the potentiometric surface made it possible to predict the probable flow directions of dye and leachate. These predictions were subsequently verified.

The Tennessee Division of Solid Waste Management maintained that tracing was irrelevant to evaluation of the site; it recommended instead that hydrogeologic evaluation consist of geophysical surveys that measure electromagnetic induction and earth resistivity. The Division then recommended that three additional piezometers be drilled so that a total of seven on the site could be measured once and used to establish a groundwater divide (theirs). This divide allegedly separated that part of the property that drains to the water supplies from that part that (in one tracer test) did not. Such thinking is, at best, muddle-headed. The site is likely to be litigated.

Computer Enhancement of Downhole-Video Borehole Logs

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ABSTRACT

Video borehole logging is a commonly used geophysical logging technique whereby direct observation of the subsurface rock or well casing is desired. Information typically obtained by video borehole logging would be borehole shape, degree and extent of fracturing, evidence of large cavities, and degree of in-filling. However, recording conditions are often not conducive to the acquisition of clear, distinct logs, thus resulting in insufficient detail regarding subsurficial geologic conditions.

By utilizing a video-graphics computer workstation, it is possible to capture the entire video borehole log into a picture file. This picture file can then be viewed in either its original form or after various computer-aided enhancements have been made to the picture file. These enhancements can be as simple as manipulating the color, lighting, and/or shading of the picture file to as complex as that of rotating selected snapshots of the video log to the vertical and flattening the picture on the computer screen.

The Henderson Road Superfund site (Upper Merion Township, Pennsylvania) is located in a karst terrane where the subsurface conditions are very poorly understood. Downhole video logging was conducted in an effort to gain more direct information on the subsurface. These video borehole logs were then enhanced using a video-graphics workstation. Enhancements included adjusting color, lighting, and shading.

Introduction

Subsurficial investigations in fractured rock and karst terranes are generally much more difficult and complex than those in granular media. Although this problem is well known from attempts to exploit ground-water resources in fractured rock and karst terranes, the situation has become especially evident recently with the advent of the Superfund program. Numerous examples can be cited in which instances of known releases of contaminants to the subsurface could not be adequately monitored because subsurface transport was restricted to discrete pathways (*e.g.* fractures and conduits) as opposed to dispersion throughout a porous granular medium.

As a result of the need to better define subsurficial conditions at Superfund sites, many researchers have begun investigating the use of alternative methods designed to provide more direct information from the subsurface. These methods are primarily of two types, hydraulic and geophysical. New hydraulic techniques range from modification of existing hydraulics equations (Sen, 1987, 1988; Thrailkill, 1988) to developing better methods for hydraulic testing of aquifers (Molz and others, 1989; Milanovic, 1981, p. 235-253) to ground-water tracing studies (Molz and others, 1986; Quinlan and Ewers, 1985). Geophysical methods range from attempts to apply existing methods in varying manners to glean additional information regarding the subsurface to the development of alternative techniques for investigating the subsurface (Lange and

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Kilty, 1988; Lange and Quinlan, 1988). One of the most useful geophysical methods recently developed and now commonly used at Superfund sites is the downhole video borehole viewer. This allows a continuous video recording to be made on any standard video recorder for later repeated viewing in an office setting.

Creation of video borehole logs provides a means for visualizing subsurficial conditions in boreholes. Shape, degree and extent of fracturing, evidence of large cavities, and degree of in-filling are some of the more useful information directly available from video recording a borehole. Unfortunately, the video recording often may not be of sufficient quality to provide the detailed information still required for defining subsurficial transport conditions. By utilization of a computer-graphics workstation, it is possible to capture the entire video borehole log into a computer picture file. Enhancements can be as simple as manipulating color, lighting, and/or shading of the picture file to as complex as rotating selected snapshots of the video log to the vertical and flattening the picture on the computer screen.

Site Description and Nature of the Problem

During the 1970's, an existing water well at what has come to be known as the Henderson Road Superfund site (Figure 1) was converted from a water-supply well to a waste-injection well. Although it is believed to have been operated as an injection well for several years, very little evidence exists to either support or refute this information. Wastes injected into the well consisted of various industrial solvents pumped from tanker trucks. Operation of the injection well ceased after Pennsylvania State Police observed the owners of the site clandestinely injecting industrial-solvent waste into the well.

The Henderson Road site consists of 7.6 acres located at the intersection of Henderson and Church Roads. It is occupied by a sanitation company, several automobile shops, and a drilling contractor. Approximately 2000 feet to the north and downgradient of the Henderson Road site is the Upper Merion Reservoir (UMR), an old rock quarry now used as a water supply for the Township of Upper Merion. Groundwater is pumped from the UMR at a rate of approximately 7.5 mgd, causing substantial drawdown in the local aquifer for several square miles in the area. The Henderson Road site is also approximately 350 feet upgradient of a local lumber company, where a water-supply well serves 15 employees (USEPA, 1988).

The Henderson Road site lies within the Piedmont Province of the Appalachian Highlands and is situated on metamorphosed limestone and dolostone (Elbrook limestone, Conestoga limestone, and Ledger dolostone). Gray to yellowish gray, the Elbrook limestone is a commonly silty, sandy, siliceous limestone interbedded with dolomite whereas the Conestoga limestone, the main rock type underlying the site, is gray to bluish-gray and finely to coarsely crystalline. Both exhibit schistocity. The Conestoga Formation unconformably overlies the Elbrook and Ledger formations, and is thereby younger than both of them. The Ledger dolostone is light gray, massive, coarsely crystalline, sparkling, and contains siliceous layers. All of the rocks have been subjected to intense deformation as can be seen by the complex series of anticlines, synclines, fractures, and faults in the general area (Nevius , 1987, p. 9-12).

Karstification of the limestones and dolostones has resulted in significant alteration of the regional geology. Sinkholes observed throughout the area and solution voids detected in boreholes provide substantial evidence that dissolution of the carbonate rocks is extensive and that subsurface conduits probably play an important role in the flow and transport of liquids and gases. However, these conditions are sufficiently complex so as to prevent a clear comprehension of the hydrogeology of the site regarding the rate, direction, and extent of subsurface transport. Several efforts, from hydraulic testing to geophysical investigations, have been attempted with the intention of better defining the subsurficial geology.

Vacuum testing of selected wells for possible vacuum extraction of subsurficial volatile gases yielded highly ambiguous results. Two radically different interpretations from the same vacuum-extraction data have been offered with no resolution as to which is more correct. In both instances, interpretations were directed toward establishing zones of interconnectedness. Unfortunately, this has not yet been possible. As a result, substantial confusion regarding pathways utilized for gaseous and liquid transport in the vadose zone still remains.

A second hydraulic analysis of the subsurface at the Henderson Road site consisted of conducting conventional aquifer tests on selected wells. Figure 2 is a semilog plot of the late-time dimensionless time-drawdown data of one on-site pump test with the theoretical Jacob Straight line plotted on the same graph. However, use of the Jacob method in fractured and/or karstic terranes is generally not valid owing to the extreme limiting conditions associated with it. By plotting both the field data and the theoretical Jacob curve on the same graph, it is possible to determine if the Jacob method is in fact a valid method of aquifer analysis (Sen, 1988). According to Sen, if the slopes of the two plots are parallel, then the use of the Jacob method could be considered valid for analysis of the site aquifer-test data. However, the slopes of the two lines are not parallel in this case, so the Jacob method of aquifer-test analysis can only be considered a first approximation. The flatter slope of the field data would indicate that the actual transmissivity is less than that calculated whereas the storativity is greater than that calculated. Also, leakage from surrounding fractures may be inferred from the field plot. In terms of the pump-and-treat operation planned for this site, accurate determination of subsurficial storage, flow, and transport parameters is of paramount importance.

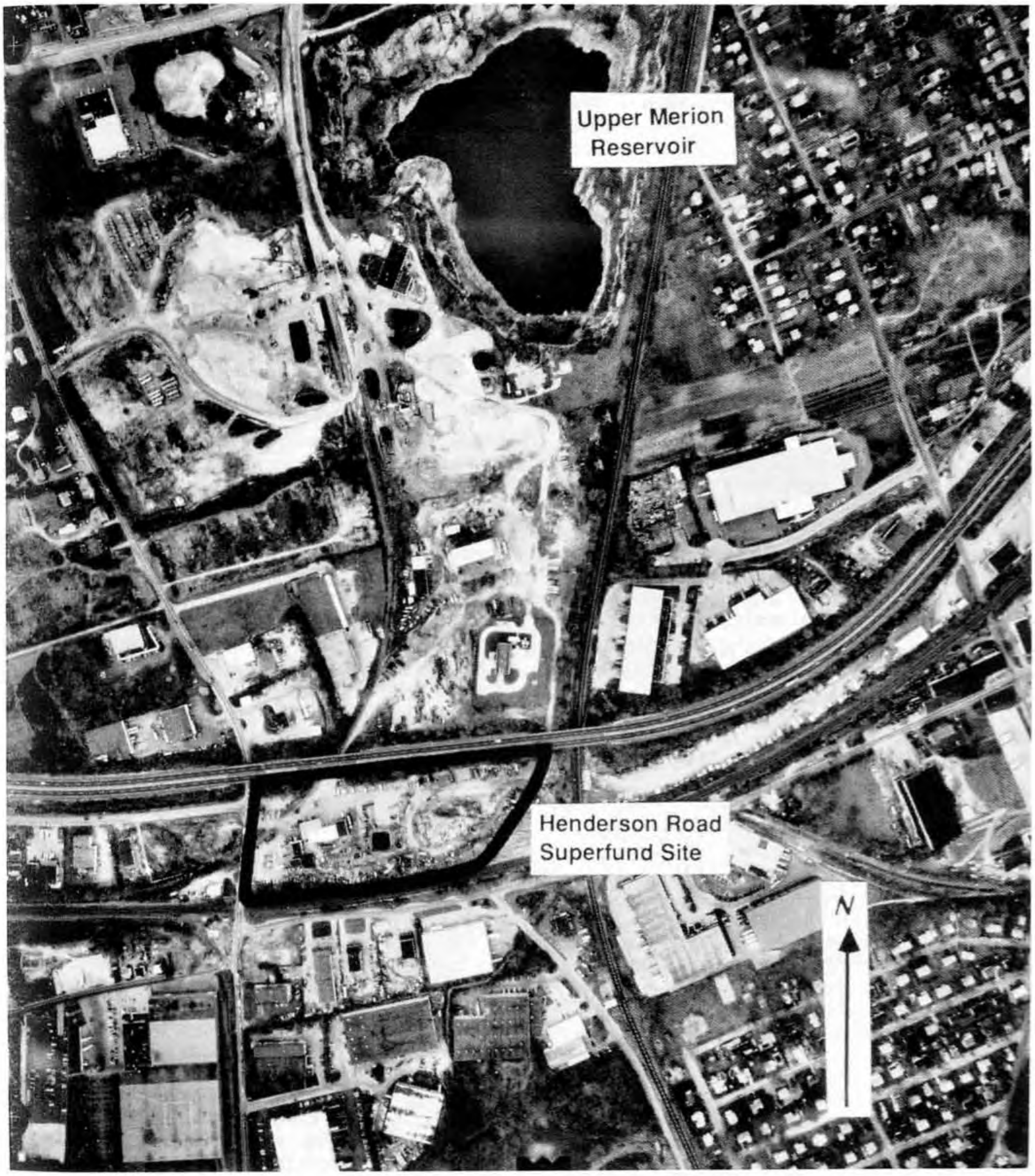

Figure 1: Aerial photograph of the Henderson Road Superfund site and the Upper Merion Reservoir, downgradient from the site. The karst landscape is masked by the high level of urbanization in the area, making field interpretations more difficult.

Borehole geophysical methods were also utilized at the site to further define the zones of greatest importance regarding subsurficial contaminant transport. These consisted primarily of caliper, resistivity, spontaneous-potential, and natural-gamma logging. Downhole television logging of the boreholes onsite was also conducted. However, none of these methods have yet been able to provide the level of detail necessary for accurate determination of

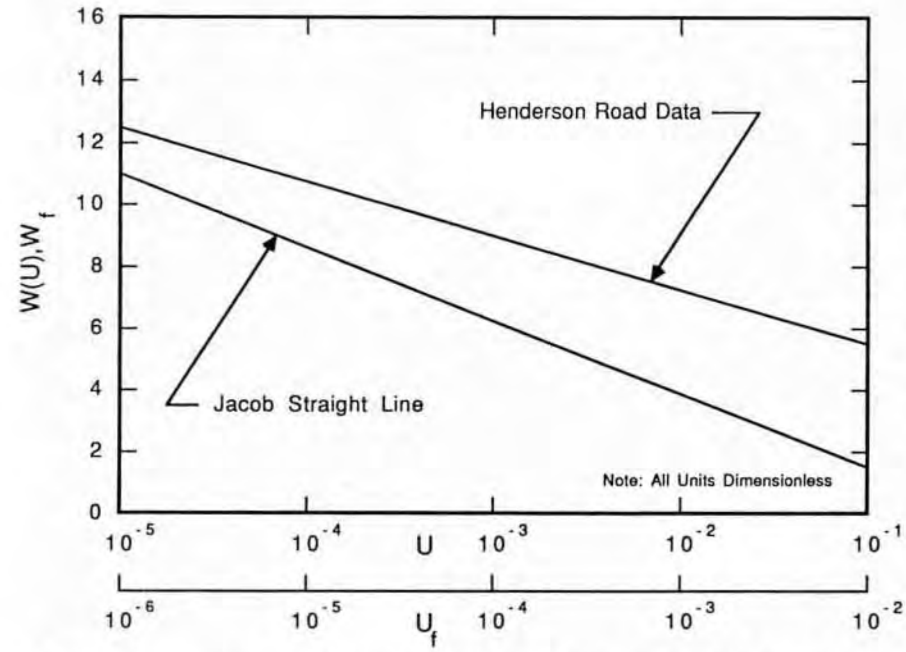

Figure 2: Comparison of the dimensionless time-drawdown data for the Henderson Road site with the theoretical Jacob's Straight Line. *Note* that the lines are not parallel, indicating the inapplicability of the Jacob method for aquifer-test analysis at this site.

subsurficial zones most likely to provide pathways for contaminant transport. In particular, the downhole-video borehole logs were especially disappointing.

Borehole video logs of the monitoring wells for the Henderson Road site were collected with the intention of providing direct observation of subsurficial geologic conditions. Specific objectives included identification of fractures and large voids, both of which may be responsible for significant amounts of contaminant transport. However, the quality of the video logs was significantly less than desired and/or required for gaining the necessary information. Generally, the camera angle was not conducive to a clear display of important geologic features, the lighting was excessively bright, and the resolution was insufficient. Clear prints of selected features from each borehole were desired for inclusion in the official public record, but these could not be obtained because of the relatively low quality of the video logs.

Owing to the importance of the information needed, it was determined that either new video logs would be required for each well or some other method would have to be devised if the information desired was ever to be acquired. Objections to relogging each well forced consideration of alternative methods for acquiring the information necessary for determining the important flow and transport zones at the Henderson Road site. An alternative that shows considerable promise is the capturing of the video logs into a computer picture file that can then be manipulated for enhanced clarity.

Description of Equipment

Video logging of the boreholes was initially conducted by suspending a television camera equipped with a 4.5 mm wide-angle lens and a manually adjusted f-stop, fixed-focus speed control. Lighting was provided by a low-light-level silicon-target vidicon tube to allow image detection by the camera with minimum lighting power requirements. The rest of the video recording equipment consisted of a nine inch, studio-quality television monitor and VHS video recorder. It is useful to note here that use of a Super VHS would have probably greatly enhanced the overall quality of the recording.

The computer equipment used to capture and enhance the video frames was a Commodore Amiga 2000 Personal Computer. Image capture was carried out on an A2000 equipped with a 68030 accelerator board, 8 megabytes of 32-bit dynamic RAM and the recently developed Video Toaster by Newtek Inc. Image enhancement was performed on an unaccelerated A2000 with 3 megabytes of 16/32-bit RAM using off-the-shelf image-processing software. The Amiga computer is manufactured by Commodore Business Machines as a high-quality graphics-and-sound capable personal computer with proprietary audio- and video-chips, multitasking operating system, and Graphical User Interface (GUI). Total equipment costs, including both hardware and software, for an accelerated system with a large storage capacity are less than $10,000, making the technology very affordable.

Image capture on the A2000 was carried out through use of the Video Toaster, a new professional-level framebuffer/video-manipulation board and software. The Toaster is essentially a computer-on-a-card which provides the A2000 with full television color and resolution while furnishing it with the capability of performing real-time video manipulation. Hence, output from Video Toaster approaches broadcast-quality video on a computer. The Video Toaster allows for the production of network-quality television on an A2000 by providing several very powerful video tools such as a 4-Input Production Switcher, Digital Video Effects, Lightwave 3D Animation, Toaster Character Generator, Toaster Paint, Dual Frame Buffers/Genlock, ChromaFX Color Processor, and Frame Grabber/Frame Store.

Discussion

Selected borehole-log frames were captured into Newtek's Video Toaster buffer from a standard VHS recorder. The video buffer was then written to a graphics file in a

24-bit color format (24 color representations for each pixel). These files were then loaded into a graphics-conversion program called *The Art Department* and manufactured by ASDG, and converted to the Amiga's standard 16-bit Interleaved File Format (IFF). Standard 24-bit graphics allows for millions of colors to be displayed on screen at one time and has generally been available only at a professional level. The current Amiga display- and graphics-software standards are 16 bit. This allows several on-screen color displays depending on resolution and display mode. The variations are 16, 32, 64, or 4096 colors on screen at one time. Resolutions range from a 320x200 to an Overscan mode of approximately 800x500. The images used in this presentation were captured in the 24-bit, millions of colors mode, and then converted to an overscanned Hold and Modify (HAM) mode that provided 4096 colors on screen at a resolution of 768x530 pixels.

Various motion-removal algorithms were used in the capture of selected frames to remove flicker that occurs when a video tape is paused while in the "play" mode. This could be accomplished because each frame consists of two fields (every other scan line) that provide the Video Toaster with the ability to grab four fields at once, thus acquiring the full amount of resolution and color fidelity (McMahon, 1991). After flicker removal, the frames may be loaded into Toaster Paint for manipulation. They may also be loaded into Lightwave 3D and mapped onto various objects or they may be loaded into the Video Toaster's other manipulation algorithms for additional enhancement.

Frames captured using the Video Toaster, once converted to the 16-bit standard, were manipulated using several off-the-shelf or public-domain image-processing programs. Because of the number of colors, resolution, or size of the files (100,000+ bytes), only one package, PIXmate, was found suitable for use in this effort. PIXmate is an image-processing software package published by Progressive Peripherals & Software, Inc. The capabilities of the software are too extensive to detail here but it allows for such things as (1) color and pallet manipulation, (2) cutting and copying segments of an image, (3) color separation, and (4) image enhancement through such bit-map operations as averaging, randomization, local-contrast enhancement, unsharp masking, and median filters. Several of these image-enhancement processes were carried out on images captured from tape of the Henderson Road site.

Figure 3 is a 64-level gray-scale image modification of the injection well at a depth of 90 ft. It was created by converting a 4096-color HAM image. *Note* the excessive brightness on the right-hand side of the figure. Numerous efforts to modify this aspect of the video frame have not yet been possible, providing evidence that greater care in the original video-recording is warranted. There is a large void on the left side of the figure, the dimensions of which are not readily discernible. To further enhance this image, an edge to the gray-scale image was overlaid onto the file producing more detail as shown in Figure 4. This was accomplished by applying a Laplacian Edge Detection Operator to Figure 3. Improvements to the clarity of selected features in Figure 3 are readily apparent in Figure 4, although further improvements would clearly be desirable. A final refinement of this image is shown in Figure 5. This enhancement was created by removing several shades of gray below a certain threshold level. Ostensibly, this should show those areas that are most directly, or strongly illuminated.

The water table at the site was encountered at a depth of 118 ft and is shown in Figure 6. Figure 6 is also a 64-level gray-scale modification created in the same manner as Figure 3. Enhancements to this image (Figure 7) were performed by taking a 32-level gray-scale image using Local Contrast Enhancement on the original 4096-color image and applying a Laplacian Edge Detection Operator to Figure 6. Various features are clearly highlighted using the computer software as can be seen by comparing Figures 6 and 7.

Figure 3: A 64-level gray-scale image of the injection well at the Henderson Road site, taken at a depth of 90 feet. This image was converted from the original 4096-color HAM mode. *Note* the excessive brightness on the right side of the image and the large void on the left side of the image.

What cannot be shown in this paper are the color images -- the original frames captured by the Video Toaster. These were more amenable to modification because of the ability to change the various shades of color. In one enhancement effort, the red component of the spectrum was enhanced to give a more "realistic" look to the image rather than the predominant yellow hue given off by the artificial light used during the video logging operation.

Figure 4: A 64-level gray-scale image of the 90-foot depth with Laplacian Edge Detection overlaid onto the original gray-scale image. *Note* the enhanced view of selected features in the image.

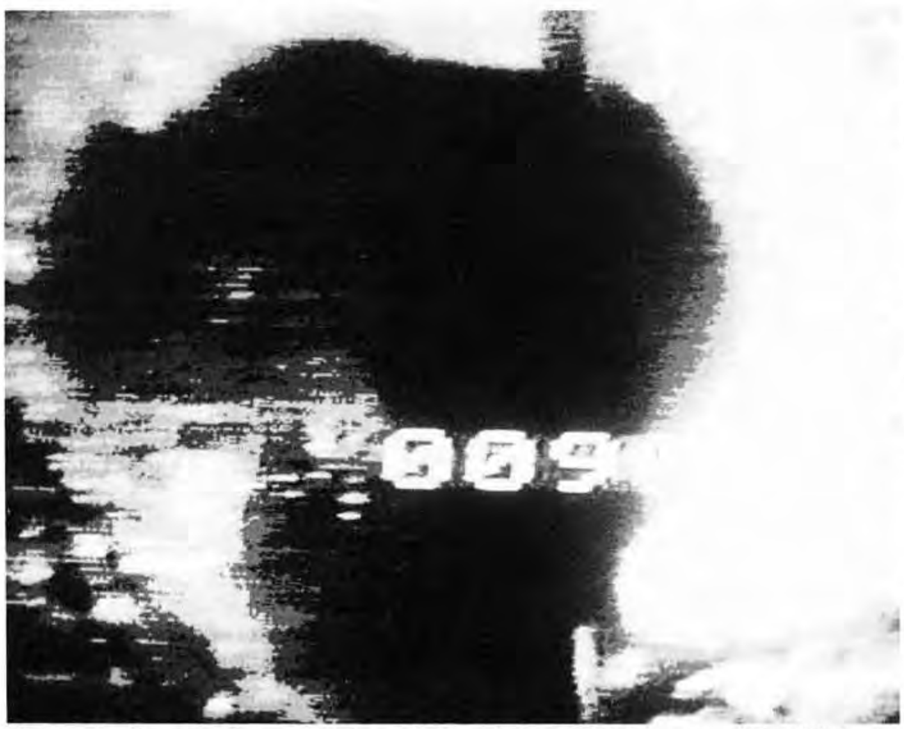

Figure 5: Gray-scale conversion of the original HAM image of the 90-foot depth with selected background gray-scale levels removed to allow for enhanced clarity of those features most strongly illuminated.

Whereas video enhancement has been shown to be useful, limitations related to the original video-tape quality are still a major impediment to the enhancement process. Improvements in video logging, therefore, should begin with proper collection of the video logs, preferably with a professional geologist overseeing the logging operation to ensure that important features are being given the necessary care required for good visual display. Use of a Super VHS recorder may also lead to better logging of features and would be more amenable to computer enhancement.

Conclusions

Video logging may be regarded as a very valuable borehole-geophysical technique for displaying actual subsurface conditions; however, it may not be enough, particularly in contaminant hydrogeology. Capturing the video log into computer picture files and manipulating these files can greatly improve observational interpretations of subsurficial conditions. As technology continues to advance, computer refinements will become ever more sophisticated.

Acknowledgments

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References

Lange, A.L. and Kilty, K.T., 1988, Detection and mapping of karst conduits from the surface by acoustic and natural potential methods: *Contract Report Prepared for the Superintendent, Mammoth Cave National Park, and the Environmental Monitoring Systems Laboratory of the U.S. Environmental*

Protection Agency, Las Vegas, Nevada: The Geophysics Group, Wheat Ridge, Colorado, 40 p.

Figure 6: A 64-level gray-scale image of the injection well at the Henderson Road site, taken at a depth of 118 feet. This image was also created from the original 4096-color HAM image. *Note* the water table evident in the image.

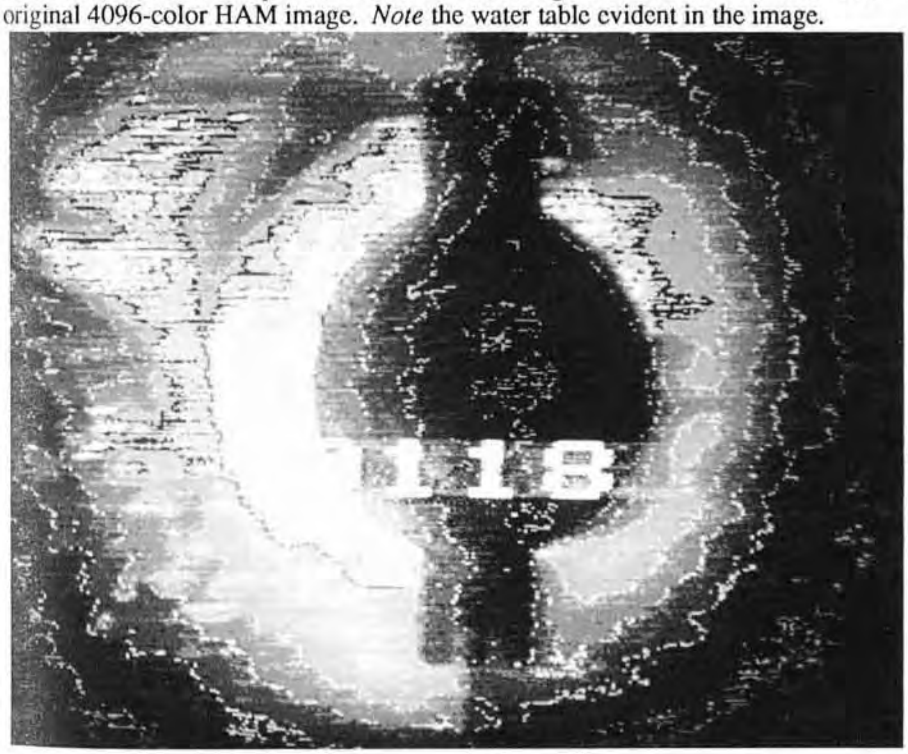

Figure 7: A 32-level gray-scale image of the 118-foot depth of the injection well in which Local Contrast Enhancement was used to enhance the original 4096-color HAM image. *Note* how clearly features are outlined with this type of enhancement.

Lange, A.L. and Quinlan, J.F., 1988, Mapping caves from the surface of karst terranes by the natural potential method: *Proceedings of the Second Conference on Environmental Problems in Karst Terranes and Their Solutions (Nashville, Tenn.)*: National Water Well Association, Dublin, Ohio, p. 369-390.

McMahon, F., 1991, Newtek's Video Toaster: A new era in Amiga Video: *Amazing Amiga Computing*, v. 6, p. 48-62.

Milanovic, P.T., 1981, *Karst Hydrogeology*: Water Resources Publications, Littleton, Colorado, 434 p.

Molz, F.J.; Guven, O.; Melville, J.G.; and Keely, J.F., 1986, *Performance and Analysis of Aquifer Tracer Tests with Implications for Contaminant Transport Modeling: EPA Project Number EPA/600/2-86/062*: Robert S. Kerr Environmental Research Laboratory of the U.S. Environmental Protection Agency, Ada, Oklahoma, 88 p.

Molz, F.J.; Morin, B.H.; Hess, A.; Melville, J.G.; and Guven, O., 1989, The impeller meter for measuring aquifer permeability variations: Evaluation and comparison with other tests: *Water Resources Research*, v. 25, p. 1677-1683.

Nevius, J.G., 1987, *A Hydrogeologic Analysis of the Carbonate Aquifer and Associated Surficial Deposits in Upper Merion Township, Pennsylvania*: M.S. thesis (unpublished), Geology Department, University of Pennsylvania, 85 p.

Quinlan, J.F. and Ewers, R.O., 1985, Ground water flow in limestone terranes: Strategy rationale and procedure for reliable, efficient monitoring of ground water quality in karst areas: *Proceedings of the Fifth National Symposium and Exposition on Aquifer Restoration and Ground Water Monitoring, Columbus, Ohio*: National Water Well Association, Worthington, Ohio, p. 197-234.

Sen, Z., 1987, Non-darcian flow in fractured rocks with linear flow pattern: *Journal of Hydrology*, v. 92, p. 43-57.

Sen, Z., 1988, Dimensionless time-drawdown plots of late aquifer test data: *Ground Water*, v. 26, p. 615-618.

Thrailkill, J., 1988, Drawdown interval analysis: A method of determining the parameters of shallow conduit flow carbonate aquifers from pumping tests: *Water Resources Research*, v. 24, p. 1423-1428.

U.S. Environmental Protection Agency, 1988, *Record of Decision for the Henderson Road NPL Site Injection Well Operable Unit, Hazardous Waste Management Division*: U.S. EPA, Region III, Philadelphia, Penn.

An Attempt to Model an Appalachian Karst Aquifer Using MODFLOW

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ABSTRACT

In southeastern West Virginia, the Middle Mississippian Greenbrier Group, a limestone with some thin calcareous shales, is 600 to 900 feet thick and strikes about N 25 E where it crops out in Greenbrier County. This area was chosen for aquifer modeling because of the previous existence of a detailed database, including dye tracing. MODFLOW, a finite-difference aquifer-modeling program available from the U. S. Geological Survey, has features that allow some of the peculiarities of karst aquifers to be incorporated. For purposes of modeling, the Greenbrier Group was treated as an aquifer consisting of a single layer. The aquifer rests on a thick shale (the Lower Mississippian Maccrady Formation) which is broadly folded into a syncline, and sharply faulted upward in the southwest near the town of Asbury. The Maccrady Formation acts as a downward barrier to ground-water flow, and a westward boundary to flow where it is faulted. Structural contours on the Maccrady were determined from geologic maps, and interpolated to an array of values fitted to the finite-difference grid. The finite-difference grid used for the model was oriented parallel to stratigraphic strike so that aquifer anisotropy could be simulated. The anisotropy factor was determined from directional measurements of 168 cave passage segments. Diffuse aquifer recharge (infiltration of rainfall) was estimated from climatic records. Concentrated aquifer recharge (cave insurgences with capture zones on detrital rocks outside the grid area) were simulated by using recharge wells. Aquifer hydraulic conductivity was determined from well-pumping tests (diffuse flow) and dye-trace travel times (conduit flow). The model was run repeatedly to determine the effects of varying the input factors. Making the anisotropy or the hydraulic conductivity higher caused the hydraulic head to drop and made the effect of the cave insurgences less noticeable. All aquifer solutions showed a ground-water divide between Spring Creek and the Greenbrier River which was close to the one actually determined by tracer tests.

Introduction

The inherent complexity of karst aquifers, such as concentrated recharge, discharge, and conduit flow, present great difficulties to normal modeling methods. A well-studied West Virginia limestone aquifer presents an opportunity to simulate water flow through a karst aquifer. The purpose of modeling in this study was to experiment with how well the model approximates reality, and how sensitive the model is to varying the input factors. MODFLOW, a finite-difference aquifer-modeling program available from the U.S. Geological Survey (McDonald and Harbaugh, 1984), has some features which allow some of the peculiarities of karst aquifers to be incorporated. A commercial preprocessor was used to generate the data files (translate structure contours, for example, to a table of numbers in the correct format). The results of the model were gridded and contoured by a commercial graphics package.

The study area covers approximately 100 square miles of the outcrop extent of the Middle Mississippian Greenbrier Group in central Greenbrier County, West Virginia. The Greenbrier Group consists of 600 to 900 feet of limestones, with a few thin shales. The Lower Mississippian Maccrady Formation, a red shale, underlays the Greenbrier Group. The limestone is partially capped by the Lillydale Shale of the Upper Mississippian Mauch Chunk Group (Bluefield Formation). In the study area, the Greenbrier Group outcrop belt strikes about N 25 E and stretches roughly between the Greenbrier River to the south, and about 15-20 miles northeast to Spring Creek (Figure 1).

The geologic structure of the study area is dominated by broad, asymmetric, northeasterly trending folds of low plunge, and several *en echelon* reverse-fault/fold complexes. Differential erosion of the limestone has resulted in deep karst valleys trending along stratigraphic strike and receiving surface drainage from surrounding detrital rocks.

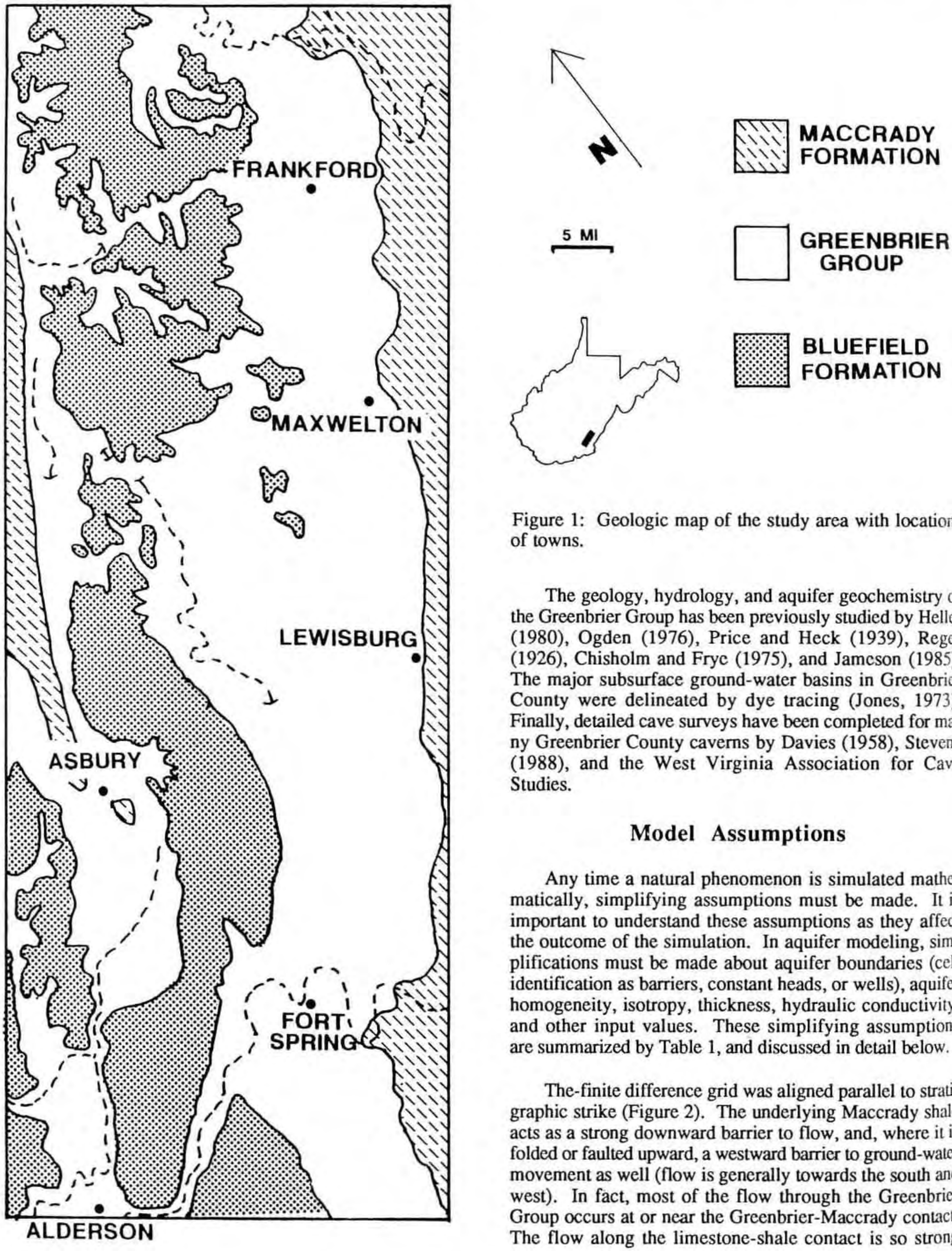

Figure 1: Geologic map of the study area with locations of towns.

The geology, hydrology, and aquifer geochemistry of the Greenbrier Group has been previously studied by Heller (1980), Ogden (1976), Price and Heck (1939), Reger (1926), Chisholm and Frye (1975), and Jameson (1985). The major subsurface ground-water basins in Greenbrier County were delineated by dye tracing (Jones, 1973). Finally, detailed cave surveys have been completed for many Greenbrier County caverns by Davies (1958), Stevens (1988), and the West Virginia Association for Cave Studies.

Model Assumptions

Any time a natural phenomenon is simulated mathematically, simplifying assumptions must be made. It is important to understand these assumptions as they affect the outcome of the simulation. In aquifer modeling, simplifications must be made about aquifer boundaries (cell identification as barriers, constant heads, or wells), aquifer homogeneity, isotropy, thickness, hydraulic conductivity, and other input values. These simplifying assumptions are summarized by Table 1, and discussed in detail below.

The-finite difference grid was aligned parallel to stratigraphic strike (Figure 2). The underlying Maccrady shale acts as a strong downward barrier to flow, and, where it is folded or faulted upward, a westward barrier to ground-water movement as well (flow is generally towards the south and west). In fact, most of the flow through the Greenbrier Group occurs at or near the Greenbrier-Maccrady contact. The flow along the limestone-shale contact is so strong

Simplifying Assumptions	Reality
Single layer	Several layers separated by thin leaky confining units
Constant thickness	Thickness varies from approximately 600 feet at Spring Creek to 900 feet at the Greenbrier River
Single K	Double K because of both diffuse and conduit flow
Anisotropy factor can be determined from cave passage frequency and orientations	Cave passages parallel to strike are larger than cross-strike passages
Homogeneous	Anisotropy probably varies with thickness, facies changes, and degree of karstification

Table 1: A summary of the simplifying assumptions made for modeling purposes.

that there are over 60 miles of mapped caverns (known as "contact caves") at this horizon. Because the Maccrady shale is such a significant hydrologic barrier, structure contours of the Greenbrier-Maccrady contact were constructed from geologic maps (Figure 3). These structural contours were fitted to the finite-difference grid dimensions, and the bottom elevations for the Greenbrier Group were interpolated for each cell. Although other thin shales exist in the Greenbrier Group that will perch or confine water, these thin shales are not nearly as hydrologically significant. Thus for modeling purposes, the Greenbrier Group was treated as a single layer of constant thickness. In the study area, the overlying shale of the Bluefield formation is regionally extensive enough to confine the Greenbrier Group only in the synclinal area below Muddy Creek Mountain (between Fort Spring and Asbury, Figure 1). MODFLOW can be programmed to determine whether flow is confined or unconfined based on the computed hydraulic head and the elevation of the top of the aquifer, and then modify the flow equations accordingly.

Spring Creek, Muddy Creek, Mill Creek, and the Greenbrier River (Figure 4) act as the major ground-water discharge zones for the area, and are treated as constant-head cells in the grid. This means that the hydraulic head does not change for these surface streams (in reality, it probably

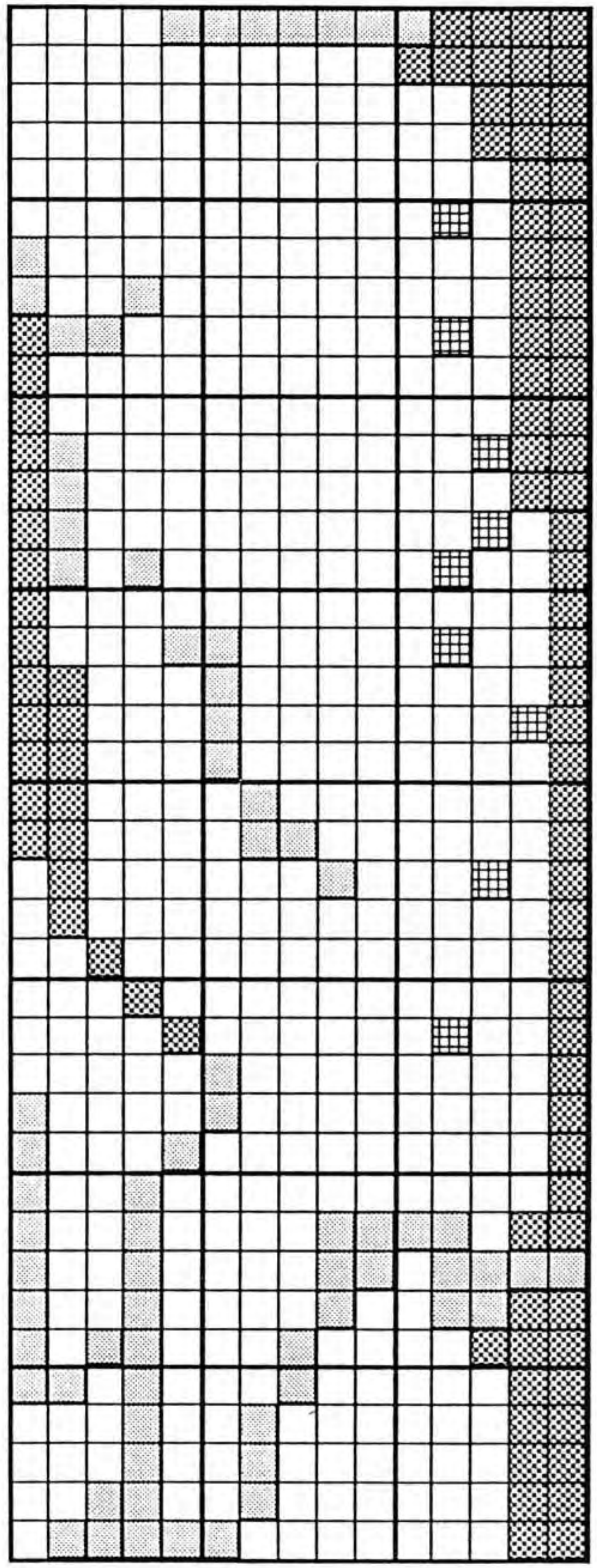

Figure 2: Finite-difference grid showing cell identification. Culverson and Milligan Creeks appear as constant-head cells in this example.

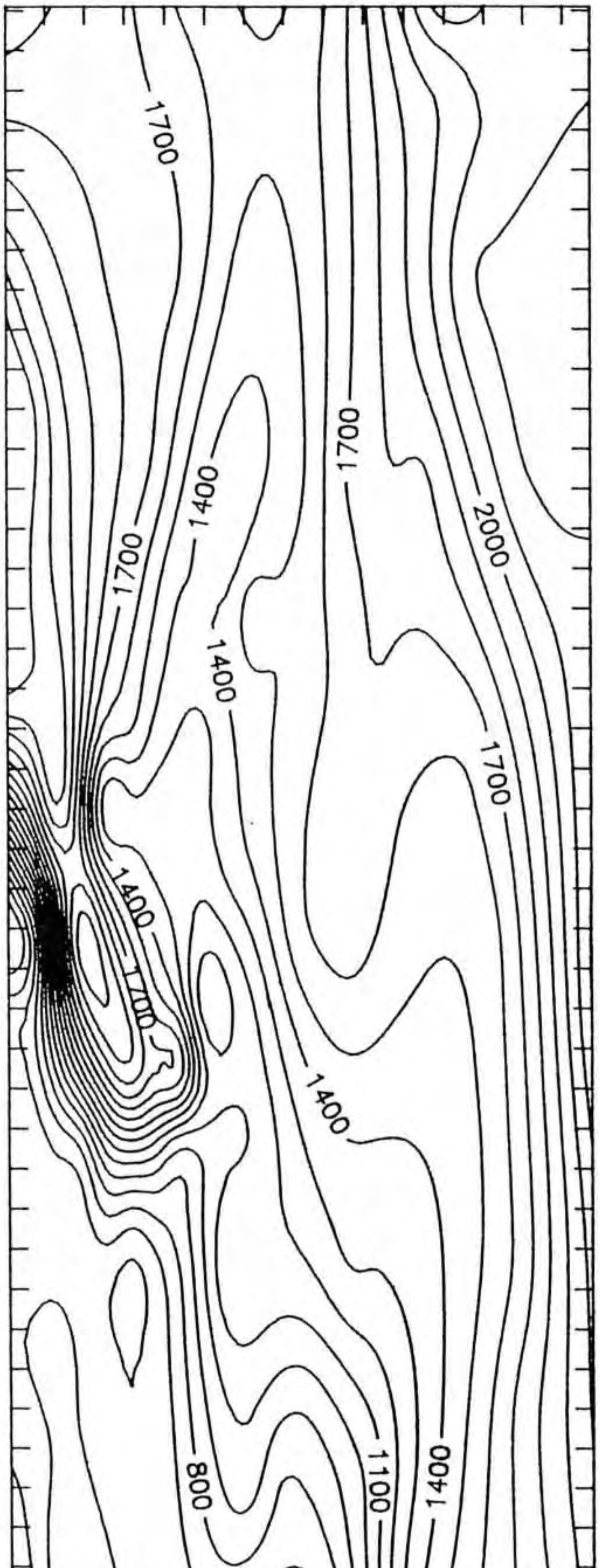

Figure 3: (At left) Structural contours (feet above mean sea level) for the contact between the Maccrady Shale and the Greenbrier Group.

varies by 20 feet or less). Culverson and Milligan Creeks were less easy to characterize because these partially segmented streams may be perched or even dry during some parts of the year. For experimental purposes, therefore, the model was set up two ways: first treating these streams as constant-head cells, and later removing them as such, but instead treating their downstream termini as cave insurgences (recharge wells, to the model).

Diffuse-aquifer recharge was estimated from climatic records. Concentrated-aquifer recharge occurs as small streams flow along the surface of the detrital rocks and are captured by caves (both mapped and those only suspected) in the Greenbrier Group. These cave insurgences were treated as recharge wells in the grid. Davis Spring, the major conduit spring that drains the area south of Maxwelton, was treated as a constant-head cell along the Greenbrier River, and not as a pumping well. Although the area does contain some widely spaced domestic water wells, the discharge from these is trivial compared to the capacity and high conductivity of the limestone aquifer. So these wells were ignored for modeling purposes.

Short-duration aquifer pumping tests (Heller, 1980; Ogden, 1976) and dye-tracing (Jones, 1973) have determined the hydraulic conductivity of the Greenbrier Group. The results of eight pumping tests yielded an average conductivity of 5.1×10^{-4} feet per minute (7.3×10^{-1} feet per day). Although it is unknown whether any of the wells tested were receiving conduit flow, this rate of hydraulic conductivity is rather typical for diffuse-type flow in fractured aquifers (Heath, 1983). Dye tracing, on the other hand, which directly measures travel times via conduit flow, yielded an average hydraulic conductivity for the Davis Spring basin of 1.4 feet per minute at low flow (or about 2000 feet per day). Although these two estimates of hydraulic conductivity for the Greenbrier Group are quite different, it is not unusual in karst aquifers to have both diffuse and conduit modes of flow at the same time (Shuster and White, 1971). For modeling purposes, however, a compromise value of 1×10^{-2} feet per minute gave the most realistic solutions.

Hydraulic conductivity in the Greenbrier Group is also probably highly anisotropic, with flow parallel to stratigraphic strike much greater than that across strike. MODFLOW allows the user to insert an anisotropy factor (TRPY) to account for this. The value of this factor was estimated by compiling a cumulative length-orientation diagram for 168 straight cave passages (Figure 5). These were then subdivided into strike-aligned versus cross-strike passages, resulting in a ratio 3.67 to 1. Because the finite-difference grid was oriented with the y axis parallel to strike, the y-to-x conductivity ratio (TRPY) was assigned this value. In reality, however, strike-aligned cave

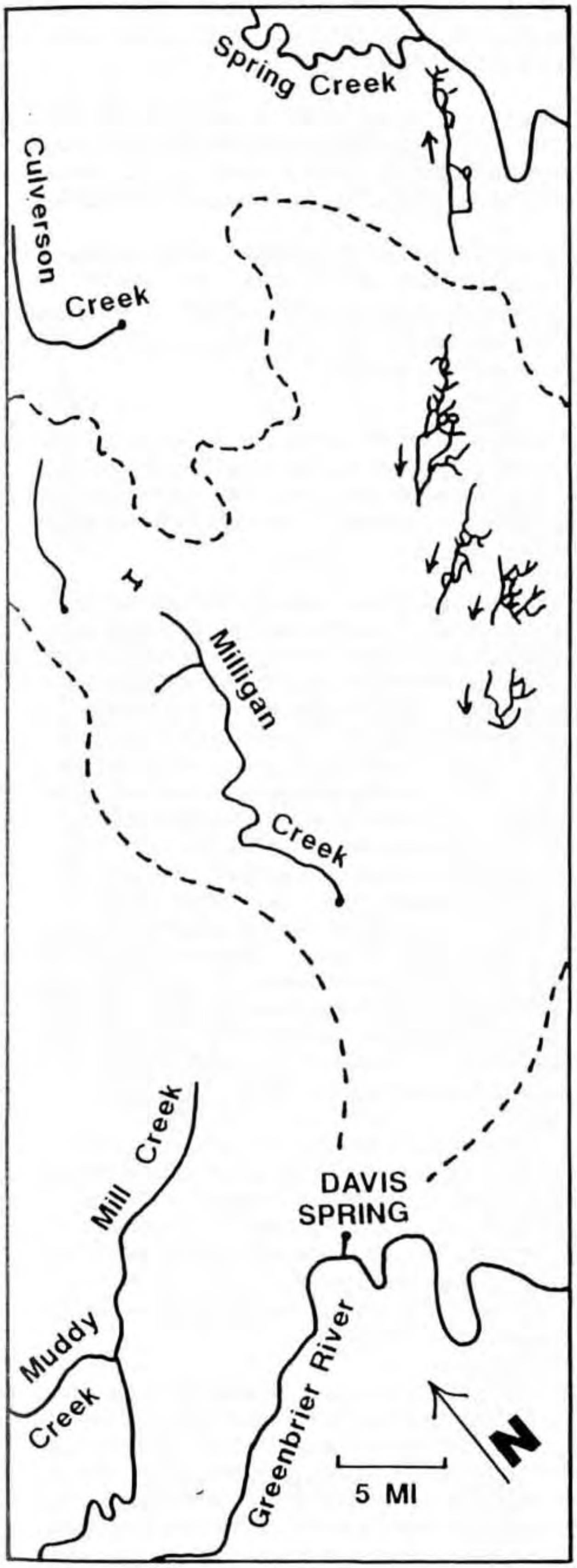

Figure 4: (At left) Map of study area showing surface drainage, locations of known contact caves (arrows), and subsurface drainage divides of the Davis Spring basin (dashed lines).

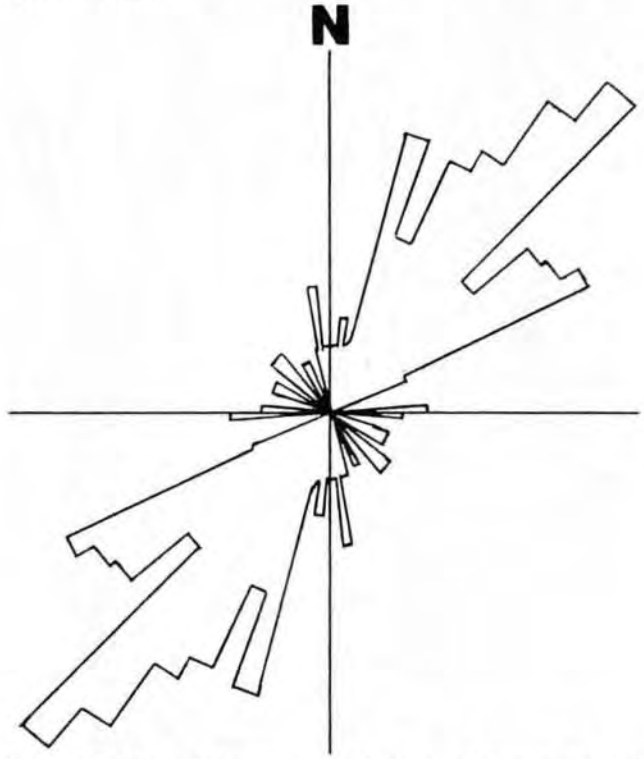

Figure 5: Cumulative length-orientation diagram for 168 straight cave-passage segments.

passages are generally larger in volume than cross-strike passages, so this TRPY should be regarded as a minimum value.

Model Results

The model was run repeatedly to determine the effects of varying the input factors. Each result was compared with: (1) mapped water-table elevations for the area, based on 35 wells (Figure 6), and (2) drainage divides for the Davis Spring Basin, based on dye-tracing results (Figure 4). Model results that were not reasonably similar to reality (given the sparsity and noisiness of the well data) were discarded. Model solutions that were realistic often showed "dry" cells around the edges, near the no-flow boundaries. This occurrence could be expected, because the aquifer is thin and topographically high in these areas.

A typical model solution is shown by Figure 7. This result treats Milligan and Culverson Creeks as constant-head cells, uses a hydraulic conductivity of 1×10^{-2} ft/min, and a TRPY of 3.67. The ground-water divide between Spring Creek and the Greenbrier River lies at about 2040 feet in this example, and coincides (roughly) with the one actually determined by dye tracing (Jones, 1973). About a

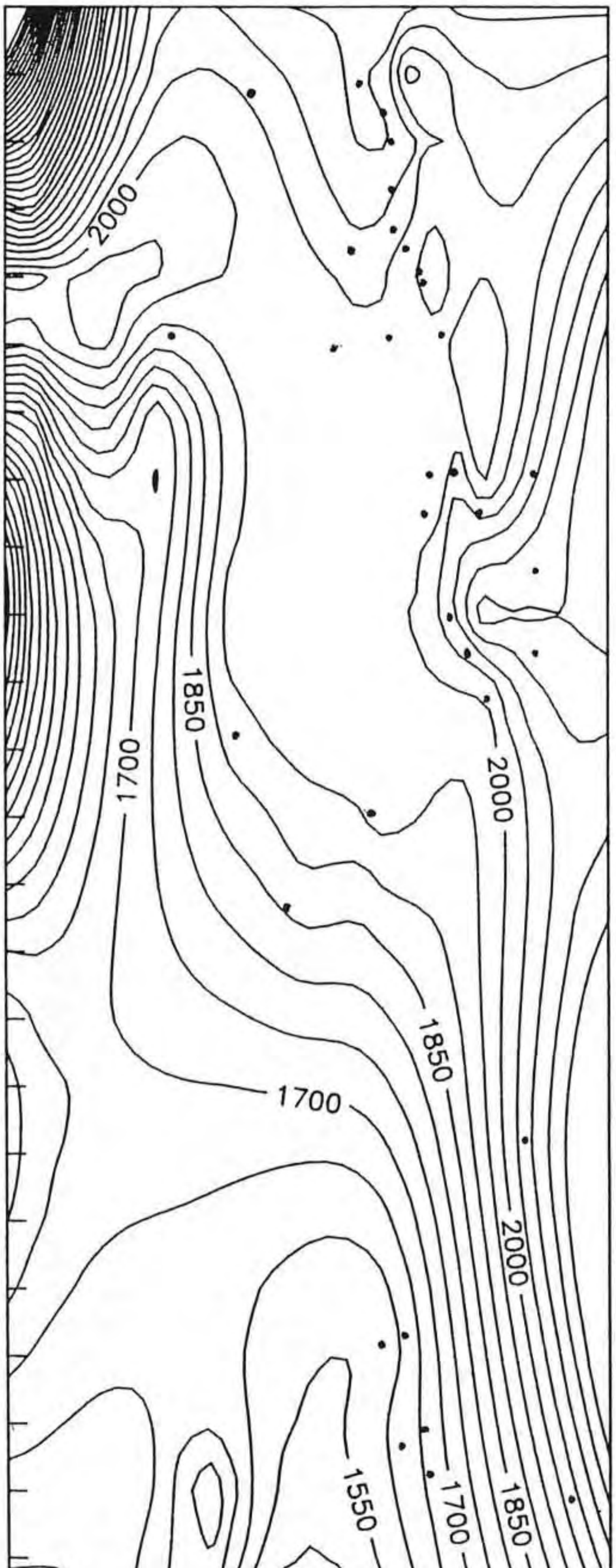

Figure 6: (At left) Contours on the water table (feet) above mean sea level) plotted from measurements of 35 wells (shown as small dots).

Figure 7: (Next page, at left) A typical model solution for a run which included Milligan and Culverson Creeks as constant-head cells, used a hydraulic conductivity of 10^{-2}, a TRPY of 3.67, and an aquifer thickness of 1000 feet.

Figure 8: (Next page, at right) A model solution for an aquifer thickness of 200 feet. The solution used a hydraulic conductivity of 10^{-2}, a TRPY of 10, and treated Milligan and Culverson Creeks as recharge wells rather than constant-head cells.

dozen cells went dry during this run. A model run with the same input values, but which treated Milligan and Culverson Creeks as recharge wells rather than constant-head cells was similar, although water levels were slightly higher.

Several model runs experimented with the shape of the aquifer. If the Greenbrier Group was treated as a single 1500-foot layer with a horizontal geometry, the solution appeared surprisingly similar to that of Figure 7, although the Spring Creek - Greenbrier River divide shifted slightly to the southeast in these cases. This seems to indicate that, in spite of the complex nature of the folding of the aquifer and its impermeable base, the hydraulic constraints (elevation of constant-head cells, locations of flow barriers) are more important in controlling the water flow. If the thickness of the aquifer is reduced while the complexly folded natural geometry is maintained, almost all of the aquifer flow becomes confined at a thickness of about 200 feet, and water levels rise too high to be realistic. A good solution was obtained in this case by increasing the TRPY to 10 (Figure 8). This solution may be the most realistic for approximating the potentiometric surface of the lowest member of the Greenbrier Group, the Hillsdale Limestone, which is confined by a thin shale.

In general, decreasing the anisotropy factor *or* the hydraulic conductivity had the direct effect of elevating the water table. By allowing the aquifer to drain less freely, this also resulted in maximizing the effects of the conduit recharge to the aquifer. Normally the low-flow cave-insurgence recharge had minimal effects on the model solutions unless the TRPY was less than 5 or the hydrualic conductivity less than 10^{-2} ft/min.

All model solutions that contained Spring Creek and the Greenbrier Rivers as constant-head cells resulted in a ground-water divide that coincided roughly with the one determined by dye tracing, and that varied little in position no matter how other factors were manipulated. This suggests that the location of this divide is a direct result of the elevation of these two major ground-water discharge zones. This hypothesis is supported by a few model runs

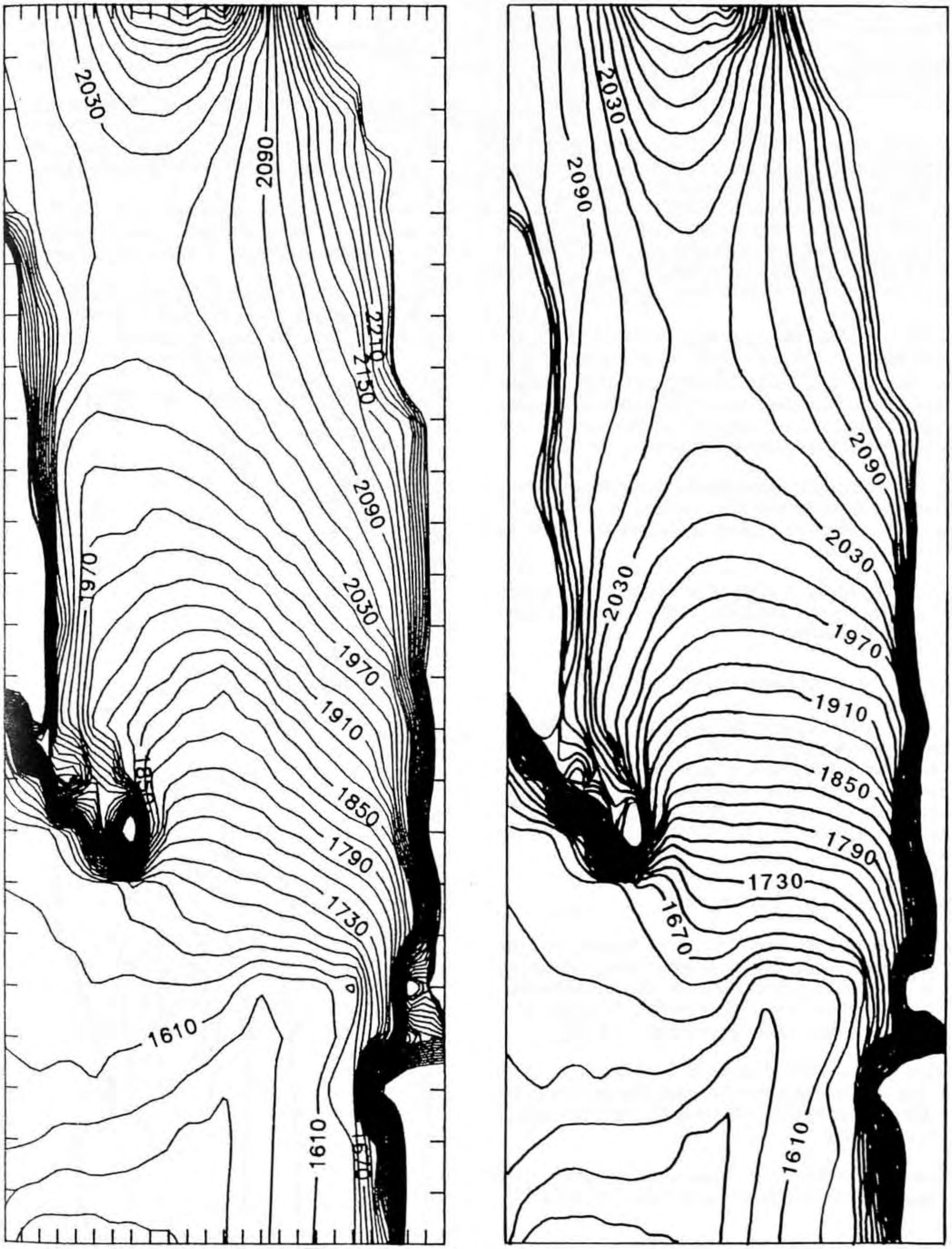
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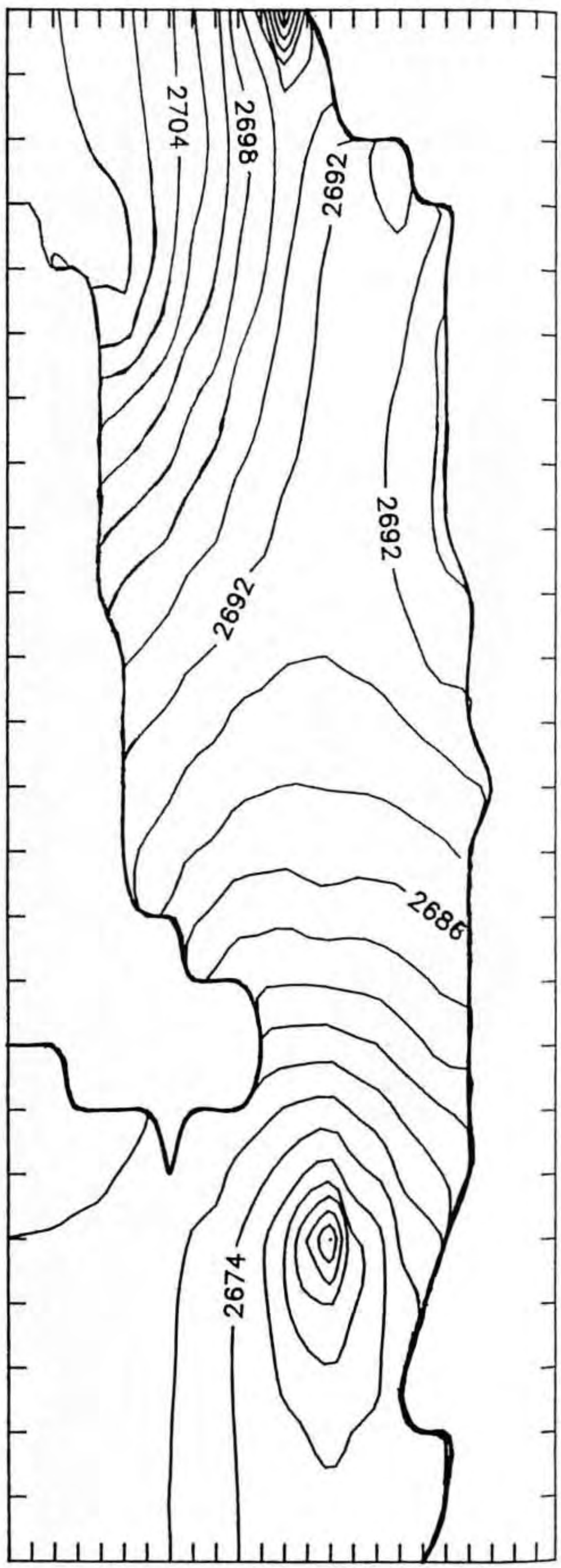
Figure 9: (At right) A model solution that used no constant-head cells at all, and that treated all springs or cave insurgences as recharge or pumping wells. The hydraulic conductivity used was 10^{-2}, the TRPY was 5, and the aquifer thickness was 1000 feet.



in which *all* constant-head cells were removed, and all known conduit springs were treated as pumping wells (Figure 9). The interesting result is a very flat water table (note that the contour interval is only two feet) with generalized flow from north to south. This solution is probably highly unrealistic, however, as there are likely to be many as yet unknown diffuse springs that discharge into the two major rivers, and that could not be simulated accurately.

Conclusions

1. The most important controlling factors in the solution for hydraulic head seem to be the hydraulic conductivity and anisotropy factor, followed by aquifer thickness as it affects confined versus unconfined flow.

2. The location of the ground-water divide between Spring Creek and the Greenbrier River seems to be a direct result of *only* the elevation of these discharge zones treated as constant-head cells.

3. The complexity in shape of the aquifer bottom, except as it affects the grid boundaries, seems to have little effect on the model solution.

Acknowledgments

The author extends gratitude to Alan Johnson of General Engineering Laboratories for his patient instruction and trouble-shooting services with both hardware and software. Appreciation is also extended to the College of Charleston Geology Department which provided the computer time, and to Mitchell Colgan for reviewing the manuscript.

References

Chisholm, J.L. and Frye, P.M., 1975, Records of wells, springs, chemical analyses of water, biological analyses of water, and standard streamflow data summaries from the Upper New River Basin in West Virginia: *U.S. Geological Survey Basic Data Report No. 4*, 80p.

Davies, W.E., 1958, Caverns of West Virginia: *West Virginia Geological and Economic Survey Volume 19*, 350 p. (Reprinted in 1965 with a 72-page supplement as volume 19A.)

Heath, R.C., 1983, Basic ground-water hydrology: *U.S. Geological Survey Water-Supply Paper 2220*, 84 p.

Heller, S.A., 1980, *A Hydrogeologic Study of the Greenbrier Limestone Karst of Central Greenbrier County, West Virginia*: Ph.D. dissertation (unpublished), West Virginia University, 167 p.

Jameson, R.A., 1985, *Structural Segments and the Analysis of Flow Paths in the North Canyon of Snedegar Cave, Friars Hole Cave System, West Virginia*: MS thesis (unpublished), West Virginia University, 421 p.

Jones, W.K., 1973, Hydrology of limestone karst in Greenbrier County, West Virginia: *West Virginia Geologic and Economic Survey Bulletin 36*, 49 p.

McDonald, M.G. and Harbaugh, A.W., 1984, A modular three-dimensional finite-difference ground-water flow model: *U.S. Geological Survey Techniques of Water-Resources Investigations*, Chapter A1, 528 p.

Ogden, A.E., 1976, *The Hydrogeology of the Central Monroe County Karst, West Virginia*: Ph.D. dissertation (unpublished), West Virginia University, 263 p.

Price, P.H. and Heck, E.T., 1939, Greenbrier County: *West Virginia Geological Survey County Report*, 846 p.

Reger, D.B., 1926, Mercer, Monroe, and Summers Counties: *West Virginia Geological Survey County Report*, 963 p.

Shuster, E.T. and White, W.B., 1971, Seasonal fluctuations in the chemistry of limestone springs: A possible means for characterizing carbonate aquifers: *Journal of Hydrology*, v. 14, p. 93-128.

Stevens, P.J., 1988, Caves of the Organ Cave Plateau, Greenbrier County, West Virginia: *West Virginia Speleological Survey Bulletin 9*, 200 p.

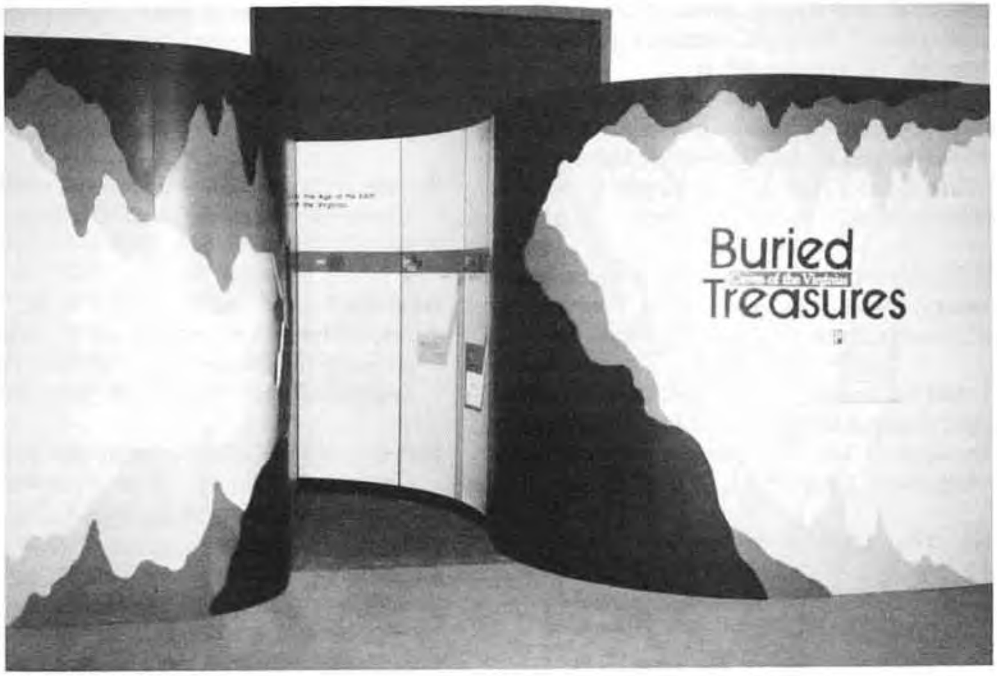


Plate I: Entrance to museum Exhibit, *Buried Treasures: Caves of the Virginias.* This traveling exhibit, designed by the Virginia Museum of Natural History, emphasizes the origin of, the biology of, and man's interaction with caves, including history, archeology, exploration, and environmental issues. The exhibit was funded, in part, by the Cave Conservancy of the Virginias and the Richmond Area Speleological Society. The premiere showing of the exhibit occurred at Radford University during the Appalachian Karst Symposium. *Photograph by Karen M. Kastning.*

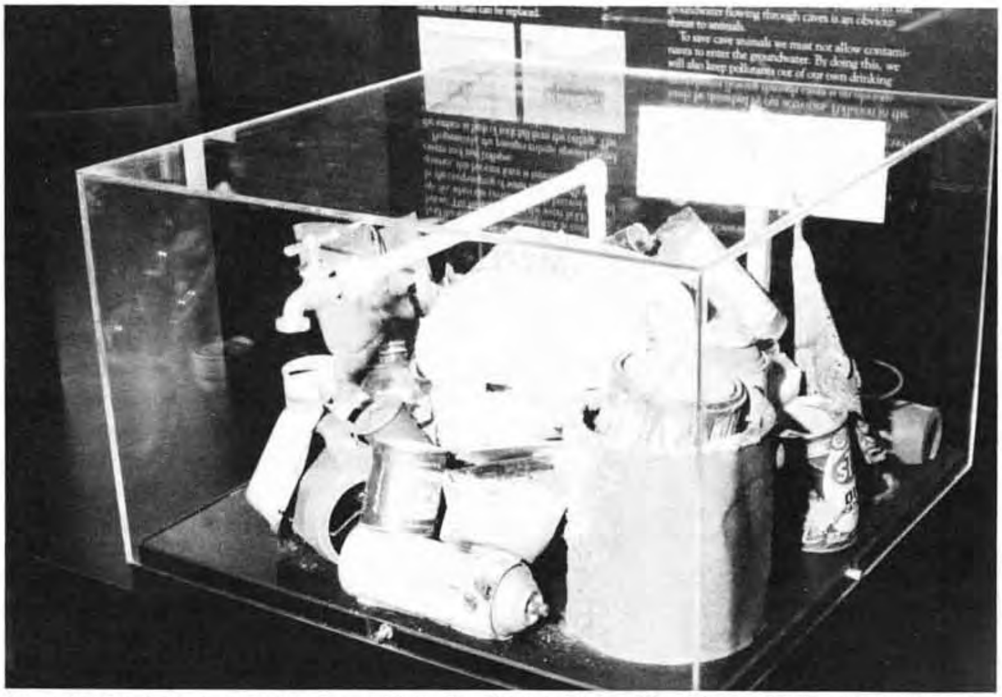


Plate J: Environmental display within the *Caves of the Virginias* exhibit shown above. Trash collected from several sinkholes within the Virginias has been arranged with a water spigot to emphasize that pollutants introduced in sinkholes can easily enter drinking water with little or no filtration. The sign asks the question, "Would you want to drink water that had flowed through this trash?" *Photograph by Karen M. Kastning.*

Hydrochemical Characteristics of the Greenbrier Limestone Karst of East-Central West Virginia

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ABSTRACT

Limestones of the middle Mississippian Greenbrier Group comprise only about 10% of the surface in southern Randolph and northern Pocahontas counties, but are very important hydrologically. The area is in the folded-plateau region, just west of the Valley and Ridge Province, with 300 m of structural relief and 700 m of topographic relief. Because clastic rocks overlying the limestones form most of the surface, there is relatively little karst-landform development on the surface, although caves are common, and streams usually sink during the drier parts of the year as they encounter the limestone. Much of the karst development occurs as underdrains of surface valleys, although in a number of cases interbasinal stream piracy does occur.

Water analyses, of about 250 samples from 50 locations, show considerable variability, even for sampling stations very close to each other. Statistical analysis places the samples into groups characteristic of surface streams, high-flow conduit springs, low-flow conduit springs, and diffuse-flow springs. Factor analysis shows that the chemical components group into carbonate-dissolution products, carbon-dioxide controlled parameters, and pollution constituents. Controlling hydrologic factors appear to be the type of recharge to the karst system that controls the carbon-dioxide related parameters, and the length of time and degree of contact of water with the rock within the system, controlling the dissolution products. Pollution constituents are not related to geological factors, but appear to be strictly anthropogenic.

Discharge measurements were seldom made during water sampling, but determinations of chemical denudation are possible in a few cases. Results range from a low of 1 mm/ka to a high of 20mm/ka of limestone removed by solution. Little correlation can be clearly demonstrated between calculated solutional denudation and any other parameter, although there appears to be an inverse relation between average gradient and calculated denudation rate.

Introduction

The karst on the eastern edge of the Appalachian Plateaus Province is developed on the Middle Mississippian Greenbrier Group. Unlike most karst areas, the northern half of the Greenbrier karst has relatively little surface exposure of limestone (Figure 1); however, the limestones underlie most of the area at depths shallow enough to be in the range of groundwater circulation. Therefore, the area is dominated by karstic hydrology and most of the water in the area has characteristics of karst water.

This paper summarizes hydrochemical aspects of the Greenbrier karst of northern Pocahontas and southern Randolph counties in eastern West Virginia. The physical hydrology of this same region was summarized in an earlier paper by Medville and Werner (1977). Most of the data used to develop the discussion of this paper are derived from investigations of the Edray Fish Hatchery springs in central Pocahontas County, and the Elk River drainage basin in northwestern Pocahontas County and southwestern Randolph County. Detailed reports of the hydrochemistry of these two areas are in preparation.

Study Area

The study area is located at the eastern edge of the Appalachian Plateau (Allegheny Plateau) Province. One of the main streams draining the area, the Greenbrier River, is generally considered to be the boundary between the plateaus and the Valley and Ridge Province to the east. Geologic structures underlying the area are broad, gentle folds that keep the dip of beds quite low, no more than a

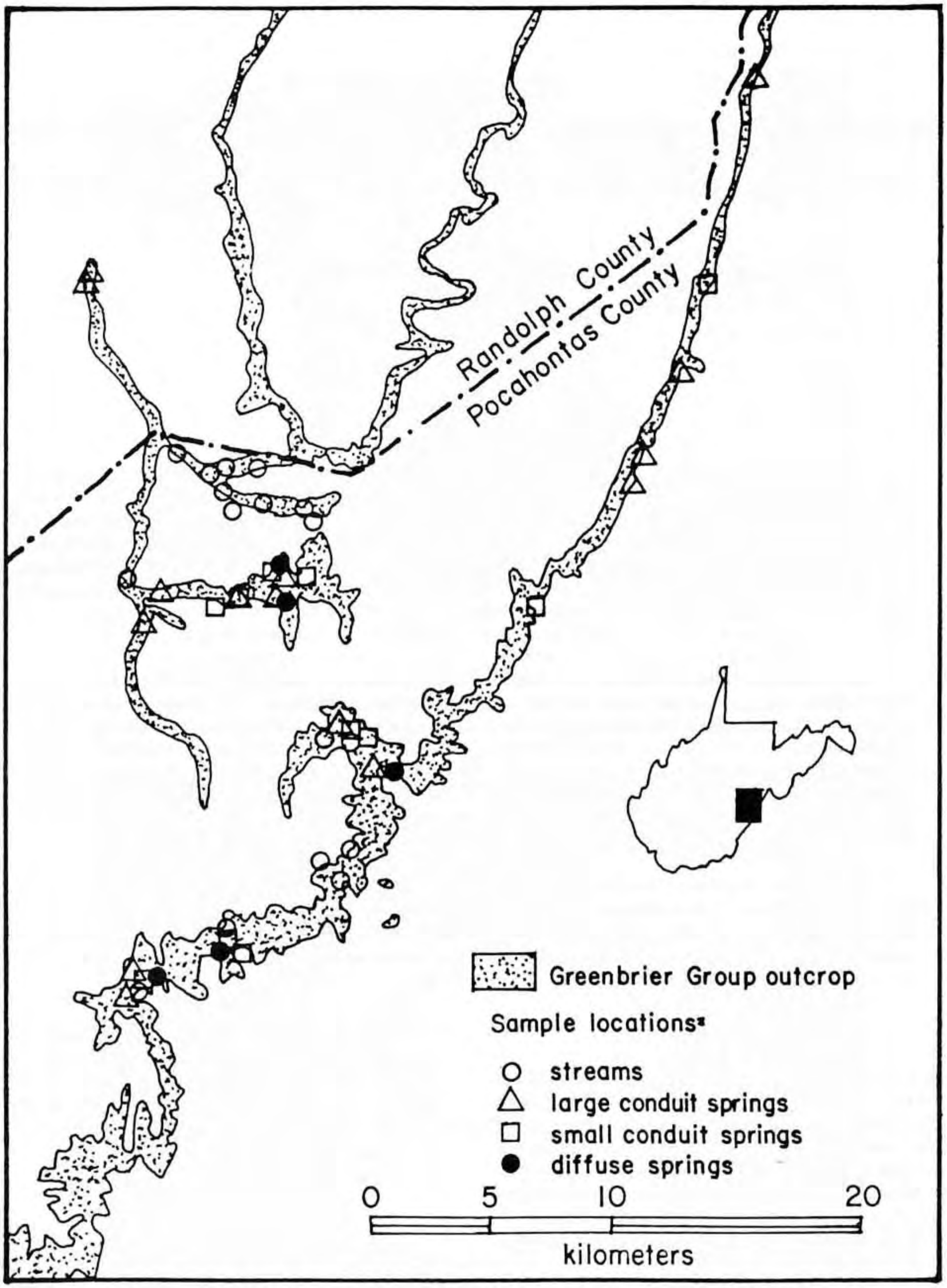

Figure 1: Location map of the study area. Not all sampling sites are shown in congested areas. Water types indicated for sample locations are based on field observations rather than the chemical character of the water. Outcrop extent of the Greenbrier Group taken from Cardwell, Erwin, and Woodward (1968).

few degrees in most of the area. Faulting is prevalent, but displacement on any single fault is generally measured in millimeters or centimeters (Eddy and Williamson, 1968; Werner, 1972), therefore none of these faults would appear on the typical geologic map.

The karst-forming rock in the area is the Middle Mississippian Greenbrier Group (Figure 2) which consists primarily of limestones with a few marker beds of shale or shaley limestone. Thickness varies from about 100 m at the northern end of the study area to 150 m at the southern end. The rocks underlying the Greenbrier are shales and sandstones, and can be disregarded in any discussion of karst development for this area. Overlying the Greenbrier are sequences of sedimentary rocks that are mostly non-marine clastic rocks, but repeated incursions of the sea, especially during the deposition of the first hundred meters of rock after Greenbrier time, have formed a number of thin, impure limestone beds. Much of the sandstone and shale in this sequence also contains calcite as cement or veins. Even the purely non-marine sequence, a considerable thickness above the last marine limestone, contains a number of fresh-water limestones. Thus, there is no sharp boundary between the carbonate and non-carbonate rocks in this area.

Precipitation averages 1200 mm/yr to 1500 mm/yr. Runoff determined from U.S. Geological Survey gauging stations in the area ranges from about 600 mm/yr to 900 mm/yr.

The Greenbrier River drains most of the eastern part of the study area, and the Elk River drains the western part. Interbasinal subterranean stream piracy through solution conduits is common, but, for the most part this is confined within the major river drainage basin (as far as is known). The only significant piracy between major surface drainage basins occurs in a tributary to the Tygart Valley River, north of the Elk River basin, into a solution conduit which eventually discharges into the Elk River basin (Medville and Werner, 1977).

The Greenbrier limestone terrane has undergone karstification throughout, but in the northern part of its extent, including the study area, there are few sinkholes. This is because limestone comprises less than 20% of the surface outcrop. South of the study area, in Greenbrier and Monroe counties, surface exposures of limestone are proportionally greater and sinkhole development is intense.

Methods.

Approximately 260 water samples were collected from nearly 80 locations (springs, and surface and cave streams) during a two-year period from October 1973 to October 1975. Chemical analyses for constituents important in carbonate dissolution were performed on 230 of these samples, and several derived values were computed. These data were used to develop the discussion in this paper. Table 1

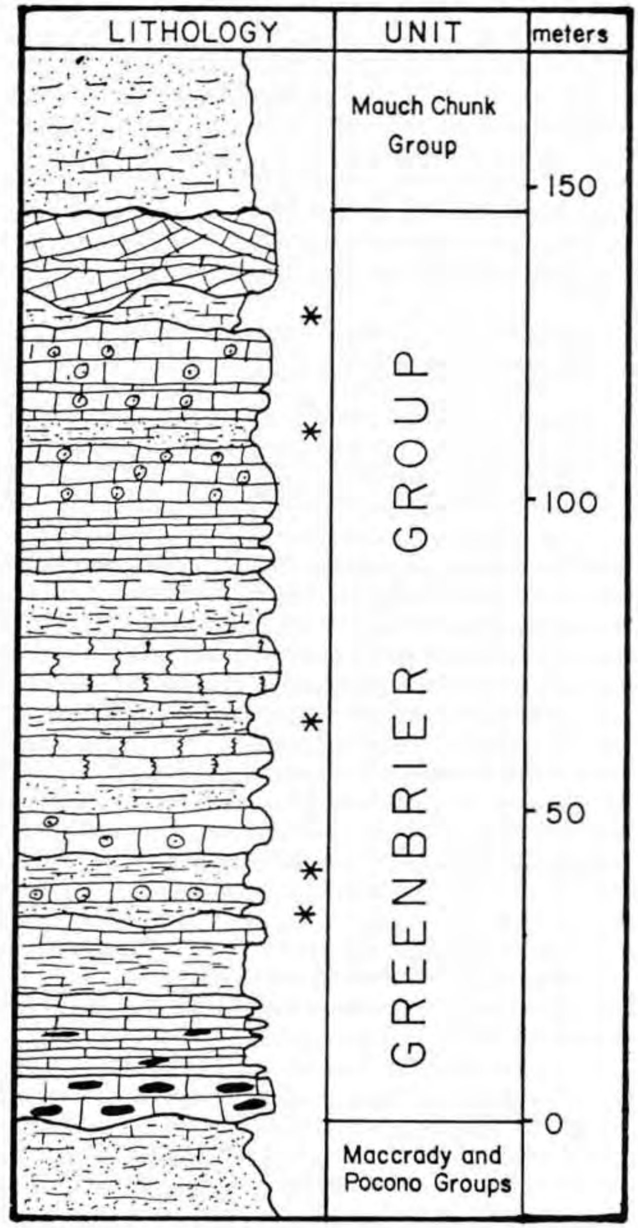

Figure 2: Stratigraphic column for the study area. Lithologies and thickness are representative of the center of the study area. The section thins to the north, and thickens to the south, primarily through corresponding changes in the lower units of the Greenbrier Group. Hydrologically important aquitards and aquicludes are indicated by asterisks.

gives the descriptive statistics associated with these data. Most locations were sampled one to three times, but several locations were sampled more often. In particular, the springs associated with the Edray State Fish Hatchery were sampled about 20 times each, and the streams and springs in the upper Elk River drainage basin were sampled five or more times.

Temperature and specific conductance were measured in the field. Although pH was usually measured with a

Variable	median	range	1st decile	9th decile
pH	7.76	6.87/8.35	7.37	8.08
Ca^{++}	34.0	6.9/92.9	15.2	63.1
Mg^{++}	2.62	0.49/12.44	1.46	4.52
HCO_3^-	102.5	16.0/298.5	44.3	185.1
Cl^-	5.6	0.5/67.6	1.6	18.4
NO_3^-	3.3	0.2/13.7	0.4	6.1
SIc	-0.19	-2.11/1.15	-1.22	0.53
SId	-0.66	-2.48/0.67	-1.64	0.07
pCO_2	-2.80	-3.63/-1.94	-3.16	-2.48
Ca^{++}/Mg^{++}	7.2	2.5/52.3	5.0	10.1

Table 1. Descriptive statistics of water-sample analyses.

portable pH meter in the field (at the vehicle), generally a few minutes after collection, some collected samples were not measured until returned to the laboratory three or four hours after collection. The portable instrument was very unstable at the sampling locations, probably because of high humidity, so attempts at *in situ* pH measurements were abandoned. A later set of measurements with a more stable meter at some of the springs showed that pH varied by 0.1 units over periods of a few minutes, and that measurements made up to 24 hours later on a sample collected in a plastic bottle and kept cool agreed with those made in the field to within less than 0.1 units.

Analysis for bicarbonate was based on titration for alkalinity with hydrochloric acid to the inflection point near pH of 4.5. Calcium and magnesium determinations were by EDTA titration.

The remaining analyses were not performed on all samples. Chloride was determined for 225 samples by titration with mercuric nitrate using diphenylcarbazone indicator, and nitrate was determined colorimetrically for 150 samples by the cadmium reduction method. Due to problems in applying the analytical techniques and the relatively low levels of the remaining ions, none of these were incorporated into the data set for the statistical analysis. Sulfate analysis by the turbimetric method was performed on approximately 20% of the samples, and never exceeded 10 mg/l, which was deemed the reliable resolution limit for the technique. On the 10% of samples tested for sodium and potassium by atomic absorption spectroscopy, no sample exceeded the detection limit of the instrument for potassium (approximately 5 mg/l), and sodium was below 5 mg/l for all samples except a few with high chloride levels (but still below 10 mg/l).

Values were calculated for the following from the chemical-analysis data: saturation indices with respect to calcite and dolomite, initial equilibrium carbon-dioxide concentration, and calcium/magnesium (molar) ratios. Chemical-equilibium constants for these calculations were obtained from a number of sources (Harned and Owen, 1958; Plummer and Busenberg, 1982; Langmuir, 1971).

General Character of the Waters

The chemical character of all the karst waters analyzed indicated that they were relatively nonpolluted. The principal dissolved components were those derived from limestone solution, with virtually no influence from other sources. A significant although relatively minor exception was the addition of chloride from road-deicing salts in the group of springs at the Edray Fish Hatchery (discussed previously in Werner, 1977). There has been little effect from fertilizers or septic tanks as indicated by low dissolved-nitrate levels, nor is acid-mine drainage involved as seen from low dissolved sulfate levels. Although gypsum occurs in some areas in the Greenbrier limestones, none has been recorded in the immediate area, and low sulfate levels reinforce this. Thus, the data set used in this analysis would appear to represent as pure a set of limestone karst waters as possible, without complicating factors.

Spatial and Temporal Variations

Although the water under discussion evolves largely in contact with the Greenbrier limestones which have significant chemical and physical variation, the variations seen in the chemical characteristics of the water do not correspond to limestone characteristics. Field observations of the springs under discussion, including traversing some of the caves through which the water flows before emerging and dye tracing to delineate some of the recharge-discharge connections, has provided information for a general interpretation of the spring types found in the area.

Various workers (Shuster and White, 1971; Harmon and others, 1972, 1973; Drake and Harmon, 1973) have classified karst waters on the basis of their positional properties with regard to the aquifer, and have attempted to relate this classification in one way or another to the chemical character of the waters. In a very general way, all waters in most karst terranes are mixtures of two end-member types, surface stream recharge and diffusing soil waters, that have passed through sequential steps of chemical evolution. Surface waters are more or less in equilibrium with atmospheric carbon dioxide, and soil waters are at equilibrium with soil carbon dioxide that may be about two orders of magnitude higher. Drake and Harmon (1973) arrived at a classification comprising six water types - allogenic surface recharge, soil-zone recharge, conduit springs, diffuse springs, wells, and surface discharge - that could be distinguished chemically on the basis of saturation with respect to calcite and equilibrium carbon-dioxide partial pressure alone.

Considering basic principles with regard to groundwater, waters of a given terrane progress through several stages as they pass through the aquifer (Figure 3); there is a source recharging the aquifer, water passes through stor-

age in the aquifer to a discharge point, and then leaves the terrane. Applying this model to the real world can lead to some difficulties. For example, field observations allow one to easily determine that water is obtained from a stream, but determining whether it is hydrologically above or below karst terrane requires additional information. Although that determination is not usually difficult, an additional complication exists. The diagram of Figure 3 seems to clearly distinguish the two types of streams, entering and leaving, but it is possible that a stream entering one karst terrane has just left another. In the case of the Greenbrier karst, lithologies are such that the rock sequence is divided into several hydrologic units, each of which would then present one cycle of the model presented in Figure 3. Furthermore, although not shown on the diagram, some streams flow on the surface across a karst terrane, being chemically modified by dissolving limestone from their beds. Thus streams are not always easily classified, other than as streams.

Somewhat lesser problems exist with spring waters. Shuster and White (1971) classified springs in the Nittany Valley of Pennsylvania on the basis of seasonal variation in chemistry as either conduit or diffuse springs. Since then, this classification has been applied in its original or slightly modified form to other areas. Although it is usually not difficult to determine the physical classification of a spring when adequate observational data are available, this chemical classification of the spring water may be more difficult because of the complex path this water may have taken through the various stations denoted on Figure 3, and the attendant complications in the chemical evolution of the water. In the area under discussion in this study, springs are classified as being conduit (cave) springs of two types, "large" and "small", and diffuse springs.

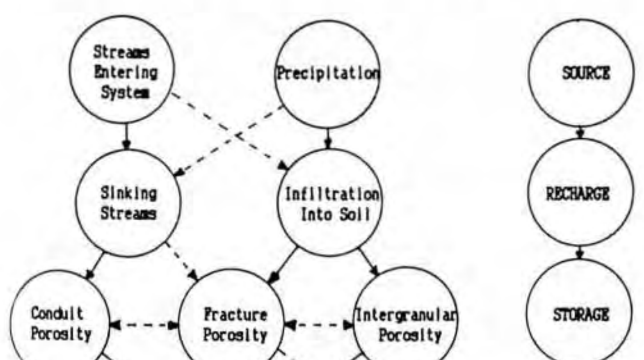

Figure 3: Water routing through a karst aquifer. The vertical column on the right is the fundamental sequence of water passing through any aquifer. The left part of the diagram is more specific to the karst aquifer with the various forms each segment of the routing sequence may take. Prominent (and most likely) paths are shown as solid lines and less likely or low mass-transfer paths are dashed lines. Only natural karst paths are shown; some streams simply flow across, and wells may abstract water from the aquifer; neither are shown.

Differentiation between small- and large-conduit springs is not based entirely on discharge, although in general, "large" springs have higher discharges than do "small" ones. The major distinctions are gradient steepness and character of the conduit. "Large cave springs" flow from relatively steep conduits with little pooling of water. The cave passage behind such a spring tends to hold a wall-to-wall stream flowing throughout. This results in faster flow-through rates and consequently less time for dissolving of rock and coming to chemical equilibrium. Conduit systems have less storage capacity and springs are more flashy than the other types, and therefore have greater coefficients of variation in all characteristics. Reviewing data presented by the workers cited above indicates that various properties of these springs fall outside the ranges reported by them, so perhaps large cave springs do not exist in their study areas.

"Small cave springs" flow from conduits that are usually less steep and contain pools. This produces a longer flow-through time for systems that have the same sink-to-rise distance, and consequently the water has more time to dissolve rock and come to equilibrium. These systems are also less flashy in discharge and chemical character. The principal source of water for both types of conduit springs is point recharge from sinking streams into solution conduits, but there is a contribution from seepage water infiltrating through soil and entering through fractures in the ceilings of conduits. Because of the differences in flow-through time and overall flow volume, small cave springs tend to be more affected by the contribution of seepage water and show that as higher content of dissolved solids and higher equilibrium carbon-dioxide partial pressures. These springs seem to correspond to all conduit springs of the other workers.

Diffuse springs derive all or nearly all of their water from surface water infiltrating through soil cover. Flow paths are tortuous and constricted and therefore discharge tends to be limited more by hydraulic properties of the fractures or pores than by availability of water at the infiltration source. Flow-through times tend to be very long, for the length of the paths, and as a result, these waters reach saturation with respect to the rocks they contact.

Temporal variation in the spring types is in agreement with the findings of Shuster and White (1971) and may be seen in Figure 4. Differences in seasonal fluctuations among springs is pronounced. Figure 4 shows coefficients of variation for pH and bicarbonate content that approxi-

mately double between diffuse springs and small-conduit springs, and again between small- and large-conduit springs, for the two year period. Although these example springs have more data available than do others in this study, comparable coefficients of variation were obtained for other cases in which multiple samplings were done.

The sampling locations were classified based on field observations, prior to an analysis of the chemical data. These classifications are the ones used in Figures 1 and 5. It is noteworthy that, in several places within the study area, all three types of springs appear within a few tens or hundreds of meters of each other, rising from the same stratigraphic horizon. If a classification based on chemical properties of the waters is considered against a classification based on field observations, similar groupings occur. This can be illustrated by plotting variables against each other as in Figure 5. Paper plots allow for only two variables; perfect class groupings are expected if only two variables were required for classification, such as in the data set of Drake and Harmon (1973). As can be seen in Figure 5, clearly more than the two variables used are required for complete separation of groups; however, despite considerable overlap, the general trends of variation among the water classes are readily apparent. Stream samples are low in dissolved solids and carbon dioxide; large-conduit springs reflect higher carbon-dioxide levels within cave passages through which they pass (although dissolved solids do not appreciably increase probably because of very fast flow-through times). Small-conduit springs, because of longer residence times, increase in dissolved-solids content, but are in equilibrium with essentially the same cave atmosphere as the large-conduit springs. Diffuse springs have contributions from some higher carbon-dioxide levels, and longer residence times, as well as more intimate contact with the rocks, allow for higher dissolved-solids concentrations.

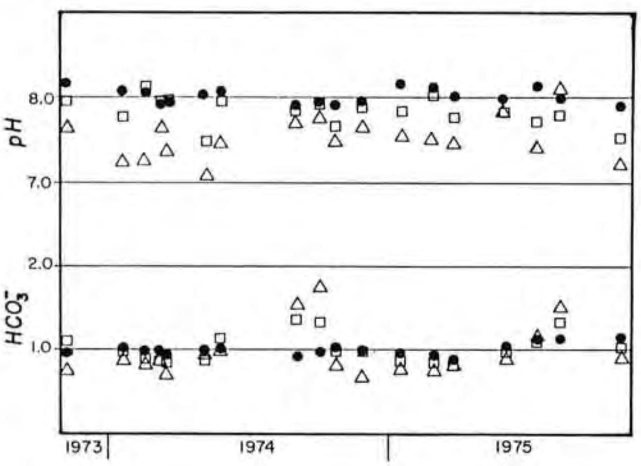

Figure 4: An example of temporal variation for the three spring types. The springs are three of the springs at the Edray Fish Hatchery. Bicarbonate is related to the mean of all values for the particular spring so as to show variation over time. Means for the springs are 60 mg/l for the large conduit spring (triangles) with coefficient of variation of 30%, 120 mg/l for the small conduit spring (squares) with coefficient of variation of 18%, and 195 mg/l for the diffuse spring (filled circles) with coefficient of variation of 7%.

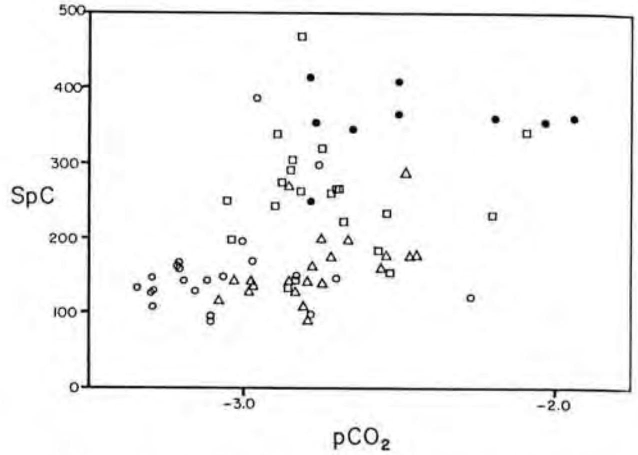

Figure 5: Cross plot of specific conductance (as representing total dissolved solids) against log equilibrium partial pressure of carbon dioxide for all sampling locations in this study. Where more than one sample was taken at a location, the mean values were used in this figure. Classification of locations is based on field evidence. Open circles are stream samples, triangles are large-conduit springs, squares are small-conduit springs, and filled circles are diffuse springs.

Statistical Analyses

A variety of statistical tests were applied to the chemical data collected. In general, the results of these tests confirmed the hypotheses generated from field observations and general expectations based on the common understanding of karst hydrochemistry.

Correlation Analysis.

Pearson's Product Moment Correlation coefficients were generated for the chemical variables and two generated time variables (Table 2). One time variable (WD) is the day of the water year, that is, 1 October is set equal to 1, 1 November becomes 32, etc.; the other (DD) is number of days away from 1 April (middle of the water year), that is, both 31 March and 2 April are set to 1, both 1 May and 2 March are equal to 30, etc. These variables were created to test the basis of commonly held wisdom with regard to karst-water chemistry. Discharge is expected to be at minimum at the beginning (and end) and maximum at the middle of the water year. Most dissolved constituents are expected to hold an inverse relationship to discharge; if so, they should correlate with DD. Conversely, dissolved mineral constituents have often been shown to peak with an increase in discharge and then decline, though not in phase with the discharge. In this case, the WD variable

Table 2. Pearson's Product Moment Correlation Coefficients of all water data.

	pH								
T	0.20*	**T**							
SpC	0.49**	0.24*	**SpC**						
Ca^{++}	0.44**	0.31**	0.94**	**Ca^{++}**					
Mg^{++}	0.41**	0.24*	0.77**	0.75**	**Mg^{++}**				
HCO_3^-	0.45**	0.30**	0.96**	0.94**	0.80**	**HCO_3^-**			
Cl^-	0.32**	0.08-	0.64**	0.43**	0.32**	0.38**	**Cl^-**		
NO_3^-	0.27*	-0.11-	0.39**	0.32**	0.23+	0.35**	0.41**	**NO_3^-**	
WD	-0.20*	-0.18+	-0.21*	-0.30**	-0.34**	-0.21*	-0.14-	-0.13-	**WD**
DD	-0.17+	0.57**	0.21*	0.22*	0.23*	0.25**	0.07-	-0.18+	-0.03-

Notes: Statistical significance of each coefficient is designated as follows:

** = alpha <0.001

* = 0.001<alpha<0.01

+ = 0.01<alpha<0.05

- = alpha>0.05

might be a better representation of a time variable expected to relate to the changes in concentration.

As expected for karst waters, the highest correlation was between bicarbonate and specific conductance (0.96); correlation between calcium and bicarbonate (0.94) and calcium and specific conductance (0.94) was nearly as high.

Because the number of observations is fairly large, several other, weaker correlations are also statistically significant at an alpha level of 0.0001. It is less certain that a cause-and-effect relationship is present, despite the apparent high significance level, but relationships indicated by at least some of these correlation coefficients can be rationalized. For example, pH correlates positively with all the limestone dissolution variables (specific conductance and concentrations of calcium, magnesium, and bicarbonate) because as solution of limestone progresses, the dissolved constituents obviously increase and that reaction serves to increase the pH.

Somewhat surprising is the general lack of correlation between the time variables and chemical variables. The best correlation is between temperature and the DD variable, at 0.57, indicating higher temperatures at the beginning and end of the water year than in the middle, which is hardly surprising. The relatively low correlation is most likely because most of the samples are spring samples that tend to vary relatively little in temperature. Both calcium and magnesium are negatively correlated with position in the water year (WD), and slightly less positively correlated (at a lower significance level) with DD, perhaps indicating a relationship between these dissolved constituents and discharge that is out of phase, or, as noted above, indicating flushing of the more concentrated waters during the initial higher-discharge period and not a strong inverse relationship with discharge.

Discriminant Analysis.

Discriminant analysis is normally used in order to group multivariate observations into classes. It is necessary to determine what the classes are to be on some basis, and so designate observations included in a training set. The procedure operates by assuming an equal probability of each observation falling into each of the designated classes, and then developing a mathematical model (equation) that would calculate a high probability, ideally 1.0, that each observation actually falls into the group that has been predefined for each observation. In practice, for most real data sets, not all probabilities reach 1.0, nor is each observation in the training set actually assigned to the predefined class. The procedure attempts the best classification possible, given the data supplied. After the model is developed, it can then be applied to other data not part of the training set to classify those observations according to the same criteria.

For this project, observations on the water sources allowed a classification into four basic water types - streams, large cave springs, small cave springs, and diffuse springs. The samples taken at the Edray Fish Hatchery and at several surficial stream sites provided a sizable, easily classified subset of data. In the case of this data set, 79 observations were predefined as belonging to one of the four water types and used to train the discriminant program. Designated important variables for this run were pH, temperature, concentrations of calcium, magnesium, bicarbonate, chloride, and nitrate ions, saturation indices with respect to calcite and dolomite, and equilibrium carbon-dioxide pressure. The discriminant equations were then applied to the training set and also to the mean values of the same parameters for all sampling stations.

The results of this analysis were somewhat surprising in how well the procedure performed to classify both the observations included in the training set and the means of observations for the remaining data. The observations classed as streams were principally surface streams flowing across the limestone outcrop, but included those that had just entered the limestone outcrop as well as those that had traversed the entire outcrop. Also included were two

observations from cave streams taken at locations in the cave where the stream had not been underground for a long distance. For these "stream" samples, all classified as "stream" at a probability of 1.0 except for one that came from a cave stream at a probability of 0.97. "Large cave springs" are those with relatively high discharge and a fairly high gradient (about 0.06). These are the main springs of the Edray Fish Hatchery and their recharge is from stream sinks approximately one mile (air line distance) from the springs (Medville and Medville, 1976). Intuitively, one might expect some problem in differentiating waters from such springs from the stream waters. However, only one sample out of 19 used in the training set as "large cave spring" (and, in fact, the only sample in the entire set) was misclassified (as a stream). About a third, however, rated a probability less than 1.0 (although always 0.8 or higher). "Small cave springs" have a lesser discharge, and a lower gradient (about 0.03), and are known to issue from a cave passage (Salmon Cave) that has a considerable number of pools (Medville and Medville, 1976). All samples classified correctly; only one had a probability of less than 1.0 (at 0.98). The last group, "diffuse springs" was based on samples from a single tufa-depositing spring at the Edray site. All of these samples classified correctly with a probability of 1.0.

A point of interest in regard to the spring samples used in this set is that they are located within 1 km of each other, and all issue from the same stratigraphic horizon near the base of the Greenbrier Group. The chemistry is sufficiently distinct among these springs that they can be grouped into one of the three spring types simply on the basis of the chemical analysis of a single sample. Unlike the set of springs studied by Shuster and White (1971) that required a temporal factor for classification, the Edray springs apparently are sufficiently different from each other to not require knowledge of that factor.

Less accurate was the attempt to classify the remaining locations. The means of all samples taken at each location were classified against the discrimination model. Only 57% of these were classified the same as a classification based on field observations. However, most disagreements involve locations that are not very well known, and only 14% of the locations were clearly misclassified, probably resulting from too few or anomalous samples for the location.

The experience here is quite different from that of Drake and Harmon (1973). They found that only one variable, saturation with respect to calcite, was sufficient to separate all their classes (with the exception of the spring classes), and the addition of log carbon-dioxide equilibrium pressure was sufficient to completely separate their six classes. In the present study, no single one or even two variables provided a very good separation of the four defined water classes. One of the best two-variable determinations is between specific conductance and log equilibrium carbon-dioxide pressure (Figure 5). Although some sense of discrimination is apparent, there is still considerable overlap. A significant difference exists between the terranes used for the two studies. Drake and Harmon (1973) obtained their water samples from areas in the Valley and Ridge of Pennsylvania, where the carbonates are all in one sequence, overlain by clastic rocks that contain almost no carbonate, even as cement. On the other hand, the Greenbrier Group of West Virginia is overlain by a mostly clastic sequence that contains a number of limestone beds as well as calcite-cemented sandstones. From a hydrochemical standpoint, then, water reaching the Greenbrier limestones has probably already passed through a carbonate aquifer, however thin or impure it might be. Even though a few of the stream samples had quite low levels of calcium or bicarbonate, none were as low as the mean of Drake and Harmon's stream samples. Thus, because of multiple cycles of the routing shown in Figure 3, waters from the Greenbrier terrane have had a more complex evolution than the central Pennsylvania waters.

Factor Analysis.

Factor analysis was done using analytical and computed variables on samples derived from springs at the Edray Fish Hatchery; stream samples were excluded. Each data group - large cave springs, small cave springs, and diffuse springs - was run separately. In general, four factors appeared (Table 3). The pollution variables - chloride and nitrate - appeared as separate factors, which is not surprising because the sources are so different. However, the "karst water" variables separated into two distinct groups. The results of rock solution - calcium, magnesium, bicarbonate, and the saturation indices - were associated on one

	Spring group		
	Large conduit	Small conduit	Diffuse
FACTOR 1	carbonate solution 40%	carbonate solution 31%	carbon dioxide 29%
FACTOR 2	carbon dioxide 21%	carbon dioxide 31%	carbonate solution 26%
FACTOR 3	chloride 14%	chloride 16%	chloride 14%
FACTOR 4	nitrate 13%	nitrate 12%	nitrate 12%
Total variation explained	88%	90%	81%

Table 3. Results of factor analysis on the springs of the Edray Fish Hatchery area. Factor names are generalized from the variables actually composing them. Percentages are variation explained by the factor in the data set for that spring group.

factor, and pH and equilibrium carbon-dioxide pressure associated on another factor. As noted above under correlation analysis, pH did not correlate highly with the rock solution factors, indicating that there was independence of the variables. The conclusion to be drawn from the statistical analysis is that the final pH of the spring water is more controlled by the partial pressure of carbon dioxide of the hydrologic system than by the amount of rock solution. This follows from the chemical pathway followed in the evolution of karst water (*see* White, 1988, p. 205); given an open system with water nearing saturation and the units in which the analytical values are measured, the pH change is small relative to the amount of change in the bicarbonate (and calcium or magnesium) ion concentration.

Ogden (1976) performed factor analysis on his data set of water analyses from springs and wells in an area in Monroe County, West Virginia, approximately 70 km to the south and also underlain by carbonates of the Greenbrier Group. Although different in detail, his results were largely the same with the exception of one additional factor representing the magnesium-dolomite component. Ca^{++}/Mg^{++} values for his water samples were lower although the reason for this is not clear. His area is closer to the Cambrian-Ordovician carbonates and there may be some movement of water or weathered rock from that terrane to the Greenbrier limestone terrane. The Greenbrier limestone itself has similar Ca^{++}/Mg^{++} ratios in both areas according to analyses given in McCue, Lucke, and Woodward (1939).

Denudation Rates

Although only a few discharge measurements were made at the time of water sampling, it is possible to determine instantaneous carbonate flux rates (based on the dissolved calcium plus magnesium) for a number of locations and times. For sixteen of the springs, low-flow measurements were made on one instance each. Several discharge measurements were made of the spring on the Elk River at the downstream end of the limestone outcrop, and it was possible to estimate flow for other sampling times from gauge records of the U.S. Geological Survey kept at Webster Springs. Discharge measurements at the Edray Hatchery were difficult because water abstraction systems are installed at the spring orifices. However, it was possible to obtain some direct discharge estimates and to relate these to records from a nearby U.S. Geological Survey gauge. Because of the uncertainties inherent in estimates of this type, the results are offered here as rough estimates.

Instantaneous flux rates for the low-flow measurements on Back Allegheny Mountain (the northeastern part of the outcrop on Figure 1) ranged from a low of 3.4 mm/ka (millimeters over the entire drainage basin per thousand years) to 15.4 mm/ka; in the Edray area (the southwestern part of Figure 1), rates ranged from 1.4 mm/ka to 7.3 mm/ka; and for the Elk River basin (the western part of Figure 1), the range was from 0.9 mm/ka to 3.0 mm/ka. One high-flow measurement for the downstream end of the Elk River basin provided a rate of 10.2 mm/ka.

Estimated annual average carbonate flux rates could be calculated for two basins in the study area. The U.S. Geological Survey has a recording gauge in the Elk River basin not far downstream from the karst area and one in Indian Draft below the karst terrane. Estimates of total runoff are available for these two basins. When a weighted average of water analyses from this study and the average runoff are used to calculate carbonate-denudation rates, the results are 15.5 mm/ka for the Elk River basin, and 16.1 mm/ka for Indian Draft.

These denudation rates are somewhat low compared to rates reported by other workers. Ogden (1982), working to the south of this area, reports rates of 19.0 mm/ka to 22.6 mm/ka. The basin areas are of comparable size, but there is a much more extensive outcrop and generally gentler relief in his area. The spring waters in his basins contain appreciably higher dissolved calcium and magnesium concentrations and are more saturated (Ogden, 1976), apparently reflecting longer residence times in the karst.

Compared to other data summarized by White (1988, p. 218), all the rates derived in this study are very low. The average runoff reported from U.S. Geological Survey stations for this area is 600 mm/yr to 900 mm/yr. Applying the regression equation of White results in denudation rates of 35.8 mm/ka to 50.5 mm/ka, or two to three times those derived from the chemical data. Analysis of the conditions prevailing in the study area relative to the "average" karst indicates that residence time in the limestone is much shorter here, mainly because of higher gradients. However, this may not be valid because there seems to be a trend toward higher carbonate flux rates at low flow for those springs in higher gradient basins within the study area. That, combined with a lack of extensive outcrop accessible to water, may explain the low chemical-denudation rates for this terrane.

The only direct measurement of erosion in a karst channel in the Greenbrier karst is that reported by Coward (1975) from a cave about 20 km to the south. He measured approximately 0.75 mm of erosion per year in an active stream channel. If that rate is compared to the overall denudation rate as calculated from the water chemistry, the implication is that approximately 2% of the surface is being actively eroded. If all of this were occurring in cave channels, then one would expect an average meter-wide cave passage approximately every kilometer (considering allowance for wetting of the walls and floor). Of course, it is unlikely that all dissolution is in conduits, or in traversable cave passage, so that estimate becomes the upper limit of actively-forming-cave frequency, with the actual frequency considerably lower. It also does not consider formerly active, but now abandoned passages.

Summary

The Greenbrier Group karst terrane of northern Pocahontas and southern Randolph counties of West Virginia has some unique characteristics in regard to its waters. In comparison to most other well studied areas, this karst area is characterized by high hydraulic gradients that do not allow the waters to dissolve as much limestone because of the shorter average-residence time. Therefore, the average chemical-denudation rates of 15 mm/ka to 16 mm/ka are two to three times lower than average for karst terranes in general.

There are three basic spring types to be found in this area; these can be classified on the basis of physical and chemical properties into large conduit, small conduit, and diffuse springs. The diffuse springs correspond well to diffuse springs elsewhere, and the small conduit springs appear to correspond to conduit springs studied in the karst of Pennsylvania. The large conduit springs do not correspond to any of the springs reported by the Pennsylvania workers, and are probably a consequence of high gradients in well developed conduits that are not common in the Valley and Ridge of Pennsylvania.

Multiple cycling through a basic karst-aquifer route for most of the waters in this terrane creates a more complex evolution of the waters than in a terrane with more distinct carbonate-non-carbonate rock sequences such as in the Valley and Ridge or some of the karst in the continental interior.

References

Cardwell, D.H.; Erwin, R.B.; and Woodward, H.P., 1968, *Geologic Map of West Virginia*: West Virginia Geological Survey, 2 sheets.

Coward, J.H.C., 1975, *Paleohydrology and Streamflow Simulation of Three Karst Basins in Southeastern West Virginia, U.S.A.*: Ph.D. dissertation (unpublished), McMaster University, 303 p.

Drake, J.L. and Harmon, R.H., 1973, Hydrochemical environments of carbonate terrains: *Water Resources Research*, v. 9, p. 949-957.

Eddy, G.E. and Williamson, D.B., 1968, The effect of faulting in Cassell Cave, West Virginia (abstract): *National Speleological Society Bulletin*, v. 30, p. 38.

Harmon, R.S.; Drake, J.J.; Hess, J.W.; Jacobson, R.L.; Ford, D.C.; White, W.B.; Fish, J.E.; Coward, J.H.C.; Ewers, R.O.; and Quinlan, J.F., 1973, Geochemistry of karst waters in North America: *Proceedings of the 6th International Congress of Speleology, Olomouc*, v. 3, p. 103-114.

Harmon, R.S.; Hess, J.W.; Jacobson, R.W.; Shuster, E.T.; Haygood, C.; and White, W.B., 1972, Chemistry of carbonate denudation in North America. *Transactions of the Cave Research Group of Great Britain*, v. 14, p. 96-103.

Harned, H.S. and Owen, B.B., 1958, *The Physical Chemistry of Electrolytic Solution*: Reinhold, New York.

Langmuir, D., 1971, The geochemistry of some carbonate ground waters in central Pennsylvania: *Geochimica et Cosmochimica Acta*, v. 35, p. 1023-1045.

McCue, J.B.; Lucke, J.B.; and Woodward, H.P., 1939, Limestones of West Virginia: *West Virginia Geological Survey, Volume 12*, 560 p.

Medville, D.M. and Medville, H.E., 1976, Caves and karst hydrology in northern Pocahontas County: *West Virginia Speleological Survey Bulletin 6*, 174 p.

Medville, D.M. and Werner, E., 1977, Karst hydrology and water chemistry in a mixed sedimentary terrain, *in* Tolson, J.S. and Doyle, F.L. (editors), *Karst Hydrogeology - Memoirs of the 12th Congress of the International Association of Hydrogeologists*: The University of Alabama in Huntsville Press, Huntsville, Alabama, p. 443-457.

Ogden, A.E., 1976, *The Hydrogeology of the Central Monroe County Karst, West Virginia*: Ph.D. dissertation (unpublished), Department of Geology and Geography, West Virginia University, Morgantown, 262 p.

Ogden, A.E., 1982, Karst denudation rates for selected spring basins in West Virginia: *National Speleological Society Bulletin*, v. 44, p. 6-10.

Plummer, L.N. and Busenberg, E., 1982, The solubilities of calcite, aragonite, and vaterite in CO_2-H_2O solutions between 0° and 90° C, and an evaluation of the aqueous model for the system $CaCO_3$-CO_2-H_2O: *Geochimica et Cosmochimica Acta*, v. 46, p. 1011-1040.

Shuster, E.T. and White, W.B., 1971, Seasonal fluctuations in the chemistry of limestone springs: A possible means for characterizing carbonate aquifers: *Journal of Hydrology*, v. 14, p. 93-128.

Werner, E. 1972, Effect of small thrust faults on cave passage cross-section: *National Speleological Society Bulletin*, v. 34, p. 143-147.

Werner, E., 1977, Chloride ion variations in some springs of the Greenbrier limestone karst of West Virginia, *in* Dilamarter, R.R. and Csallany, S.C. (editors), *Hydrologic Problems in Karst Regions*: Western Kentucky University, Bowling Green, p. 357-363.

White, W.B., 1988, *Geomorphology and Hydrology of Karst Terrains*: Oxford University Press, New York, 464 p.

Nitrate Levels in the Karst Groundwaters of Tennessee

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ABSTRACT

Numerous wells and springs were sampled in the Ordovician and Mississippian carbonates of Tennessee to determine nitrate levels in order to ascertain the impact of man's activities on groundwater quality. Ten municipal springs in east Tennessee were sampled monthly for one year, and the results showed that nitrate levels remained below 2.5 mg/l (nitrate-nitrogen). Springs recharged from agricultural lands had significantly higher levels. Nitrate levels were also measured for sixteen domestic wells and twenty-four small springs in east Tennessee. The range and average for the wells were 0.02 to 2.64 mg/l and 1.05 mg/l, respectively. For the springs, the range and average were 0.02 to 5.60 mg/l and 1.95 mg/l, respectively.

Four large springs in central Tennessee were also sampled monthly for one year. The average nitrate levels for the springs recharged primarily by storm-water runoff from the city of Cookeville ranged from 0.69 to 0.94 mg/l; whereas, a spring recharged by agricultural lands outside of Cookeville had an average nitrate level of 3.09 mg/l. Sixty-four wells and springs around Cookeville were also sampled. The range and average for the wells in Mississippian-age beds were <0.1 to 31 mg/l and 1.2 mg/l, respectively. The range and average for the wells in Ordovician-age beds were <0.1 to 8.1 mg/l and 2.8 mg/l, respectively. Mississippian springs had a range of <0.1 to 2.9 mg/l with an average of 1.4 mg/l. Ordovician springs ranged from <0.1 to 3.3 mg/l with an average of 1.0 mg/l.

Municipal water-quality data supplied to the Tennessee Division of Water Supply was tabulated from twenty-nine wells and springs in the Cambrian-Ordovician carbonates of eastern Tennessee for the 1981-85 period. Multiple measurements produced a range of 0.01 to 3.27 mg/l and an average of 0.60 mg/l. Sixty-six measurements from 18 municipally used wells and springs in the Mississippian carbonates of central Tennessee showed a range of 0.01 to 3.67 mg/l and an average of 1.21 mg/l. Twenty-five nitrate measurements from municipal water supplies in the Central Basin Ordovician carbonate aquifer yielded a range of 0.01 to 2.81 mg/l and an average of 0.44 mg/l at six sites.

The results of this study show that nearly all samples were well below the 10 mg/l health limit for nitrate, but significantly higher levels occurred in rural areas where there is a greater acreage of agricultural activity. In general, nitrate levels were found to be higher in the Mississippian carbonates than in the Cambrian-Ordovician carbonates.

Introduction

In Tennessee, groundwater is the source of drinking water for 51 percent of the 4.76 million residents (Hutson, 1985). Approximately half of the state is underlain by fractured and cavernous carbonate rock that provides little filtration of contaminants as they move towards the water table. This is particularly true for nitrate which can cause methemoglobinemia or "blue babies disease" in newborn infants. As a result, the U.S. Environmental Protection Agency (EPA) (1982) has adopted a drinking-water limit of 10 mg/l nitrate-nitrogen. Natural background levels of nitrate-nitrogen are generally less than 3 mg/l. Higher concentrations suggest the influence of human-related sources such as crop fertilizers, septic tanks, land disposal of municipal and industrial waste, and animal wastes from livestock and poultry. Preliminary data from EPA's September 1, 1989 press advisory showed that about 50 percent of drinking-water wells tested had nitrate residue.

Growing concern about the potential impact of man's activities on nitrate levels in groundwater has promulgated a number of national and regional surveys (U.S. EPA, 1978; Zurawski, 1978; Madison and Brunett, 1985; Young, 1986; Adams and others, 1986; Canter, 1987; and Soileau, 1988). Madison and Brunett (1985) found that the seven states within the Tennessee River drainage had lower than national levels of nitrate with less than 1 to 4 percent of the wells exceeding the 10 mg/l drinking standard and less than 17 percent exceeding 3 mg/l. Soileau (1988) added EPA STORET data to Madison and Brunett's (1985) USGS WATSTORE data and found a mean concentration of 1.3 mg/l of nitrate-nitrogen and a range of <0.01 to 28.3 mg/l for 587 Tennessee Valley region wells sampled from 1956 to 1986. Very few of these wells were located in the state of Tennessee owing to an extreme paucity of data. Thus, the purpose of this paper is to present recent nitrate data that has been assimilated by the authors for the karst areas of central and eastern Tennessee to demonstrate the wide variability of nitrate concentrations both areally and temporally.

Sources of Data, Methodology, and Hydrogeology

Most of the data presented in this paper were collected and analyzed by the authors, utilizing the laboratory at the Center for the Management, Utilization and Protection of Water Resources of Tennessee Technological University. Analyses were performed on preserved samples by a Technicon Autoanalyzer GTPC (Standard Method 429 [APHA, 1989]). In addition, recent unpublished data from local studies and municipal water-quality data supplied to the Tennessee Division of Water Supply (1981-85) was incorporated in the survey.

The water samples were obtained from wells and springs in the flat-lying Ordovician- and Mississippian-aged carbonates of central Tennessee and the folded Cambrian-Ordovician-aged carbonates of eastern Tennessee (Figure 1). Most of the research has been conducted in and around Cookeville and Johnson City. Cookeville occurs on the Eastern Highland Rim Province which is underlain by flat-lying Mississippian limestones (Figure 2). Streams originating on the Cumberland Plateau to the east flow over shales and sandstones until underlying limestone beds are intersected. Some streams sink when they reach the Bangor Limestone whereas others sink into the lower Monteagle, St. Louis, or Warsaw limestones. Subterranean water moves through caves, pits, and solutionally enlarged fractures until emerging as spring flow. The lower Warsaw is sandy, contains shale beds, and acts as an aquiclude. As a result, numerous springs and caves are found at the St. Louis/Warsaw contact. The Fort Payne chert-rich limestone forms an areally extensive bench along the western edge of the Eastern Highland Rim Province. Discharge from the Fort Payne aquifer occurs along the Highland Rim Escarpment at the Chattanooga Shale contact. This water moves down the escarpment and then commonly sinks into cavernous strata within the Leipers-Catheys/Bigby-Cannon Ordovician formations of the Central Basin Province.

The geology of the Valley and Ridge Province of eastern Tennessee is significantly more complex. Surface water moves off mountains composed of clastics and igneous rock and sinks into strongly folded and faulted Cambrian-Ordovician carbonate rocks comprising the valley floors. The groundwater initially moves down gradient along bedding planes and fractures and then migrates along the strike (Ogden and others, 1989). Nearly all of the karst occurs in the Knox Group because it is over 3,000 feet thick and thus has a large outcrop area. Groundwater samples were also taken from springs in the Shady Dolomite and Copper Ridge Dolomite.

Results

Time-Series Analyses of Large Springs

As part of a wellhead-protection project in eastern Tennessee, nine springs were sampled monthly between July 1989 and June 1990 (Ogden and others, 1990). Johnson City springs emerge from quartzite, and the recharge area is entirely within National Forest boundaries. The low levels of nitrate are representative of a hydrogeologic environment free of man-made contaminants (Figure 3). Hampton Springs emerges from thick river alluvium overlying dolomite. Nitrate levels are also low even though the entire town of Hampton has individual home septic tanks. Blue, Big, Rockhouse caves, and Jonesboro springs emerge from limestone and exhibit conduit flow (based on dye tracing and geologic evidence). Gradual increases in nitrate were observed during the winter months at these four springs. During winter months, plant uptake of nitrates is at a minimum and nitrates are available to migrate to the water table. In addition, nitrates can be added to the groundwater from septic tanks during winter months when

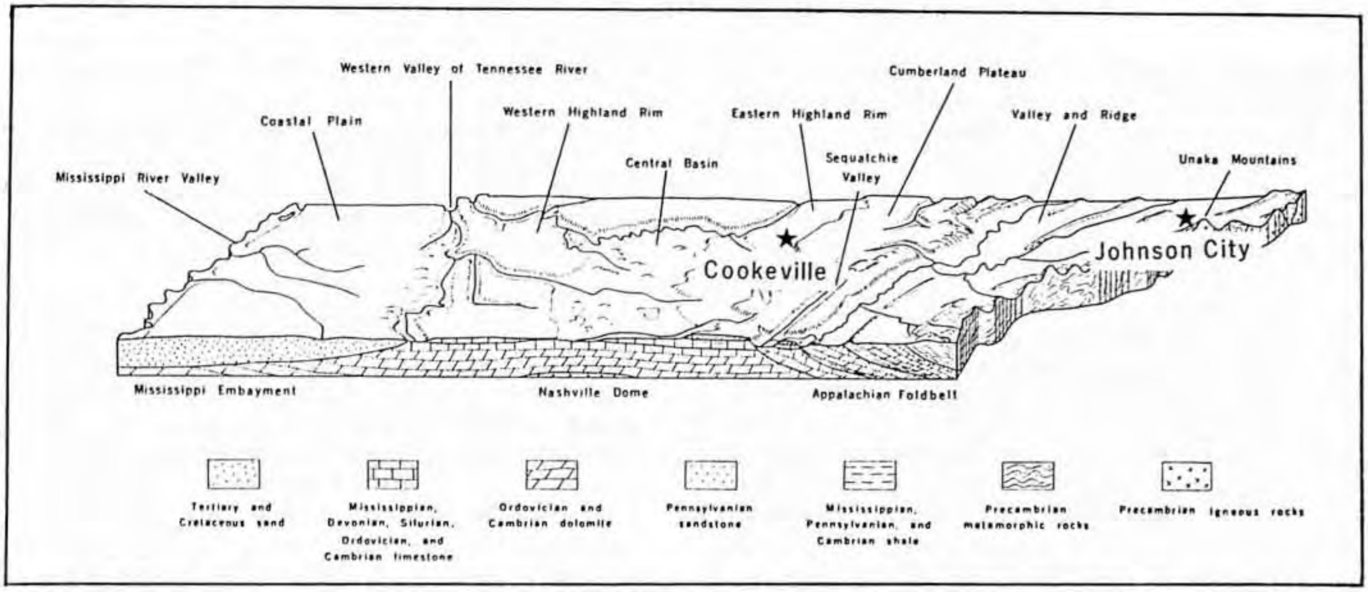

Figure 1: Relief map of Tennessee showing physiographic provinces and major geologic structures. The region around Cookeville lies on flat-lying carbonate rocks of the Eastern Highland Rim. The region around Johnson City lies on folded carbonate rocks in the Valley and Ridge Province (from Miller, 1974).

the soil is at field-capacity conditions. Overland flow to sinking streams through fields used by cattle is also greatest during the wet winter and spring months.

Lee and Hamilton springs emerge from dolostone, and the suspected recharge areas do not exhibit the degree of surficial karst features seen in the limestone terranes of the area. A significant component of diffuse flow is believed to exist within these dolomitic aquifers. Other measured chemical parameters not presented in this paper showed lower coefficients of variation than at the limestone springs. Regardless, conduit versus diffuse flow does not appear to be a primary factor affecting nitrate levels. Land use in the recharge areas is more important. Hamilton, Lee, and Big springs have the highest levels of nitrate; this is attributed to the recharge areas having a significantly greater percentage of pasture lands than do other spring basins.

Nitrate measurements were also made for these springs during a May 21, 1990 storm event (Figure 4). Figures 3 and 4 show that in general, nitrate levels vary little within

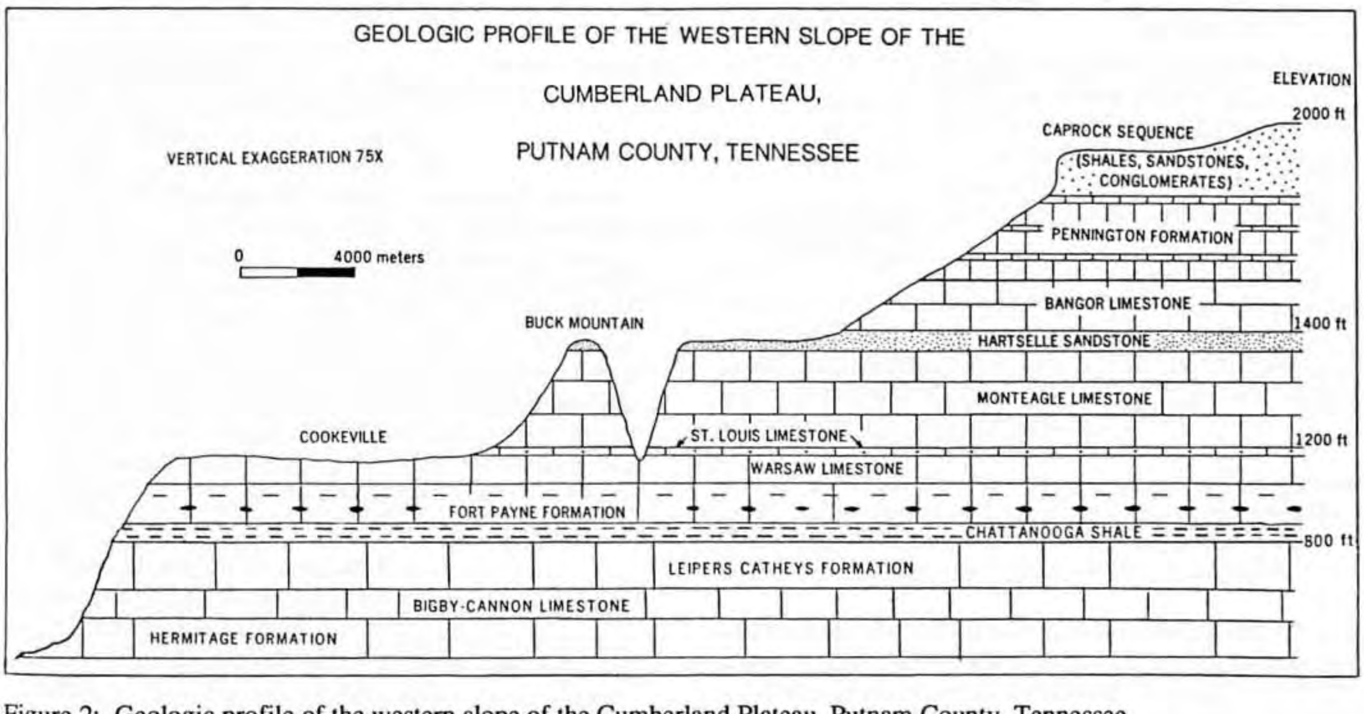

Figure 2: Geologic profile of the western slope of the Cumberland Plateau, Putnam County, Tennessee.

each groundwater basin with little change taking place due to storm events. As a result, any single measurement for nitrate will provide an estimate of the yearly range within approximately ±0.5 mg/l.

A similar time-series analysis of nitrate levels was performed at four large springs in the Cookeville area (Figure 5). All of these springs exhibit conduit flow and show significant turbidity during storm events. Hidden Hollow, Pigeon Roost, and Big springs all receive much of their recharge from storm-water runoff from paved areas in Cookeville. In contrast, City Springs is recharged almost exclusively by farmlands, and this is the likely cause for significantly higher levels of nitrate. Although City Springs has more nitrate, its overall water quality is much better than at the other springs. The cave water at City Springs supports a much greater diversity of species, and several blind cave fish have been observed (Pride and others, 1989).

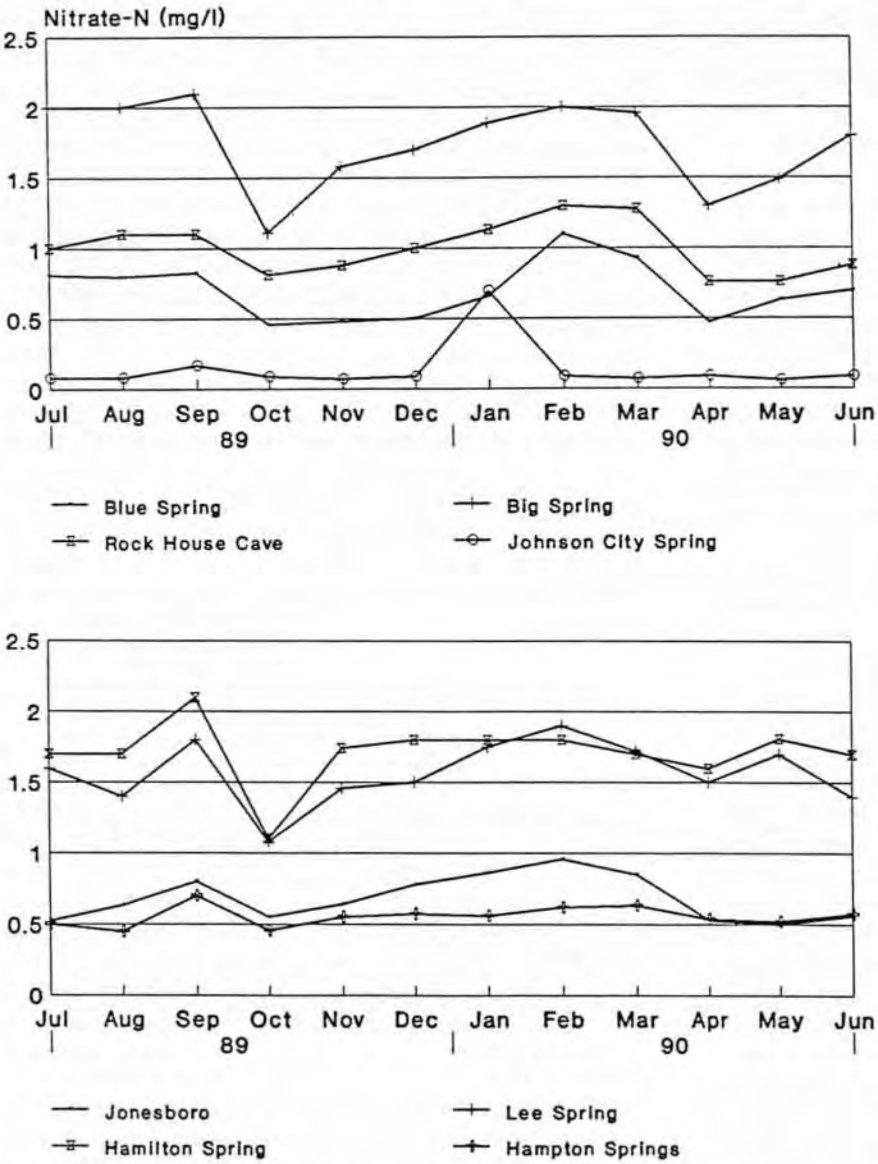

Figure 3: Monthly nitrate measurements at eight eastern Tennessee springs.

Domestic Wells and Small Springs

Nitrate levels were also measured for sixteen domestic wells and twenty-four small springs in eastern Tennessee. The range for the wells was 0.02 to 2.64 mg/l, and the average was 1.05 mg/l. For the springs, the range and average were 0.02 to 5.60 mg/l and 1.95 mg/l, respectively.

Collar (1989) analyzed sixty-four wells and springs for nitrate levels in the Cookeville area. He found that Fort Payne (Mississippian) wells had a range of <0.1 to 31 mg/l with an average of 1.2 mg/l; whereas, the Ordovician-aged wells had a range of <0.1 to 8.1 mg/l with an average of 2.8 mg/l. Springs in the Fort Payne Limestone had a range of <0.1 to 2.9 mg/l with an average of 1.4 mg/l, and the springs in the Ordovician-aged carbonates had nitrate levels that ranged from <0.1 to 3.3 mg/l with an average of 1.0 mg/l. Collar (1989) found no statistically significant differences in these averages using the Mann-Whitney U Test.

Public Drinking-Water Supplies

Every three years the municipalities supplying public water must submit a full analysis of their water quality to the Tennessee Division of Water Supply. This information was obtained and tabulated for the 1981 - 1985 period and divided into groups based on the geographic distribution of principal aquifers in Tennessee according to Bradley and Hollyday (1985). One hundred and four samples from 29 municipal wells and springs in the Cambrian-Ordovician carbonate aquifer of eastern Tennessee yielded a range in nitrate values of 0.10 to 3.27 mg/l and an average of 0.60 mg/l. Sixty-six samples from 18 municipal water supplies in the Mississippian carbonate aquifer of central Tennessee produced a range in nitrate values of 0.10 to 3.67 mg/l with an average of 1.21 mg/l. Six municipally used wells and springs in the Ordovician carbonate aquifer of the Central Basin Physiographic Province yielded a range of 0.01 to 2.81 mg/l and an average of 0.44 mg/l nitrate-nitrogen from twenty-five measurements.

Summary and Conclusions

A survey of numerous wells and springs in the Cambrian-Ordovician and Mississippian carbonate aquifers of Tennessee has revealed that nearly all waters are well below the safe drinking limit for nitrate-nitrogen of 10 mg/l. Table 1

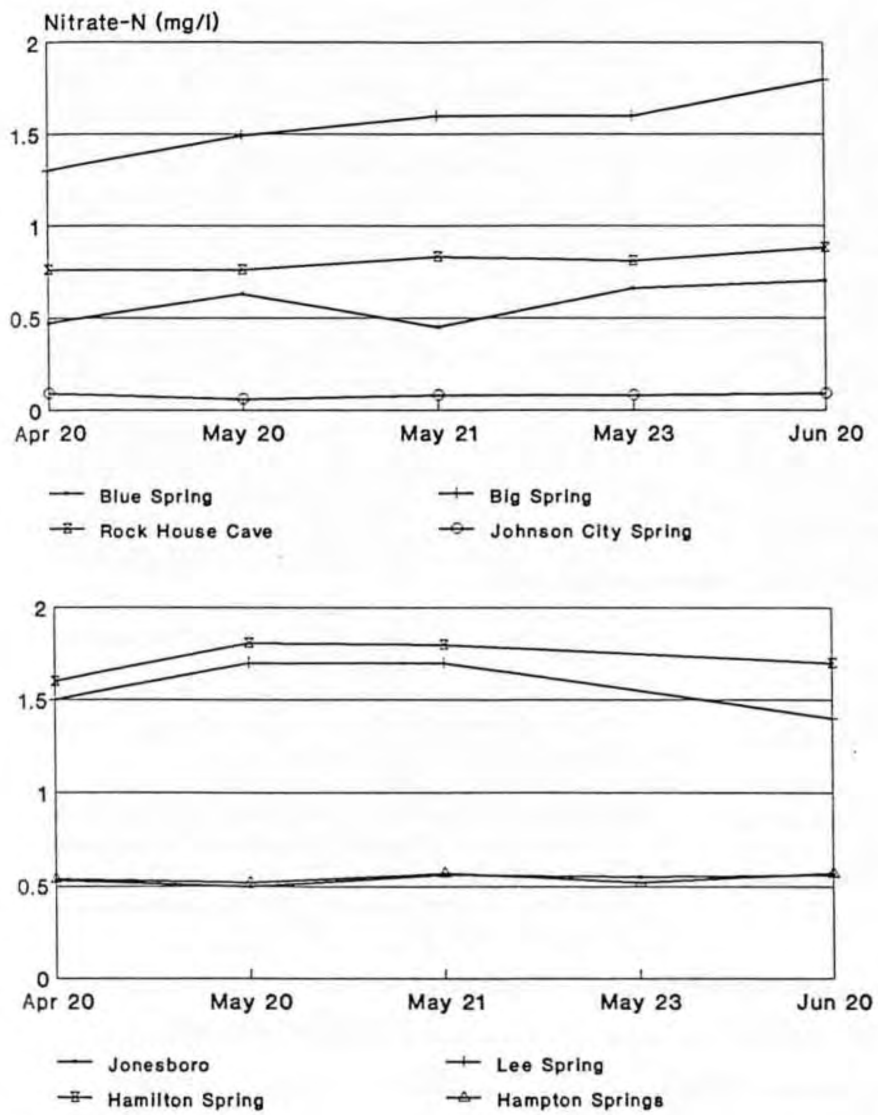

Figure 4: Nitrate measurements at eight eastern Tennessee springs during a May 1990 storm event.

summarizes the data presented in this report. In general, groundwaters in the Mississippian carbonates have higher levels of nitrate than in the Cambrian-Ordovician carbonates, although a great degree of variability occurs among formations within these two age groups. The time-series analysis of the spring-water data has shown that little variability of nitrate levels occurs throughout the year, but slight increases are observed during the winter when plant use of nitrate is at a minimum and recharge is the greatest. A predominance of conduit versus diffuse flow does not appear to cause a significant difference in yearly average nitrate levels or the amount of change resulting from small to moderate storm events.

Significantly higher levels of nitrate are found in springs that largely drain land grazed by cattle. Fertilizer application to croplands may be a factor, although this type of land use is minimal in the study areas. Septic tanks likely contribute some nitrate to the groundwater, but a clear relationship between septic-tank density and nitrate levels was not seen at the springs where the recharge areas had been delineated by dye tracing. Therefore, it is felt that more concern must be placed on agricultural practices to insure groundwater protection.

Acknowledgments

This research was funded in part by the University of Tennessee - Water Resources Research Center (USGS) and the Center for the Management, Utilization and Protection of Water Resources - Tennessee Technological University. Some matching funds were also provided by the First Tennessee Development District. The authors thank the following students for assisting in this project: Walter Crawford, John Mason, and Scott Wheeler.

References

Adams, V.D.; Wilson, T.M.; Brown, R.T.; Gordon, J.A.; and Mills, H.H., III, 1986, Design of a statewide ground-water monitoring network: *Tennessee Water Re-sources Research Center, University of Tennessee, Knoxville, Research Report No. 13*, 199 p.

American Public Health Association, American Water Works Association, and Water Pollution Control Federation, 1989, *Standard Methods for the Examination of Water and Waste Water, 14th edition*: American Public Health Association, Washington, DC.

Bradley, M.W. and Hollyday, E.F., 1985, Tennessee ground-water resources, *in* National water summary 1984; Hydrologic events, selected water-quality trends, and ground water resources: *U.S. Geological Survey Water-Supply Paper 2275*, p. 391-396.

Canter, L.W., 1987, Nitrates and pesticides in ground water: An analysis of a computer-based literature search, *in Ground Water Quality and Agricultural Practices*: Lewis Publishers, Chelsea, Michigan, p. 153-174.

Collar, P.D., 1989, *Radon in homes, soils, and caves, and water quality of wells and springs in north central Tennessee*: M.S. thesis (unpublished), Tennessee Technological University, 342 p.

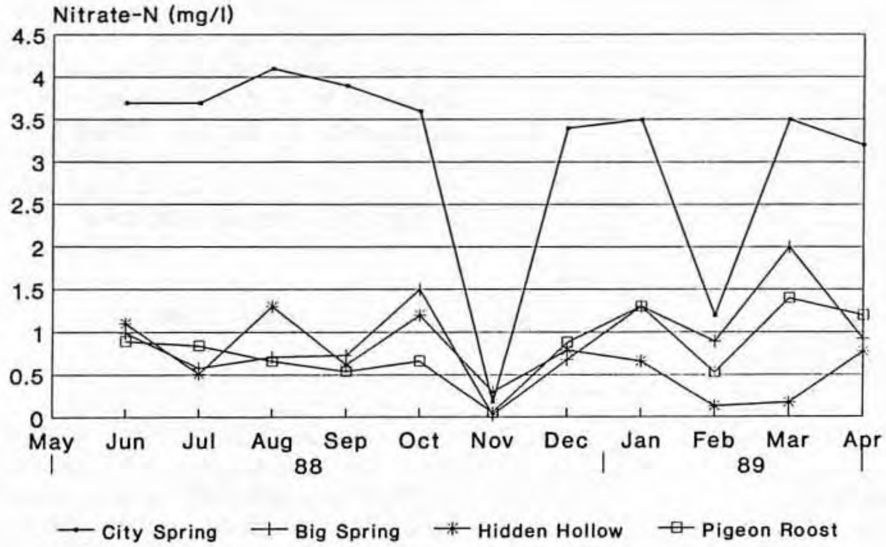

Figure 5: Monthly nitrate measurements at four central Tennessee springs near Cookeville.

Hutson, S.S, 1985, Ground-water use by public-supply systems in Tennessee in 1985: *U.S. Geological Survey Water Resources Investigations Report 89-4092*, 1 plate.

Madison, R.J. and Brunett J.O., 1985, Overview of the occurrence of nitrate in ground water of the United States, *in* National water summary 1984; Hydrologic events, selected water-quality trends, and ground water resources: *U.S. Geological Survey Water Supply Paper 2275*, p. 93-105.

Miller, R.A., 1974, The Geologic History of Tennessee: *Tennessee Division of Geology Bulletin 74*, Nashville, 63 p.

Ogden, A.E.; Curry, W.; and Cummings, J., 1989, Morphometric analysis of sinkholes and caves in Tennessee comparing the Eastern Highland Rim and the Valley and Ridge physiographic provinces, *in* Beck, B.F. (editor), *Engineering and Environmental Impacts of Sinkholes and Karst: Proceedings of the Third Multidisciplinary Conference on Sinkholes and the Engineering and Environmental Impacts of Karst, St. Petersburg Beach, Florida, 2-4 October 1989*: A.A. Balkema, Rotterdam and Brookfield, p. 135-142.

Ogden, A.E.; Hamilton, K.G.; and Brown, T.L., 1990, Delineation of wellhead protection areas for municipal-used springs of eastern Tennessee: *Tennessee Water Resources Research Center, University of Tennessee, Knoxville, Report 124*, 134 p.

Pride, T.E.; Ogden, A.E.; and Harvey, M.J., 1989, Biology and water quality of caves receiving urban runoff in Cookeville, Tennessee, USA: *Proceedings of the 10th International Congress of Speleology, Budapest, Hungary*, p. 27-29.

<table>
<tr><th rowspan="3">Data Source</th><th colspan="4">Mississippian Carbonates</th><th colspan="4">Ordovician Carbonates</th></tr>
<tr><th colspan="2">Wells</th><th colspan="2">Springs</th><th colspan="2">Wells</th><th colspan="2">Springs</th></tr>
<tr><th>Range</th><th>Average</th><th>Range</th><th>Average</th><th>Range</th><th>Average</th><th>Range</th><th>Average</th></tr>
<tr><td>Wellhead Protection Project (Ogden et al., 1990)</td><td></td><td></td><td></td><td></td><td>0.02 - 2.64</td><td>1.05</td><td>0.45 - 2.10</td><td>1.21</td></tr>
<tr><td>Septic Tank Survey (Wilson et al., 1989)</td><td></td><td></td><td></td><td></td><td></td><td></td><td>0.02 - 5.60</td><td>1.95</td></tr>
<tr><td>Cookeville Sinkhole Project (Pride et al., 1989)</td><td></td><td></td><td>0.69 - 4.10</td><td>1.38</td><td></td><td></td><td></td><td></td></tr>
<tr><td>Putnam/Jackson County Water Quality Survey (Collar, 1989)</td><td><0.10 - 31.0</td><td>1.20</td><td><0.10 - 2.90</td><td>1.40</td><td><0.10 - 8.10</td><td>2.80</td><td><0.01 - 3.30</td><td>1.00</td></tr>
<tr><td rowspan="3">Tennessee Division of Water Supply (1981 - 1985)</td><td colspan="4">Wells and Springs</td><td colspan="4">Wells and Springs</td></tr>
<tr><td colspan="2">Range</td><td colspan="2">Average</td><td colspan="2">Range</td><td colspan="2">Average</td></tr>
<tr><td colspan="2">0.01 - 2.81</td><td colspan="2">0.44</td><td colspan="2">0.01 - 3.27</td><td colspan="2">0.60</td></tr>
</table>

Table 1: Range and average of nitrate concentrations (mg/l Nitrate-N) in karst groundwaters of Tennessee.

Soileau, J.M., 1988, An overview of ground-water nitrate and fertilizer nitrogen consumption in the Tennessee Valley region: *Tennessee Valley Authority, Office of Agriculture and Chemical Development Bulletin 204*: Muscle Shoals, Alabama, 29 p.

Tennessee Division of Water Supply, 1981-1985, *Water Quality Records for Municipal Water Supplies, Nashville, Tennessee.*

U.S. Environmental Protection Agency, 1978, *National Water Quality Inventory: 1977 Report to Congress: EPA 440/2/84-006*, Washington, DC.

U.S. Environmental Protection Agency, 1982, Maximum contaminant levels (Subpart B of Part 141, National Interim Primary Drinking Water Regulations): *U.S. Code of Federal Regulations, Title 40, Parts 100 to 149, Revised July 1, 1982*, p. 315-318.

U.S. EPA Press Advisory, 1989, *EPA Releases Interim Results of National Survey of Pesticides in Drinking-Water Wells, Washington, DC.*

Wilson, T.M.; Reinhard, J.D.; Gordon, J.A.; Ogden, A.E., 1989, A study of septic tank systems in Sullivan County, Tennessee: *Report to the Tennessee Valley Authority, Chattanooga, Tennessee*, 90 p.

Young, H.C., 1986, Distribution maps of Tennessee Valley region: *National Fertilizer Programs, Agricultural Programs, Major Land Uses, Livestock Inventions, Farm Products and Sales: Circular Z-211 TVA/OACD*, Muscle Shoals, Alabama, 46 p.

Zurawski, A., 1978, Summary appraisals of the nation's ground-water resources--Tennessee region, including parts of Tennessee and adjacent states: *U.S. Geologi-cal Survey Professional Paper 813-L*, p. L1-L35.

Plate K: A sinkhole in rural Pulaski County, Virginia, filled with farm and household trash. It has been estimated that thousands of these sinkhole dumps exist within the groundwater recharge zones of the Appalachian karst of Virginia and West Virginia. (*See* papers by Erchul, p. 147, Kastning and Kastning, p. 123, and Hubbard, p. 135, in this volume). *Photograph by Ernst H. Kastning.*

Impacts of Barnyard Wastes on Groundwater Nitrate-N Concentrations in a Maturely Karsted Carbonate Aquifer of South-Central Kentucky

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ABSTRACT

In maturely karsted areas such as the Sinking Valley groundwater basin of south-central Kentucky, where conduit development is extensive and soil cover is thin, human and animal wastes can pose a serious threat to the local groundwater system. The water chemistry of domestic wells, other groundwater-access points, and surface streams throughout the basin were monitored between June of 1987 and March of 1988. In general, nitrate-N concentrations were less than 5 mg/l. Nitrate-N levels in one domestic well (Harris Well), however, exceeded the drinking-water standard. The suspected nitrogen source for the elevated nitrate-N levels appears to be a barnyard within 200 feet of the well; septic discharge from the house occurred down slope from the well and was also a potential factor.

Nitrogen sources contributing to groundwater-nitrate occurrences throughout most of the basin consisted primarily of N-fertilizers, animal wastes and decomposed plant matter. Nitrates behaved as non-point-source contaminants and were leached through the soil and epikarst and into the groundwater system during recharge events and/or under field-saturated conditions, by diffuse autogenic recharge and diffuse flow. Following a rain event, nitrate-N levels at most groundwater-sampling locations would typically decrease initially, in response to dilution by concentrated autogenic recharge. Following this initial decrease, nitrate-N levels would typically increase above base-flow concentrations as the diffuse autogenic recharge component became a larger contributing factor toward groundwater flow. Sources of nitrate and other contaminants that were introduced at discrete locations in the limestone resulted in deleterious effects on local groundwater quality. Such was the case of Harris Well, where a nearby barnyard is located over thinly-soiled limestone. Groundwater in Harris Well showed an inverse response of nitrate-N concentrations, as well as specific conductance and hardness, to rainfall events. Following the event, these levels rebounded to or near base-flow levels, indicating the existence of a nitrogen source that was continually leached into the groundwater and well.

Introduction

The outcrop of Mississippian limestones along the Cumberland escarpment in eastern south-central Kentucky (Figure 1) is a unique and, thus far, relatively unstudied karst region. The area is characterized by a thick sequence of soluble limestones, thin soil cover, moderate relief, and low structural control. The Sinking Valley groundwater basin within this region is wholly agricultural and occupies approximately 52 square miles. The groundwater system exhibits conduit flow and a focused point of groundwater discharge, thus providing an excellent setting in which to determine the behavior of nitrates, to establish the major sources contributing to groundwater nitrate, and to determine the effects of groundwater-recharge events on nitrates.

Harris Well was among eleven sampling locations throughout the basin that were sampled at least biweekly between June 13, 1987 and February 16, 1988 (Figure 2). These locations included other domestic wells, springs, caves, and surface streams. In addition to concentrations of

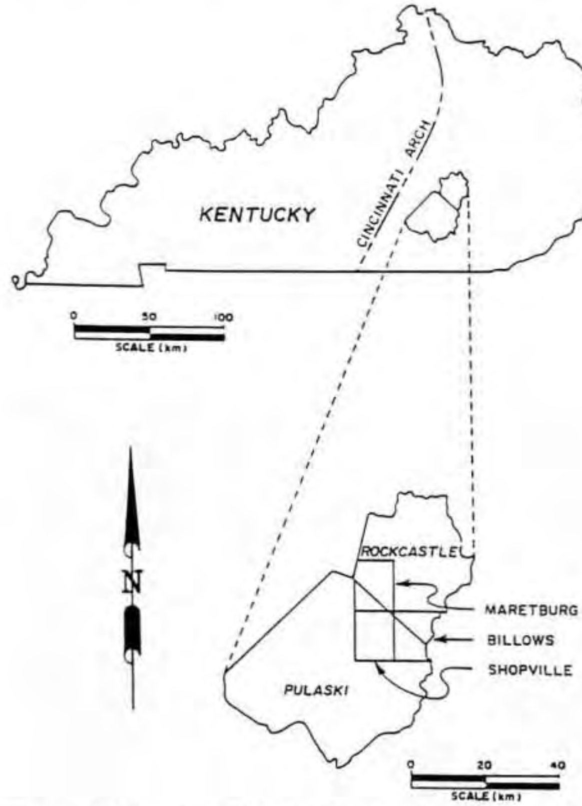

Figure 1: Location of the study area in three quadrangles in Pulaski and Rockcastle counties, Kentucky.

nitrate-N and nitrite-N, monitored water-chemistry parameters included calcium and magnesium hardness, alkalinity, specific conductance, pH, chlorides, and coliform bacteria. Calculated parameters P_{CO_2} and SIc were also determined at each sample location in an effort to characterize the groundwater recharge and flow components (Drake and Harmon, 1973). Environments of sampling locations in the study area displayed differences in certain water-chemistry parameters, thus allowing the dominant recharge and flow mechanism to be identified.

A network of rain gauges was distributed over the basin to monitor precipitation (Figure 2), while groundwater discharge was monitored during much of the study at a karst window near the groundwater-basin discharge point. The period of study was unusually dry with respect to rainfall and low-flow conditions occurred between July and December.

Background

Groundwater flow in mature karst has been the focus of several investigations in the attempt to characterize recharge, storage, and flow mechanisms of particular aquifer systems. These properties can, in turn, be applied to contaminant (nitrate) behavior.

White (1969) distinguished the flow behavior of different carbonate aquifers by proposing two end-member models, diffuse flow and conduit (or free) flow. Although greatly oversimplified, these end-member models can be compared with discharge from actual carbonate aquifers. Gunn (1985) provided a conceptual model for conduit-flow dominated karst aquifers. The model examines the three main system components: inputs, stores, and transfer mechanisms.

Depending upon the nature of the karst aquifer, the storage and movement of nitrate will behave in different ways. In dye-tracer studies in the Mendip Hills, England, Friederich and Smart (1981) found that dye spread laterally through the subcutaneous zone at flow rates on the order of 100 meters per day. However, dye was still present ten months after its artificial injection into the epikarst, indicating significant storage in this zone.

Similarly, Gerhart (1986) found that nitrate-N concentrations in a shallow, fractured dolomite aquifer are influenced by two different recharge mechanisms: direct recharge through fractures and sinkholes that affects nitrate concentrations of groundwater for two to three days; and gradual recharge that occurs through small channels and micropores in the unsaturated zone and affects nitrate levels for several weeks or more.

An important factor influencing groundwater nitrate-N levels is, of course, the contributing nitrogen source. The complexity of nitrate within the soil-groundwater system makes its behavior difficult to understand. There are several nitrogen sources potentially contributing to nitrate-N levels in the Sinking Valley groundwater basin, including animal and human wastes, fertilizers, and decayed plant matter.

The increasing confinement of animals in large numbers for meat, milk, and egg production has caused large quantities of animal waste to accumulate in small areas (Wadleigh, 1968), creating the potential to contaminate underlying groundwater. Aldwell and others (1983) reported that waste from farmyards and other areas of livestock concentration is the most common and widespread source of groundwater pollution in Ireland. Brown (1990) found that diffuse autogenic recharge had the greatest overall effect upon nitrate-N levels in a conduit-dominated karst aquifer in south-central Kentucky. Localized groundwater conditions were greatly affected by concentrated nitrogen sources (barnyards), however, and exhibited excessive levels of nitrate-N and coliform bacteria.

Studies by the U.S. Environmental Protection Agency show that 29% of the U.S. population disposes of domestic waste through individual on-site disposal units, and investigators estimate that many of these septic tank-soil absorption systems are not operating satisfactorily (Scalf and others, 1977). In addition to nitrates, other products of human waste associated with septic drainage include bacteria, suspended solids, biological oxygen demand (BOD),

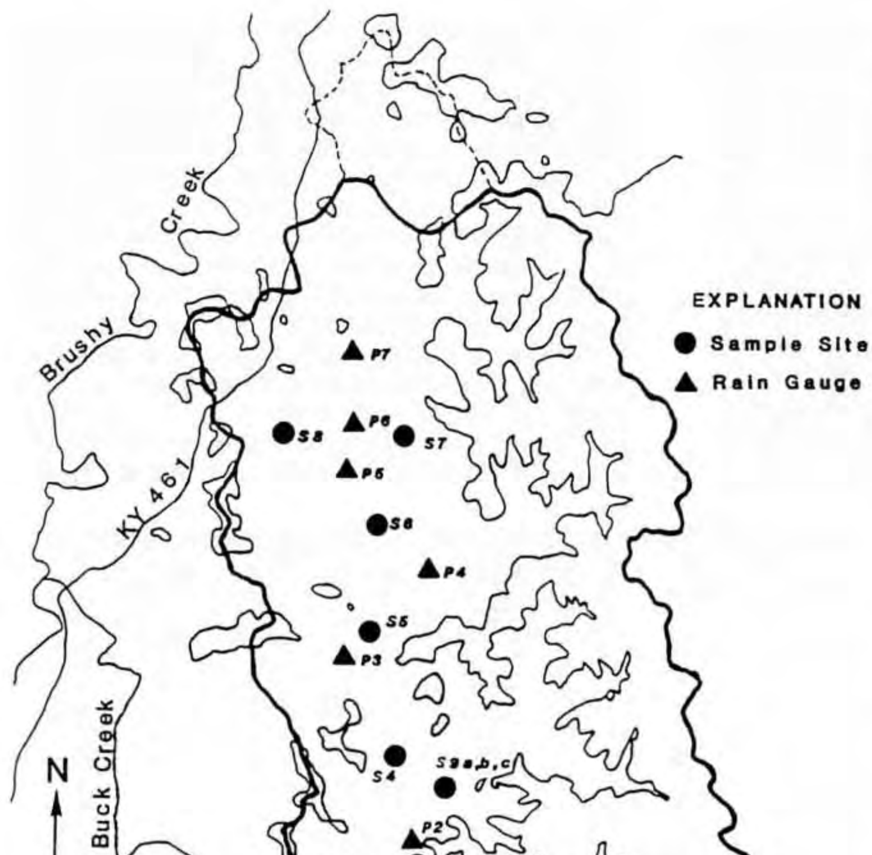

Figure 2: Location of sampling sites and rain gauges in the Sinking Valley Groundwater Basin.

viruses, ammonia, chlorides, phosphates, and sodium, as well as waterborne diseases such as infectious hepatitis.

Application of nitrogen fertilizer to agricultural areas has increased tremendously since the end of World War II, and fertilizers are generally considered to be the primary source of nitrate concentration (Madison and Brunett, 1985). Halberg (1986) ascertained that a store of nitrogen from fertilizer exists in the karst soil mantle in Iowa. Leaching of nitrate-N into the groundwater system was found to be directly related to recharge events in the spring.

An additional, but somewhat less influential source of nitrates in groundwater results from the leaching of nitrates following harvest. Soil conditions in autumn promote rapid nitrification of organic nitrogen and, following harvest, can lead to considerable leaching of nitrate ions (Addiscott, 1986).

Hydrogeologic Setting

Using terminology proposed by White (1969), the Sinking Valley groundwater system can be classified as a conduit-dominated carbonate aquifer. This categorization is supported by the presence of numerous sinkholes, dry valleys, sinking streams, extensive solutional development, and characteristically high groundwater flow rates throughout the Sinking Valley region.

A thick sequence of Mississippian limestones within the Newman Formation makes up most of the Sinking Valley aquifer system. The Ste. Genevieve and upper member limestones have undergone significant dissolution and provide the pathway for most groundwater flow. The underlying St. Louis Limestone contains abundant chert in the upper horizon that inhibits dissolution. Overlying these limestone units are mudstones, sandstones, and limestones of the Pennington Formation, forming the upper part of knobs and ridges, and conglomerates and sandstones of the Lee Formation (Pennsylvanian) that cap the hills.

Previous dye-tracing studies have demonstrated that groundwater flow in the basin follows a dendritic pattern, converging toward a central conduit that discharges through a karst spring into Buck Creek, a surface stream to the west (Romanik, 1986). The conduit network transports groundwater down dip in a southerly direction before making a shift in the southern part of the basin to follow strike toward the southwest. Short Creek, a karst window, was a major monitoring point immediately upstream from the groundwater-basin discharge spring.

A survey of land use, agricultural practices, and septic disposal was performed to determine the major sources of nitrates in the Sinking Valley. The survey data indicate that although the basin was not farmed intensively, nearly 20% of the area was used to grow crops including corn, tobacco, hay, and alfalfa, whereas over 35% was dedicated to pasture. Livestock are raised in pastures and barnyards throughout the basin; such locations represent a concentrated area in which nitrogen is introduced into the ground.

Septic-disposal practices in the study area were found to range from septic tank - leach field systems to direct discharge of raw sewage onto the ground surface.

Nitrate-N Levels in the Groundwater System

Nitrate-N levels in the Sinking Valley groundwater system demonstrated a seasonal relationship. Groundwater nitrate-N levels in Short Creek (Figure 3), representative of conduit flow, were found to increase in the spring months, apparently resulting from the application of fertilizers. During summer and early fall months and baseflow conditions, nitrate-N concentrations were relatively stable, as low rainfall and the effects of evapotranspiration prevented the leaching of nitrates through the soil and epikarst and into the groundwater system. At the onset of significant groundwater recharge and the cessation of evapotranspiration, groundwater nitrate-N levels throughout the basin increased sharply as nitrates in the soil and epikarst were carried by recharge (infiltration) into the groundwater system. Following the major period of groundwater recharge in the fall and early winter, nitrate-N concentrations in groundwater declined, probably due to the depletion of nitrates in the soil and epikarst, coupled with dilution by increased groundwater flow. Thus, seasonal trends were observed in groundwater nitrate levels.

Nitrate-N levels in groundwater at most sampling locations, including basin discharge monitored at Short Creek, exhibited a strong relationship with recharge-producing rain events. Early on in the course of a rain event, groundwater chemistry was largely representative of conduit flow, or quickflow, supplied dominantly by concentrated autogenic recharge. Low levels of nitrate-N, calcium hardness, and specific conductance were detected in groundwater at this time (Figure 4). As diffuse recharge (infiltration) and diffuse flow began to play a larger role in groundwater flow, nitrates stored in the soil and epikarst were leached into the groundwater system, resulting in an increase in nitrate-N levels. This increase usually peaked after the rain event and remained above base-flow levels for several days.

Harris Well

Harris Well was the only sampling location where nitrate-N concentrations exceeded the drinking-water standard of 10 mg/l, and remained above the limit for most of the study (Figure 5). Nitrate-N concentrations in this well behaved differently than other sample locations in that nitrate-N, along with hardness and specific conductance (Figure 6), demonstrated a negative correlation with rainfall events and obtained highest concentrations between rain events. Nitrate-N in groundwater at all other locations increased only in response to a recharge-producing rain event. A comparison of groundwater parameters at Harris Well and Short Creek during low and high flow is provided in Table 1. Parameters in both sample locations decreased following the increased groundwater flow caused by recharge events between October 4, 1987 and January 20, 1988; decreased nitrate-N levels in Short Creek over this period reflect the flushing and dilution of

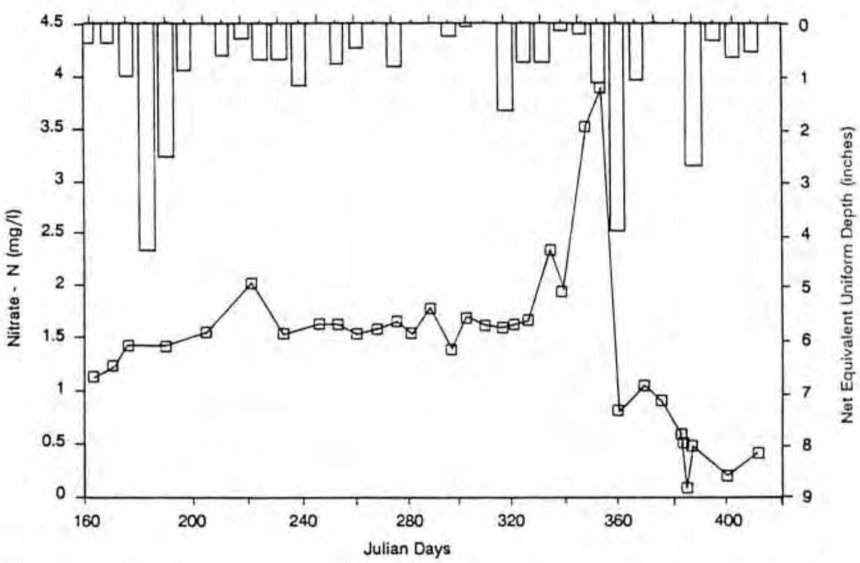

Figure 3: Nitrate-N concentrations and precipitation vs. time at Short Creek.

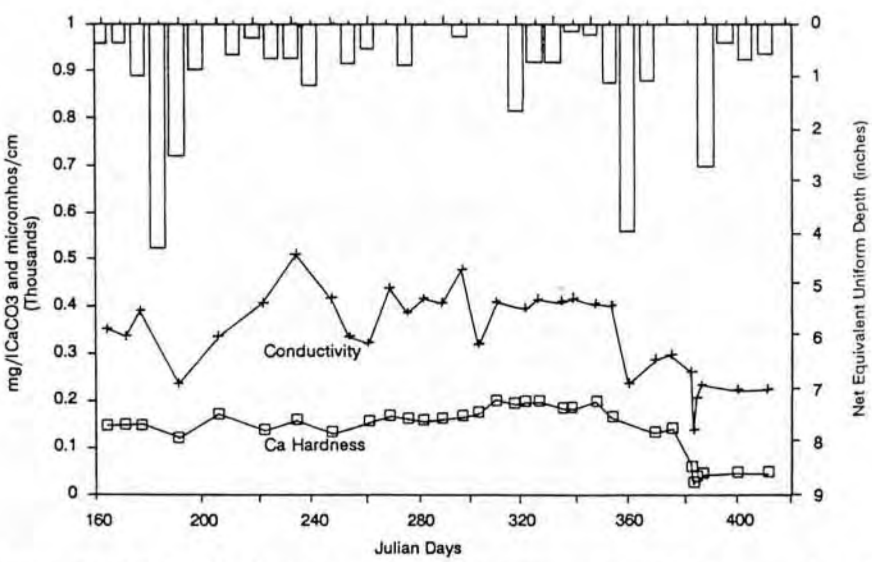

Figure 4: Calcium hardness, specific conductance, and precipitation vs. time at Short Creek.

nitrates stored in the soil and epikarst. In addition to nitrate-N, other parameters in Harris Well were anomolously high, including specific conductance (737 µmhos/cm), chlorides (52.5 mg/l), and fecal coliform (400 colonies/100 ml).

Based on the strikingly different relationship of nitrate-N with rainfall from the rest of the groundwater system, the significantly higher levels, and the associated parameters indicative of animal or human waste, it is apparent that the groundwater nitrate originated from a major source other than nitrogen fertilizers, as was attributed to other sample locations.

Groundwater in Harris Well appeared to reflect contamination by localized animal and possibly human wastes. The probable nitrogen source was a cattle barnyard approximately 180 feet southeast and uphill from the uncased well. Another, less likely source exists about 50 feet west of the well, where sewage was discharged from the back of the house onto the ground. Contaminants from this source are downgradient from the well and would have a more tortuous path to travel.

From the nitrate-N and precipitation vs. time relationship it is evident that nitrate was introduced to and stored within the soil and epikarst and continuously leached into the groundwater and Harris Well through diffuse and conduit (mixed) flow. Following a rain event and the resultant flushing and dilution effects caused by concentrated recharge and conduit flow, nitrate-N concentrations in the groundwater decreased.

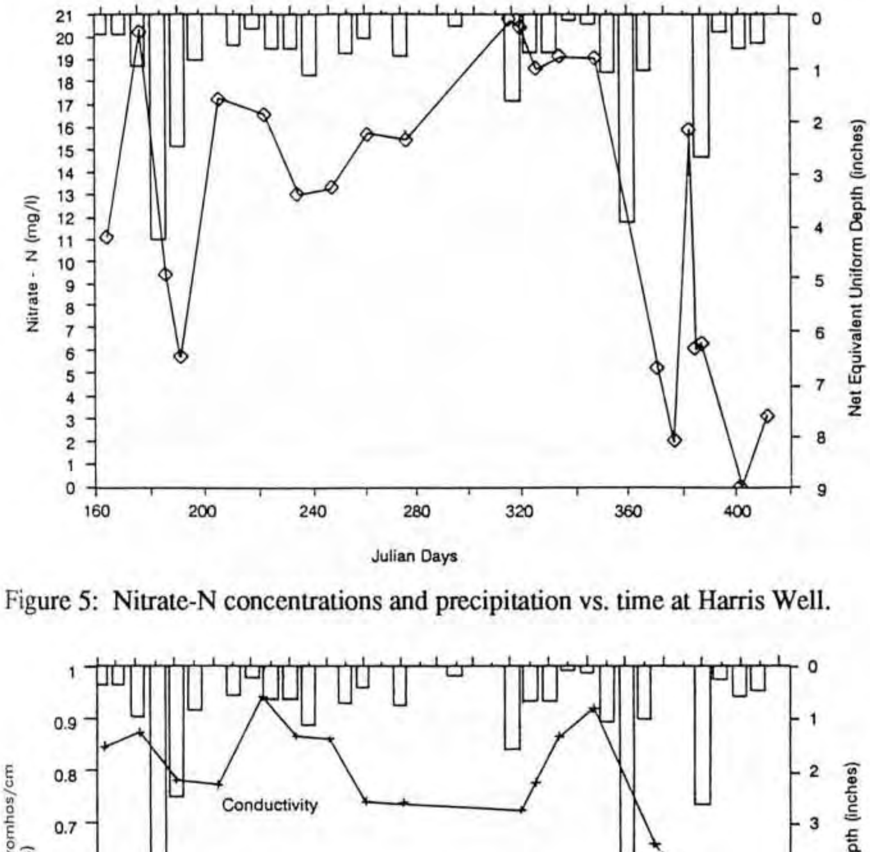

Figure 5: Nitrate-N concentrations and precipitation vs. time at Harris Well.

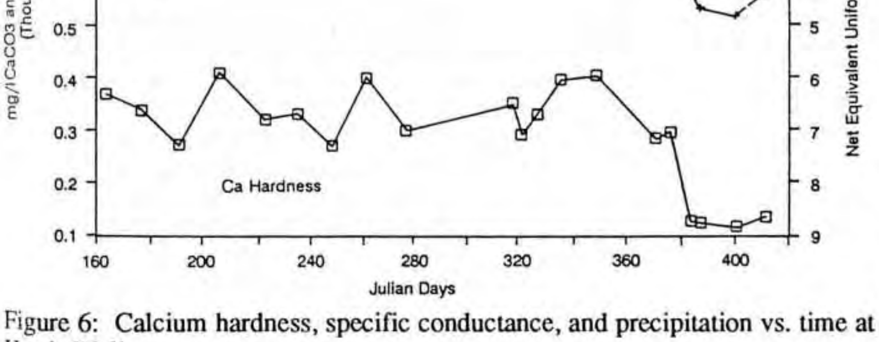

Figure 6: Calcium hardness, specific conductance, and precipitation vs. time at Harris Well.

Conclusions

Thinly mantled, conduit-dominated karst aquifers in eastern south-central Kentucky such as the Sinking Valley basin are extremely vulnerable to contaminants from the surface. Although it was determined that nitrogen fertilizers were not applied in quantities sufficient to generate excessive nitrate-N levels in groundwater, animal husbandry and ineffective septic systems can result in high levels of nitrates, as well as bacteria and other associated contaminants, in localized groundwater environments.

In most sampling locations of the Sinking Valley groundwater system, elevated groundwater nitrate-N levels (above 1 mg/l) appeared to result mostly from the application of fertilizers in the spring and subsequent leaching by rainfall events. Increased nitrate-N concentrations exhibited strong correlations with recharge events following the growing season.

In Harris Well, however, a nitrogen source of a different nature was apparent, as indicated by much higher concentrations, different behavior, and associated occurrence with elevated levels of chlorides and coliform bacteria. High nitrate-N levels occurred in Harris Well between rain events, but exhibited sharp reductions immediately following recharge events, apparently associated with flushing and dilution of groundwater affected by barnyard wastes and/or human wastes.

Table 1: Groundwater chemistry parameters at Harris Well and Short Creek during low- and high-flow conditions.

	Date	Nitrate-N (mg/l)	Total Hardness (mg/l)	Chloride (mg/l)	Specific Conductance (μnhos/cm)	Fecal Coliform (per 100 ml)
HARRIS WELL: Low flow	October 4, 1987	16.26	324	52.5	737	400*
High flow	January 20, 1988	8.33	131	—	584	—
SHORT CREEK: Low flow	October 4, 1987	1.67	178	4.25	387	20*
High flow	January 20, 1988	0.55	31	—	140	—

*Measured December 2, 1987.

References

Addiscott, T., 1986, Farmers, fertilizers and the nitrate flood: *New Scientist*, v. 8, p. 50-54.

Aldwell, C.R.; Burdon, D.J.; and Sherwood, M., 1983, Impact of agriculture on groundwater in Ireland: *Environmental Geology*, v. 5, p. 39-48.

Brown, C.J., 1990, *The Effects of Ground-Water Recharge on Nitrate-N Concentrations in a Conduit-Dominated Carbonate Aquifer, the Sinking Valley Ground-Water Basin, South-Central Kentucky*: M.S. thesis (unpublished), Eastern Kentucky University, Richmond, 145 p.

Drake, J.J. and Harmon, R.S., 1973, Hydrochemical environments of carbonate terrains: *Water Resources Research*, v. 9, p. 949-957.

Friederich, H. and Smart, P.L., 1981, Dye tracer studies of the unsaturated-zone recharge of the carboniferous limestone aquifer of the Mendip Hills, England *in* Beck, B.F. (editor), *Proceedings of the Eighth International Congress of Speleology, Bowling Green, Kentucky, July 18-24, 1981*: National Speleological Society, Huntsville, Alabama, v. 1, p. 283-286.

Gerhart, J.M., 1986, Ground-water recharge and its effects on nitrate concentration beneath a manured field site in Pennsylvania: *Ground Water*, v. 24, no. 4, p. 483-489.

Gunn, J., 1985, A conceptual model for conduit flow dominated karst, *preprint from* International Symposium on Karst Water Resources, Ankara (in press), p. 1-10.

Halberg, G.R., 1986, Overview of agricultural chemicals in groundwater: *Proceedings of the Agricultural Impacts on Ground Water Conference, Omaha, Nebraska, August 1986*: National Water Well Association, p. 1-63.

Madison, R.J. and Brunett, J.O., 1985, Overview of the occurrence of nitrate in ground water of the United States, *in* National water summary 1984; Hydrologic events, selected water-quality trends, and ground-water resources: *U.S. Geological Survey Water Supply Paper 2275*, p. 93-105.

Romanik, P.B., 1986, *Delineation of a Karst Groundwater Basin in Sinking Valley, Pulaski County, Kentucky*: M.S. thesis (unpublished), Eastern Kentucky University, Richmond, 93 p.

Scalf, M.R.; Dunlap, W.J.; and Kreissl, J.F., 1977, *Environmental Effects of Septic Tank Systems*: EPA - 60013-77-096, 31 p.

Wadleigh, C.H., 1968, Wastes in relation to agriculture and forestry: *U.S. Department of Agriculture Miscellaneous Publication 1065*, 23 p.

White, W.B., 1969, Conceptual models for carbonate aquifers: *Ground Water*, v. 7, p. 15-21.

Preliminary Assessment of the Impact of Class V Injection Wells on Karst Groundwaters

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ABSTRACT

The U. S. Environmental Protection Agency has contracted the authors to investigate the impact of Class V injection wells on groundwater in karst terranes to determine the need for regulations over these shallow injection methods. Any sinkhole that has been modified to better accept drainage, including storm-water runoff, is considered a Class V injection well. In addition, service-station-bay drains that lead to septic tanks, pits, or dry wells are a focus of the study.

The study is being conducted in Cookeville and in Johnson City, Tennessee to compare contaminant transport in both flat-lying and folded rock. To date, over 75 service stations have been surveyed as to bay-drain-disposal methods. Sampling sites have been chosen to document potential impacts. Groundwater flow paths and velocities have been determined using dye-tracing methods. Major cave systems underlie both Cookeville and Johnson City. Both cities have modified sinkholes to better accept storm-water runoff. Initial sampling of some springs has shown a significant degradation of water quality and an impact on the diversity and density of benthic macro-invertebrates.

An additional objective of the study is to review the Class V injection-well regulations of eastern states containing karst and to formulate a model regulation to protect karst groundwater. This review has shown that most states do not provide for specific protection of vulnerable karst waters. In fact, several states containing large karst regions have no Class V injection program at all.

Introduction

Part C of the Safe Drinking Water Act (Public Law 93-523) authorizes the U.S. Environmental Protection Agency (EPA) to establish regulations to assure that potable groundwater is not endangered by underground injection of waste. Guidelines for underground injection and the classification of injection wells come under Part 146.04 of the Federal Underground Injection Control (UIC) program (U.S. EPA, 1981). This program created five classes of underground injection wells. Classes I through IV include such categories as radioactive and hazardous waste injection and disposal of brines from the oil and gas industry.

Class V wells are generally defined as those which inject only non-hazardous fluids into or above strata that contain an underground source of drinking water (USDW). USDWs not only include aquifers that are currently serving as drinking-water sources but also aquifers which are of acceptable quality for possible future use. Class V wells include any type of injection well not covered in the UIC definition of Classes I, II, III or IV. EPA has classified Class V injection wells into six groups based in part on the expected quality of the injected fluid. The following is a listing of these groups as found in regulations for the State of Florida which is particularly sensitive to karst aquifers:

Group 1 - wells associated with thermal energy exchange processes, that include air-conditioning return-flow wells and cooling-water return-flow wells. Cooling-water return-flow wells may be part of a closed-loop system, with no hazardous additives.

Group 2 - recharge wells, saltwater-intrusion barrier wells, connector wells, and subsidence-control wells (associated with aquifer overpumping).

Group 3 - wells that are part of domestic waste-treatment systems, swimming-pool drainage wells, injection wells used in experimental technologies, wells used to inject spent brine into the same formation from which it was withdrawn after extraction of halogens or their salts.

Group 4 - non-hazardous industrial and commercial disposal wells, that include laundry waste, dry wells, sand-backfill wells, and nuclear disposal wells used to inject radioactive wastes (provided the concentrations of the waste do not exceed drinking water standards contained in Chapter 17-22, FAC) and injection wells used for *in situ* recovery of phosphate, uraniferous sandstone, clay, sand, and other minerals extracted by the borehole-slurry mining method (lignite, tar sands, oil shale, coal).

Group 5 - lake-level drainage and storm-water drainage wells.

Group 6 - geothermal wells and "other" wells.

The U.S. EPA has funded twenty-four projects nationwide under its Shallow Injection Well Initiatives Program to help evaluate the impact of Class V injection wells on groundwater and to establish best management practices. The authors of this paper were chosen to help evaluate Class V injection-well practices in the karst of Tennessee. Any sinkhole that has been modified to accept waste, including storm-water runoff, is considered a Class V injection well.

Objectives

The primary goal of the proposed study is to determine if Class V injection wells have created a groundwater pollution problem. Evaluating the effects of service-station-bay drains (Group 6) that lead to septic tanks, pits, or dry wells are a primary target of the investigation. Other high-priority Class V injection wells that are being investigated include:

1) Agricultural and municipal drainage into improved and unimproved sinkholes (Group 1).

2) Industrial drainage into sinkholes (Group 4).

3) Domestic wastewater drainage into sinkholes (Group 5).

Additional objectives of the research are:

1) Review case histories of contamination from Class V injection wells in karst through a literature review.

2) Evaluate the effectiveness and applicability of UIC-Class V injection-well regulations in states having karstic limestone terranes and provide suggestions to modify regulations for better protection of human health and the environment.

3) Sample selected runoff waters entering sinkholes and analyze chemical constituents in nearby wells and springs to determine contaminant levels. Analyses for benzene-toluene-ethylene-xylene (BTEX), methyl tertiary butyl ether (MTBE), total petroleum hydrocarbon (TPHC), ethylene glycol., zinc, and lead will help evaluate the impact from service-station-bay drains and urban storm-water runoff.

4) Perform a benthic macro-invertebrate study at sites being sampled chemically to document effects of contaminants on biota in streams.

5) Demonstrate the applicability of dye tracing to show the connection between selected sources of pollution and contaminated wells and springs.

Hydrogeology of the Study Sites

Two geologically different karst terranes in Tennessee were chosen so that the results would have applicability throughout much of the karst in the eastern United States (*see* Figure 1 of Ogden and others, this volume, p. 199). Around Cookeville, the Mississippian-aged carbonates of the Eastern Highland Rim Province are flat-lying, allowing groundwater to move along a wide range of orientations corresponding to joint and photo-lineament trends (Ogden and others, 1989). In the Valley and Ridge Province around Johnson City, the Ordovician-aged carbonates are complexly folded and faulted, and groundwater moves predominantly along stratigraphic strike within solutionally enlarged bedding planes. Most of the groundwater flow paths in and around Cookeville have been documented by Faulkerson and others (1981) and Hannah and others, (1989). These traces have delineated the boundaries of three spring-water basins that receive storm runoff from Class V injection wells. No groundwater tracing had been conducted in Johnson City until initiation of this project. Field investigations to date show that much of the storm drainage in northern Johnson City flows through a cave system and emerges at a spring along Knob Creek.

Methods

Samples were collected where water enters Class V injection wells and at springs influenced by these waters. As a control, samples were also gathered from a sinking stream and a spring with a predominantly forested recharge

area. Samples were analyzed for constituents expected from service stations as well as from agricultural practices. Dissolved oxygen (DO), pH, conductivity, and temperature were measured in the field with a Hydrolab field monitor. Laboratory analyses used as indicators of contamination from agricultural activities and septic tanks included nitrate, chloride, fecal coliform, and fecal streptococcus bacteria. Zinc, chromium, lead, BTEX, MTBE (a gasoline additive), TPHC, and ethylene glycol levels were measured as indicators of waste products and leaks from service stations and parking-lot runoff.

To enhance the interpretation of the impact of Class V injection wells on spring waters, samples of benthic macro-invertebrates were collected at riffles and pools at both the injection points and the springs. Samples were collected with a modified kick net and a Surber sampler (0.09 m^2). All organisms were preserved in a 10% formalin solution, enumerated, and identified to the lowest taxonomic level possible.

Groundwater tracing has been conducted using fluorescein and rhodamine dyes and activated charcoal detectors. Optical brighteners and cotton detectors have also been used for tracing. A control packet was placed at a spring known not to be hydrologically connected to the tracer input site during each test.

Results to Date

Service-Station-Bay Drain Survey

Approximately 75 sites have been visited in the Johnson City and Cookeville areas to determine service-station-bay drain disposal practices. This survey yielded the following results:

1) Most service stations *claim* to have their bay drain connected to the city sewer.
2) Others have no idea where the drain leads to.
3) Only a few say that the drain leads to a septic system or drain well.

These results have made it very difficult to document if degradation in groundwater quality is occurring. Personnel from regulatory agencies with the authority to perform detailed on-site inspections will be needed to determine actual disposal practices. It is anticipated that this survey will be expanded to the Jefferson City - Morristown areas where significant karst occurs within the city limits.

Groundwater Tracing

Four groundwater traces from Class V injection wells in Johnson City have been conducted to date. These results have documented the groundwater flow paths to a spring along Knob Creek that drains much of the northern part of town. Six traces were conducted in the Cookeville area. Groundwater travel times are very rapid and are in the order of several thousand feet per day. Groundwater tracing efforts will continue within these drainage basins.

Water-Quality Sampling

The high costs of analyzing for organic compounds requires detail scrutiny of monitoring locations before sampling begins. At this point in the research, fifteen sites have been sampled around Cookeville and Johnson City. Field parameters such as pH, conductivity, and dissolved oxygen indicate degraded water at springs drained by urban runoff. The samples are presently at the lab awaiting analysis for metals and organics. More sites will be sampled during the present wet season and the data compared to subsequent sampling during the dry season.

Benthic Macro-invertebrate Survey

Many springs and sinking streams will be sampled for the density and diversity of benthic macro-invertebrates before the research is concluded. These sites include springs that drain from Class V injection wells within the city limits, rural springs that drain from forested country-side with some agricultural activity, and waters entering sinkholes that have been modified to accept drainage.

In springs polluted by Class V injection wells in urban settings, the number of benthic macro-invertebrate taxonomic groups was found to be almost half that of springs not affected by injection wells. In the springs polluted by injection wells, only 4 to 5 taxonomic groups of benthic macro-invertebrates were found, whereas, in springs where injection wells are not in the drainage basins, up to 9 taxonomic groups were found (Table 1).

Taxonomic Group	Urban Drained Springs	Rural Drained Springs
Chironomidae	67	42
Ephemeroptera	0	223
Plecoptera	0	98
Coleoptera	4	35
Crustacea	1	45
Gastropoda	0	6
Isopoda	3	107
Nematoda	40	2
Trichoptera	0	89
Total	115	647

Table 1. Benthic macro-invertebrate taxonomic groups and numbers found in urban springs receiving pollution from Class V injection wells versus rural springs not receiving Class V recharge from injection wells.

State Survey of Class V Injection-Well Regulations

A regulatory survey and literature review was conducted to gather information on the scope of groundwater contamination problems associated with Class V injection wells and to compare the regulatory frameworks existing within state programs. This will assist in developing recommendations for improved permitting, monitoring, and control over a variety of shallow injection wells. Karst limestone terranes provide a "worst-case" setting for this study because the underlying aquifer systems are extremely vulnerable to pollution by surface runoff and waste-disposal practices.

A telephone survey of states with significant occurrences of karst limestone bedrock was undertaken in order to:

- communicate with regulatory staff responsible for underground injection control (UIC) or groundwater protection programs,
- obtain copies of the states' UIC regulations pertaining to Class V wells,
- document case histories of contamination from Class V wells, and
- gain ideas and information on alternatives for preventing groundwater contamination by Class V injection wells.

Regulatory agencies in sixteen states have been interviewed, including the Alabama Department of Environmental Management, Arkansas Department of Pollution Control and Ecology, Florida Department of Environmental Regulation (DER), Georgia Department of Natural Resources, Georgia Geological Survey, Illinois Department of Human and Natural Resources (DHNR), Indiana Department of Environmental Management, Kentucky Division of Groundwater Protection, Minnesota Pollution Control Agency (MPCA), Missouri Department of Natural Resources, New York Department of Environmental Conservation (DEC), Ohio Environmental Protection Agency, Pennsylvania Department of Environmental Resources, Tennessee Department of Health and Environment (TDHE), Texas Water Commission, Virginia Water Control Board, and the West Virginia Division of Natural Resources.

Ten of the sixteen states interviewed reported having no rules dedicated to Class V wells. Most of these do, however, give special approval to groundwater return-flow wells for heat-pump and air-conditioning systems. Several commented that misused or abandoned Class V wells are regularly discovered during investigations of unrelated complaints and violations. An Indiana regulator stated that so many different branches come in contact with Class V wells, that the reports are scattered and formal rules had never been organized. He advocated the implementation of a permitting system with permit fees financing inspections and enforcement.

Minnesota prohibits any use of wells for injection purposes (except return-flow wells) and requires landowners to seal them as they are discovered. Investigators have found a number of small-quantity generators illegally disposing of hazardous wastes in septic tanks and shallow wells. The MPCA also cited several instances of Class V wells being used for sewage disposal (cesspools). Investigators of underground storage tanks often find Class V wells being used inappropriately when conducting site investigations at service stations and auto-maintenance shops. The MPCA attempts to inform the public about "disposal wells" and contamination of groundwater by distributing newsletters to the agricultural and business communities.

Illinois is another state that uses a public-information campaign to control Class V injection wells. The DHNR maintains "good relationships" with county Farm Bureau associations and soil-conservation districts. These groups encourage members of the agricultural community to divert feedlot, fertilizer, and other polluted runoff away from wells and sinkholes. DHNR maintains a registry of dry wells used for flood control that is updated regularly with a questionnaire sent out to developers, municipalities, contractors, and others. The use of retention basins to pretreat stormwaters and allow gradual percolation/evaporation is common in Illinois.

The Missouri Division of Geology and Land Survey issues a joint permit with the Division of Environmental Quality for large commercial or institutional heat-pump withdrawal/injection wells. The construction and operation permit is designed to control the temperature differential at a radius of 200 feet from the well. Observation wells are installed at this distance for the purpose of recording monthly temperatures and total dissolved solids. The permit also requires random inspections and an annual report by the permittee. No other Class V wells are approved in Missouri except those constructed for remedial purposes.

Ohio and Arkansas have conducted inventories of Class V wells in their states for the purpose of developing appropriate regulations. Ohio will have regulations available in 1991 that will contain special provisions for karst terranes.

The states of Alabama and Tennessee have a registration requirement for Class V injection wells. The regulators acknowledge that, with limited tracking and enforcement capability, they are notified of only a small fraction of the disposal wells in use. Regulators determine if there is a possibility of adverse impact on an underground source of drinking water based on the type of fluid injection and

other general information submitted. Class V wells are exempt from siting and construction requirements, although TDHE reserves the right to disapprove any well that is considered to be of substandard construction. Alabama requires new Class V wells to be drilled by a licensed well driller and issues a ten-year operating permit. TDHE requires owners/operators to apply for a Plugging and Abandonment Permit that justifies well abandonment and outlines a closure plan using a cement plug.

Georgia, Florida, and West Virginia have the most comprehensive regulations covering Class V wells that have been reviewed to date. In West Virginia, owners/operators of Class V wells must notify the Chief of the Water Resources Division of the following information: construction features of the well, nature and volume of injected fluids, alternative means of disposal available, environmental and economic impacts of well disposal and its alternatives, facility name, location, ownership, legal contact, nature and type of injection well, and operating status. All Class V wells are authorized by rule for a period of five years. When the Chief discerns a potential violation of drinking-water standards, an individual permit can be issued. The permits may include conditions for operating, mechanical integrity, monitoring, reporting, and plugging and abandonment.

Before Class V wells are approved in West Virginia, regulations specify that hydraulic connections with underground sources of drinking water should be considered in addition to the potentially affected population. The area of concern is called the "zone of endangering influence", which is defined as the "...horizontal distance from the injection well in which the pressures in the injection zone may cause the migration..." of contaminants into an USDW.

The regulations suggest two possible methods for determining the zone of endangering influence. The first takes an analytical approach using the Theis equation, based on the following assumptions.

- the injection zone is homogenous and isotropic,
- the well penetrates the entire thickness of the injection zone,
- the well diameter is insignificant compared to the radius of influence when injection time is longer than a few minutes, and
- emplacement of fluid into the injection zone creates an instantaneous increase in pressure.

Because such idealized aquifer characteristics occur very rarely in the real world, the regulations also allow a more qualitative approach for determining the zone of endangering influence. A fixed radius of not less than 1/4 mile can be used for a well or a close cluster of wells based on the following considerations:

- chemistry of the injection and formation fluids,
- geology and hydrogeology,
- population, groundwater use and dependence, and
- historical practices in the area.

Regulations in West Virginia specifically mention two types of Class V wells often found in carbonate formations:

- wells for waste disposal into solution cavities, and
- sinkholes used for the disposal of sewage or any other waste.

Alabama, Florida, Georgia and West Virginia require that Class V wells be constructed by water-well contractors licensed within the respective states. None offer specific construction-design standards owing to the variety of wells and their uses. Florida, however, reserves the right to apply Class I design standards if the situation warrants. The DER also requires a well-completion report and may ask for samples of formations penetrated. Georgia requires casing five feet into the injection zone and grouting/sealing of the annular space.

The DER can impose operating, pretreatment, monitoring, and reporting requirements on cooling-water return-flow wells with additives, experimental/remedial wells, spent-brine return-flow wells, non-hazardous commercial/industrial disposal wells, geothermal wells, and "other" wells. The latter three must be plugged with cement when abandoned to prevent the movement of fluids between USDWs.

One concern aroused during the interviews was that the recently issued NPEDS municipal storm-water regulations will increase the use of Class V injection wells for controlling urban runoff. With many states having such loose regulatory control over Class V wells, there are implications for impacting groundwater quality with re-routed storm-water drainage.

The New York DEC has begun to address this problem with its State Pollutant Discharge Elimination System. If a facility has a discharge, the Division of Water Resources must see that the discharge meets minimum standards and that groundwater under the site meets designated standards. Detention-basin discharges draining large parking areas are required to have no visible oil and grease, or else an oil/water separator will be required. The regulator stated that non-discharging collection basins that allow storm water to percolate/ evaporate are not monitored except in heavily industrialized or potentially contaminated areas.

During the interview, the regulators seem interested in developing practical alternatives for managing Class V

injection wells, and several requested copies of our final report. Most noted the wide range of locations, uses, and types of Class V wells as being the main barrier to a uniform regulatory program.

Summary and Conclusions

Our preliminary results have shown that many more Class V injection wells occur in the study areas than anticipated. Initial sampling shows that degraded surficial recharge to these wells adversely impacts groundwater. Regulatory authority will be needed to assess the impact of individual service-station-bay drains on groundwater quality.

In addition to the regulations from the states, we have compiled a number of case histories and local/model ordinances from wellhead-protection programs, watershed districts, and areas with sole-source aquifer designation. These will be evaluated and developed into recommended guidelines and "best-management practices" for preventing contamination by Class V injection wells.

Some combination of the West Virginia, Florida, and Georgia UIC regulations would provide a relatively flexible framework for regulating Class V injection wells. It is anticipated that the use of Class V injection wells for storm-water drainage and other types of disposal will increase in rapidly developing urban areas and even in some smaller communities. The importance of evaluating and monitoring the impact of these wells on USDWs should be emphasized. We suggest that dye-tracing techniques would be applicable in karst terranes to delineate the potential zone of impact. Other factors to consider would include the density of disposal wells in the area, the time of travel between the injection point and the nearest discharge or withdrawal point, other potentially impacting landuses, possible chemical interactions among injected fluids, and anticipated alterations to the subsurface flow systems caused by additional volumes of fluids.

Acknowledgments

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References

Faulkerson, J.; Burden, D.; Burden, K.; Edwards, C.; Kinley, T.; Lee, T.; Sparks, V.; Starnes, D.D.; Walls, E.; and Webster, S., 1981, *Karst Hydrology, Morphology, and Water Quality in the Vicinity of Cookeville, Tennessee*: Report to the City of Cookeville, Tennessee Technological University, Cookeville, Tennessee, 67 p.

Hannah, E.D.; Pride, T.E.; Ogden, A.E.; and Paylor, R., 1989, Assessing groundwater flow paths from pollution sources in the karst of Putnam County, Tennessee, *in* Beck, B.F. (editor), *Engineering and Environmental Impacts of Sinkholes and Karst: Proceedings of the Third Multidisciplinary Conference on Sinkholes and the Engineering and Environmental Impacts of Karst, St Petersburg Beach, Florida, 2-4 October 1989*: A.A. Balkema, Rotterdam and Brookfield, p. 183-188.

Ogden, A.E.; Curry, W.A.; and Cummings, J.L., 1989, Morphometric analysis of sinkholes and caves in Tennessee comparing the Eastern Highland Rim and the Valley and Ridge physiographic provinces, *in* Beck, B.F. (editor), *Engineering and Environmental Impacts of Sinkholes and Karst: Proceedings of the Third Multidisciplinary Conference on Sinkholes and the Engineering and Environmental Impacts of Karst, St Petersburg Beach, Florida, 2-4 October 1989*: A.A. Balkema, Rotterdam and Brookfield, p. 135-142.

U.S. Environmental Protection Agency, 1981, 40 CFR Parts 122 and 146 Underground Injection Control Program criteria and standards: *Federal Register*, v. 46, no. 190, p. 48243-48255.

The Carbonate Aquifer of the Northern Shenandoah Valley of Virginia and West Virginia

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ABSTRACT

A Cambrian and Ordovician carbonate sequence over 12,000 feet (3,700 m) thick is exposed west of the Blue Ridge in the northern Shenandoah Valley of Virginia and West Virginia. The rocks are highly folded and faulted and the nearly vertical attitude of the rocks tends to force groundwater to follow more permeable bedding planes parallel to the stratigraphic strike. Dye-tracer tests have generally shown a half-radial flow pattern with travel times of three to five months for distances of one to three miles (1.6 to 4.8 km). Karstification of the land surface and the aquifer is subdued due to the high-density fracture system, that tends to disperse rather than concentrate groundwater flow, and the generally low hydraulic gradients ($dh/dl \approx 0.01$). Groundwater circulation may dip well below local base level and solution cavities have been observed at depths of 150 feet (45 m) below major surface streams. Permeability is only through secondary fractures although the groundwater appears to be a continuum; so the aquifer is termed "fracture-diffuse". Both tracer-test and pumping-test results show that permeability is almost one order of magnitude higher parallel to the strike than normal to the strike. However, potentiometric contours tend to parallel the strike, so the hydraulic gradient is normal to the strike. The almost vertically dipping rock and the ready communication of surface sinkholes with groundwater suggest that the aquifer is generally unconfined. However, most wells are drilled to depths greater than 100 feet (30 m) in spite of relatively shallow, (average about 30 feet or 10 m) static water levels in the region. Storativity values calculated from several multiwell pumping tests range from 0.01 to 0.001 and suggest confined flow conditions in the deeper fractures despite the absence of an obvious confining layer.

Introduction

The northern part of the Shenandoah Valley is a broad carbonate lowland bounded by the precambrian Blue Ridge Mountains to the east and clastic ridges to the west. The study area lies within the Valley and Ridge physiographic province and includes parts of Clarke and Frederick counties in Virginia and Berkeley and Jefferson counties in West Virginia (Figure 1). The carbonate sequence is split into two sections by the Martinsburg Shale which crops out along the Frederick-Clarke and Berkeley-Jefferson county borders. The western carbonate belt is narrower, higher, and generally exhibits more relief than the eastern belt. This paper concentrates on the hydrogeological setting of the eastern carbonate belt which is exposed over about 75% of Clarke and Jefferson counties.

An overview of the karst setting of the northern Shenandoah Valley was presented in maps by Hubbard (1983, 1990). Regional groundwater studies have been conducted by Bieber (1961), Hobba and others (1972),

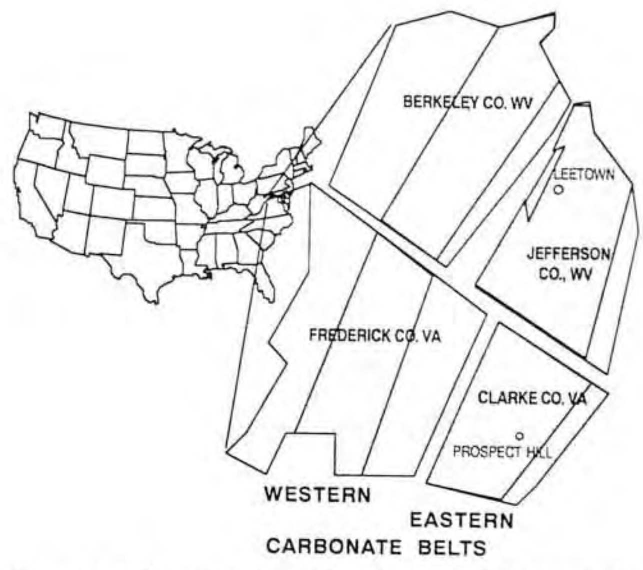

Figure 1: Location map of study area showing major carbonate outcrop belts.

Hobba (1976, 1981), Kozar and others (1991), Trainer and Watkins (1975) and Wright (1991). Detailed site studies were presented by Jones and Deike (1981), Jones (1987), and Jones and Jones (1988). Papers by Hack (1965) and White and White (1974) discuss the geomorphology of the region.

Geologic Setting

The study area lies in the Massanutten Mountain synclinorium and is underlain by Cambrian and Ordovician carbonates. The carbonate sequence is about 12,500 feet (3,800 m) thick and is highly folded and faulted. The east limbs of synclines and the west limbs of anticlines are oversteepened and locally overturned. Many low-angle faults dip to the east. The strike is N 20 E, but some of the fold axes trend in a more northerly direction than the strike of the formational outcrop belts. The role of faults in controlling groundwater movement is not clear. Hobba (1981) felt that faults act as groundwater corridors and that 67% of the large springs in Jefferson County were located on or near faults. However, detailed site investigations (Jones and Deike, 1981) suggest that most of the fault zones are well cemented with calcite and probably act as a partial barrier to groundwater. This could force groundwater to the surface at some points and may result in increased solutional activity on the upgradient side of the fault.

The carbonate sequence is a series of massive limestones, thin, shaley limestones, dolostones, and occasional bands of clastic, shaley sandstones (Figure 2). The highest water yields are from the Stonehenge (Beekmantown) and Conococheague formations. The Beekmantown Group also contains the highest density of mapped sinkholes, with 5 sinkholes/square mile in Jefferson County (Kozar and others, 1991).

The development of karst features is generally subdued in this region. The most obvious karst features are exposed ridges of carbonate rocks trending parallel to the strike and numerous large springs. These springs have mean flows from 1 to 4 cfs (0.028 to 0.113 cms) and show little response to individual storm events. Sinkholes tend to be small and shallow, and most caves in the eastern belt are small, isolated features. Berkeley County has 50 caves (3 over 1000 feet long), Jefferson County has 42 caves (3 over 1000 feet long), Frederick County has 17 caves (1 over 1000 feet long), and Clarke County has 5 caves (none over 100 feet long). The more significant karst features tend to develop near streams (Hack, 1965). This may be especially pronounced near entrenched streams where the hydrologic gradient is steepened and groundwater flow rates are accelerated.

Conduit development is minimal in the eastern carbonate belt. Only part of the dye recovery from one water-tracing test, out of fourteen tests conducted to date, exhibited the travel times and recovery pattern associated with conduit flow. The results from three tracer tests in the western belt in Berkeley County were more charac-

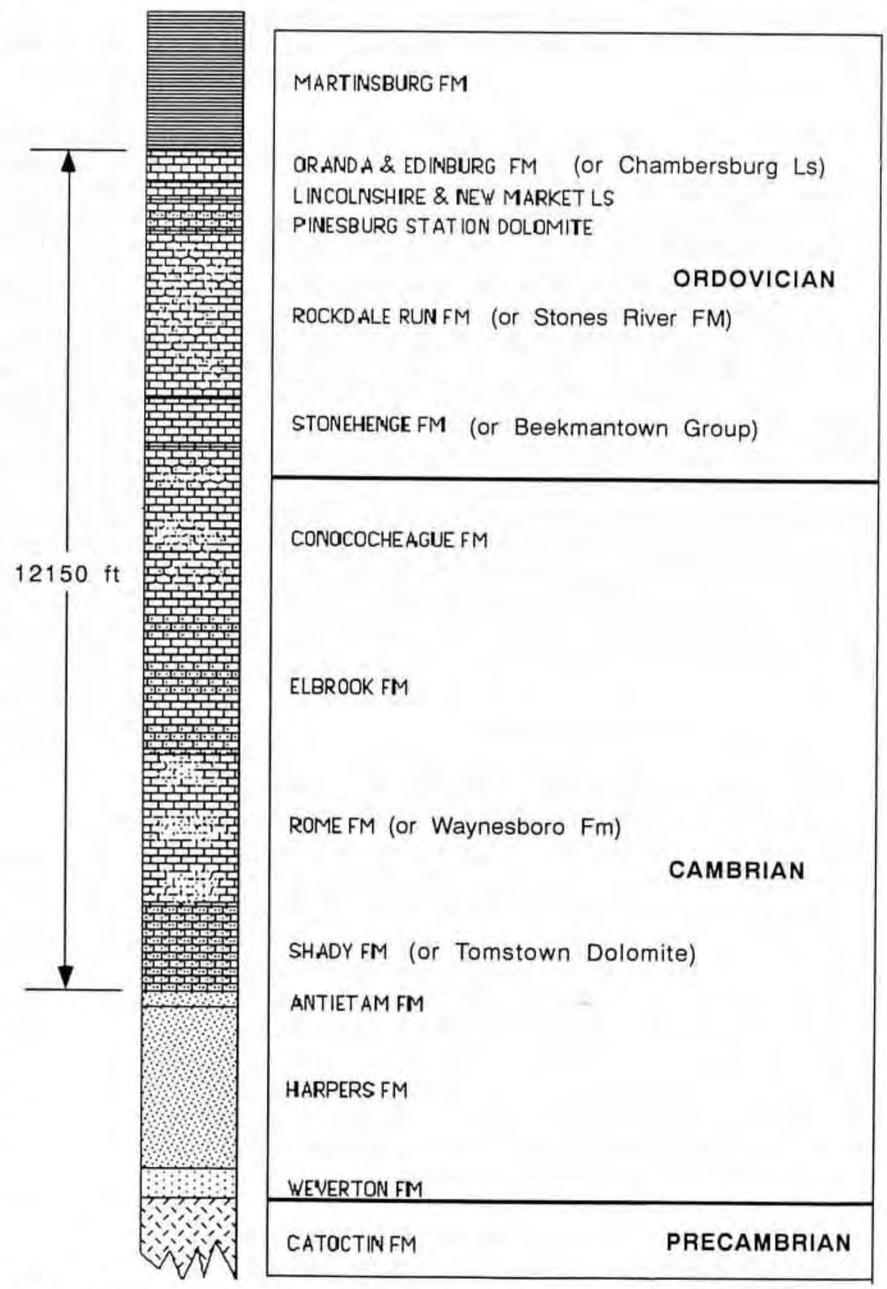

Figure 2: Generalized geologic column for the northern Shenandoah Valley.

teristic of conduit-flow tests, but a diffuse-flow component was still present. The general lack of conduit-flow development is attributed to a combination of: (1) High density fracture system which tends to disperse rather than concentrate groundwater flow; (2) Generally low relief and low hydraulic gradients ($dh/dl \approx 0.01$); (3) Lack of surrounding non-carbonate highlands to concentrate aggressive water at certain points on the carbonates.

The nearly vertical attitude of the rocks tends to force groundwater to flow in troughs of more permeable units or fractures parallel to the strike. The potentiometric contours also tend to parallel the strike, so groundwater movement predicted orthogonal to the potentiometric contours must move across lithologic barriers of lower permeability (Jones and Jones, 1988). The aquifer has entirely secondary permeability and is strongly anisotropic. Groundwater does appear to be a continuum so the aquifer is termed "fracture-diffuse".

Hydrology

Total annual precipitation for the study area is about 38 inches (965 mm) and potential evapotranspiration is 30 inches (762 mm). Discharge records at Leetown (Jones and Deike, 1981) show 13 inches (330 mm) of runoff so the actual evapotranspiration is probably about 25 inches (635 mm) and groundwater recharge is about 9 inches (230 mm) annually.

Fourteen groundwater-tracer tests have been conducted in the eastern carbonate belt of Jefferson and Clarke counties (Jones and Deike, 1981; Jones, 1987; Kozar and others, 1991). Sinking streams are rare in this area, so the dye was injected into sinkholes and flushed into the aquifer using 1000 to 2000 gallons (3.8 to 7.6 cubic meters) of water from tank trucks. Dye recovery was at multiple springs one to three miles (1.6 to 5 km) from the injection point. Travel times were three to five months with the exception of one test site situated along a fault where part of the dye was recovered in less than two weeks (Kozar and others, 1991).

Movement of the tracer is often parallel to the stratigraphic strike and sometimes almost parallel to the potentiometric contours. The flow pattern is generally half-radial and travel times are not very dependent on straight-line distances between the injection and recovery points (Figure 3). The flow rates do not appear to be linear with respect to time. Dye recovery from most quantitative tests in the area appears to correlate with storm events. Dye recovery lagged precipitation by four to five days during a test to Prospect Hill Spring in Clarke County (Figures 4 and 5).

The assignment of distinct catchment areas to individual springs in the area is probably not practical because of the diffuse nature of groundwater movement. The springs all appear to draw water from a common aquifer, so groundwater divides between springs are not sharp. Dye was also found in pumped wells during several of the tracer tests. The size of the recharge area may be estimated from discharge records and a water-budget calculation (Jones and Deike, 1981; Wright, 1991), but pollutants may arrive at a spring from distances well outside the calculated recharge area.

Groundwater flow appears to approach darcian conditions if a relatively large volume of the aquifer is considered. Velocities are low enough that the flow should be generally laminar, and conduit flow is rare and integrated into the regional potentiometric gradient. Trainer and Watkins (1975) estimated the average transmissivity (T) of the upper Potomac River Basin carbonates at 500 square feet/day (sq ft/d) (46sq m/d) and average storativity at 0.03. Multi-well pumping tests from the Rockdale Run Formation and the Stonehenge Limestone in Maryland gave T values of 17,000 sq ft/d (1580 sq m/d) and 26,700 sq ft/d (2480 sq m/d) and storativities of 0.02 and 0.01 respectively. Kozar and others (1991) estimated

TRACER TEST FROM GILPIN SINK, CLARKE CO., VIRGINIA

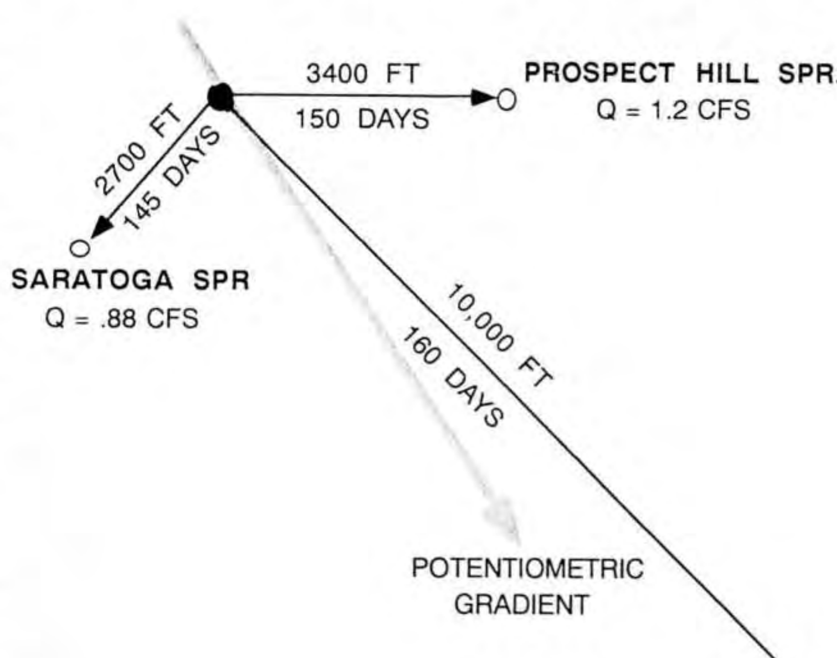

Figure 3: Pattern of dye recovery from the injection of 1 lb of fluorescein sodium dye in Gilpin sinkhole near Boyce, Clarke County, Virginia. The slope of the potentiometric surface is shown by the stippled arrow.

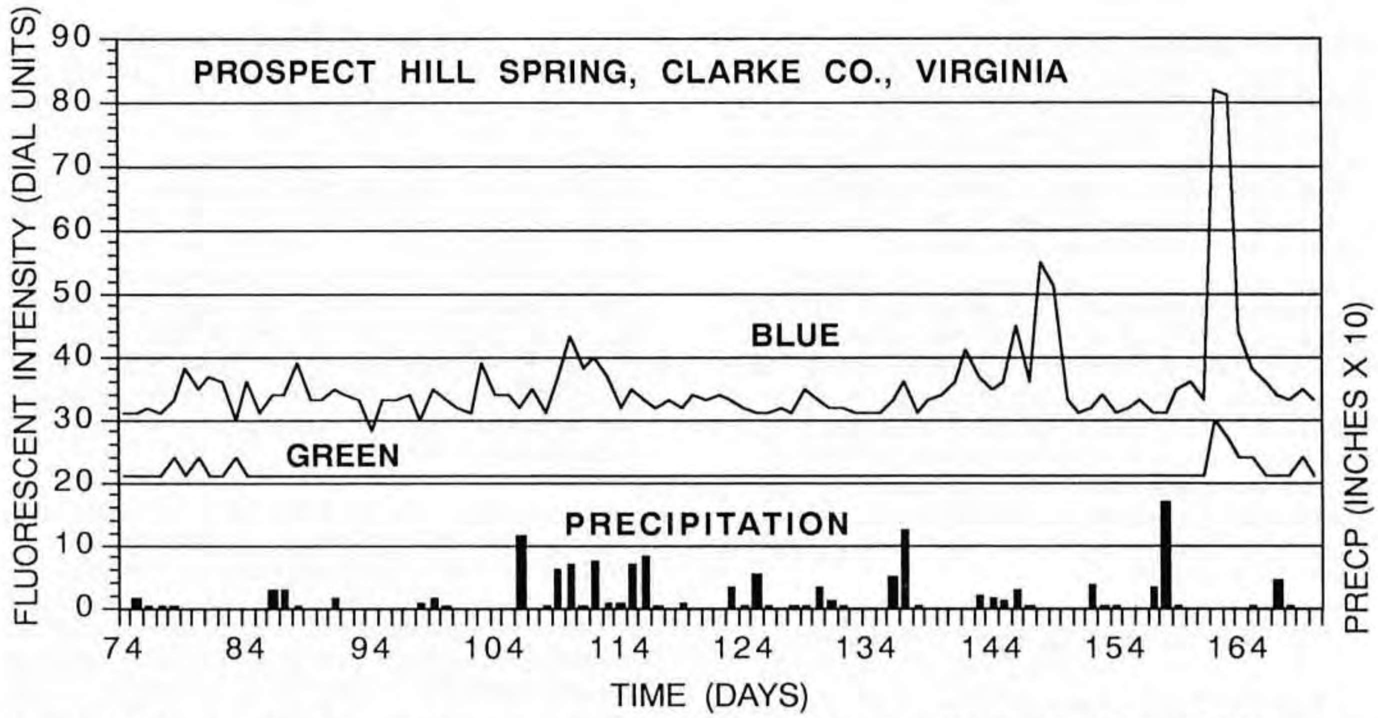

Figure 4: Dye recovery and precipitation at Prospect Hill Spring, Clarke County. The blue dye, Tinopal CBS-X, and the green dye, fluorescein sodium, were injected on the same date, but the blue dye was injected in a sinkhole 2.25 miles west of the spring and the green dye in a sinkhole 0.6 miles west of the spring. Note the long travel times and the earlier arrival of the blue tracer.

Jefferson County transmissivity at 4000 sq ft/d (372 sq m/d) parallel to the strike and at 500 sq ft/d (46 sq m/d) perpendicular to the strike, using the gradient of the water table and stream flow recession. This indicates an isotropy ratio of about 8:1 in the strike direction.

Several multi-well pumping tests have been conducted in the area, but many of the assumptions behind aquifer tests are not met by field conditions. Transmissivity values from these tests seem to range between rather narrow limits, but they are not constant at all times and places (Trainer and Watkins, 1975). The results of aquifer tests in this area are not exactly reproducible and different types of analysis will give somewhat different values from the same test. The validity of the use of the analysis and the degree to which the results are representative of the aquifer must be questioned for all pumping tests in this region, but a certain broad consistency in the results builds some confidence in the applicability of the technique.

A pumping test conducted in the Rockdale Run Formation at Leetown, Jefferson County, is shown in Figure

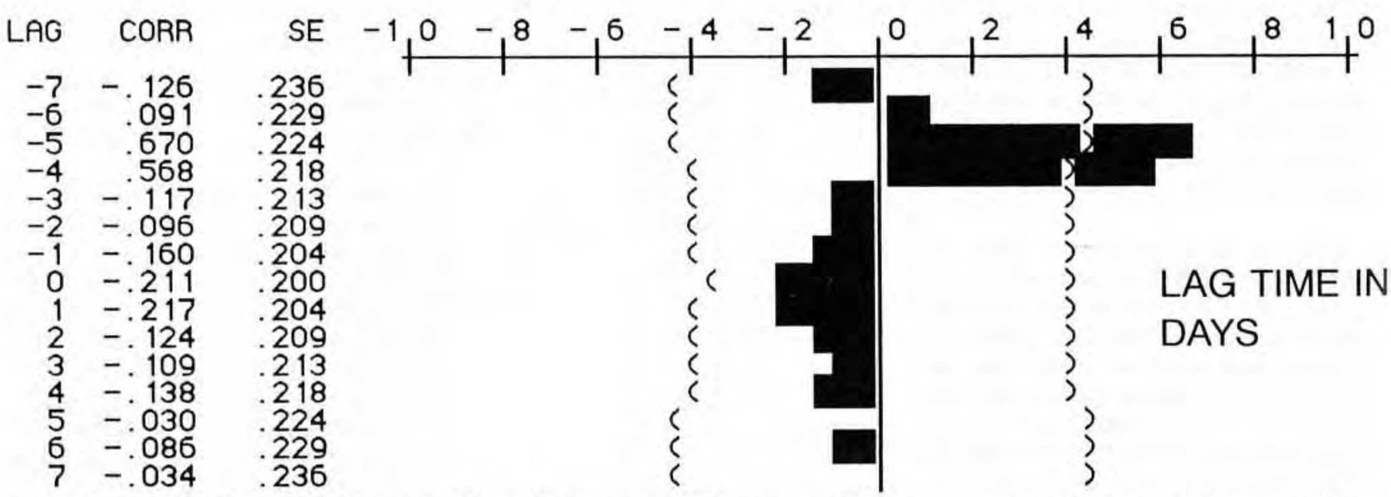

Figure 5: Plot of the cross-correlation of blue dye recovery and storm events at Prospect Hill Spring, Clarke County. Dye-recovery peaks lagged storm peaks by 4 to 5 days during this test.

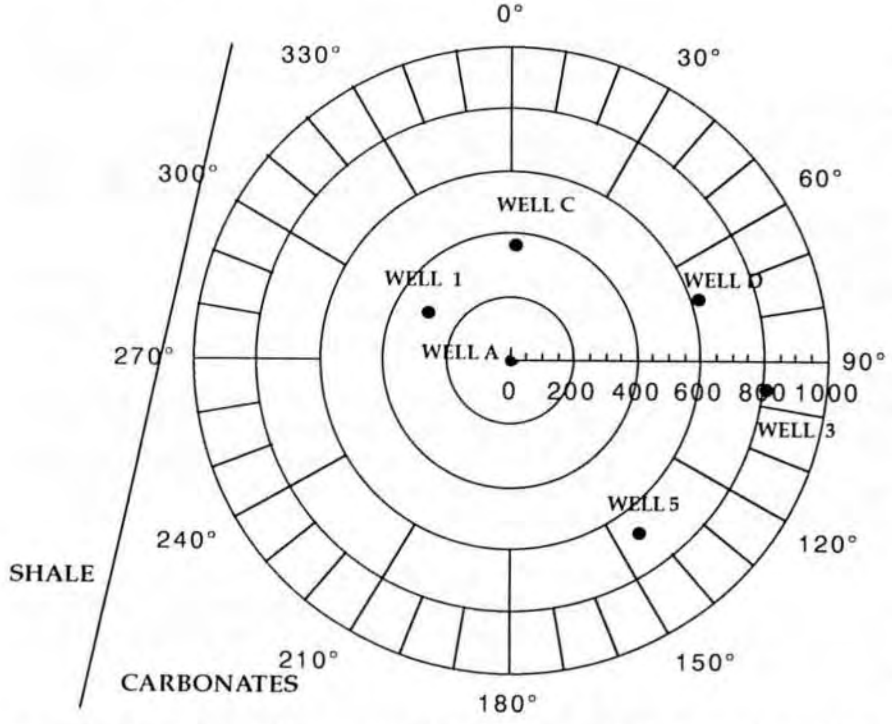

Figure 6: Plan view of arrangement of observation wells and pumped well "A" (center) for pumping test conducted at Leetown, Jefferson County, West Virginia. Note the presence of an impermeable shale boundary west of the test area.

6. Well "A" is 120 feet (37 m) deep and was pumped at 100 gpm (0.38 m^3/min) for 48 hours. The boundary of the carbonates and the Martinsburg Shale is about 1200 feet (366 m) west of the test area. Transmissivities calculated using the Theis nonequilibrium formula were 7280 sq ft/d (676 sq m/d) at well "C" and 15,400 sq ft/d (1430 sq m/d) at well "D". Storativity was 0.001 at "C" and 0.002 at "D". Another calculation method, the Jacob time-drawdown method, is presented in Figure 7 for well "D". This analysis gives a transmissivity of 10,600 sq ft/d (985 sq m/d) and a storativity of 0.03.

One major source of noise may be the assumption that the test is dealing with a confined aquifer. Storativities of confined aquifers are generally less than 0.01, so the values for this region are right on the boundary between confined and unconfined. Also, highly fractured aquifers may have low storativities under water-table conditions. There is no obvious confining bed, although a more open solutionally enlarged zone may exist at some depth. Wells drilled in the area generally go much deeper than the potentiometric surface to obtain an adequate water supply. The aquifer is in ready communication with the surface at sinkholes. In short, the aquifer appears to have both confined and unconfined characteristics. One possible interpretation of some of the irregularities in the pump-test results is that the initial drawdown and release of water from the aquifer is under confined conditions, but water-table conditions develop after some period of time and specific yield begins to control the flow of water to the well. This may also explain why many wells in the area produce turbid water after pumping for 15 or 20 minutes.

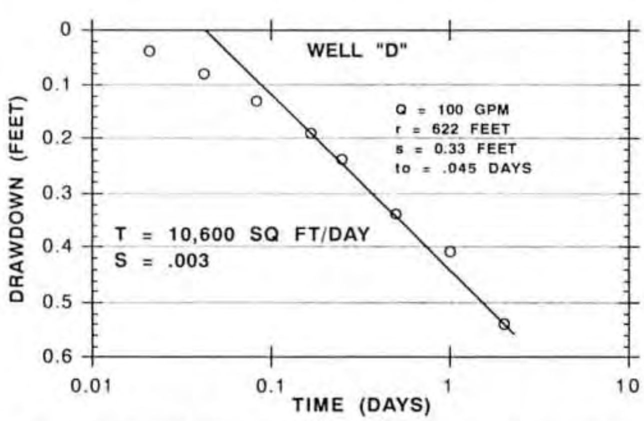

Figure 7: Semi-log plot of drawdown at well "D" versus time, from pump test at Leetown. Transmissivity (T) and storativity (S) calculated using the Jacob method.

Water circulation could theoretically be quite deep in this region owing to the nearly vertical attitude of the rocks. The potentiometric surface averages about 50 feet (15 m) below the land surface and ranges in elevation from 400 to 600 feet (120 to 180 m). Wells drilled deeper than 400 feet (120 m) below land surface do not usually encounter significant yields at depth. A study of base-level controls on cave development in the Potomac River Basin (White and White, 1974), showed that most cave passages in this area are at about 400 feet (120 m) elevation. This is above the level of the entrenched Potomac River and Opequon Creek which is at about 380 feet (115 m). Karstification does occur at depths well below present base level, for a well at the Veterans' Administration Hospital in Berkeley County encountered a 6-foot (2-m) -high solution cavity at an elevation of 227 feet (70 m), or 273 feet (83 m) below land surface.

Conclusions

The carbonate aquifers of the northern Shenandoah Valley yield large quantities of water and share characteristics of both karst and diffuse-flow aquifers. The upper part of the aquifer is in direct communication with sinkholes on the land surface and any substances that enter the aquifer in the recharge areas may be transmitted through a large volume of the aquifer in a matter of months. The carbonate aquifer of this region should be considered a "sole-source aquifer". This is a valuable resource that should be protected.

References

Beiber, P.B., 1961, Groundwater features of Berkeley and Jefferson Counties, West Virginia: *West Virginia Geological Survey Bulletin 21*, 79 p.

Hack, J.T., 1965, Geomorphology of the Shenandoah Valley, Virginia and West Virginia, and origin of the residual ore deposits: *U.S. Geological Survey Professional Paper 484*, 84 p.

Hobba, W.A., Jr., 1976, Groundwater hydrology of Berkeley County, West Virginia: *West Virginia Geological Survey Environmental Geologic Bulletin 13*, 21 p.

Hobba, W.A., Jr., 1981, Groundwater hydrology of Jefferson County, West Virginia: *West Virginia Geological Survey Environmental Geologic Bulletin 16*, 21 p.

Hobba, W.A., Jr.; Friel, E.A.; and Chisholm, J.L., 1972, Water resources of the Potomac River Basin, West Virginia: *West Virginia Geological Survey, River Basin Bulletin 3*, 110 p.

Hubbard, D.A., Jr., 1983, Selected karst features of the northern Valley and Ridge province, Virginia: *Virginia Division of Mineral Resources Publication 44*, one sheet.

Hubbard, D.A., Jr., 1990, Geologic and hydrologic components maps for Clarke County, Virginia: *Virginia Division Mineral Resources Publication 102*, 2 sheets.

Jones, W.K., 1987, Overview of the groundwater resources of Clarke County, Virginia, with emphasis on the carbonate aquifers west of the Shenandoah River: *Clarke County Ground Water Protection Plan*, p. 7.1 - 7.22.

Jones, W.K. and Deike, G.H., III, 1981, *A Hydrologic Study of the Watershed of the National Fisheries Center at Leetown, West Virginia*: Environmental Data, Frankford, West Virginia, 74 p.

Jones, W.K. and Jones, L.E., 1988, The study of groundwater movement in highly fractured aquifers: *Proceedings of 21st IAH Congress, Guilin, China*, p. 620.

Kozar, M.D.; Hobba, W.A., Jr.; and Macy, J.A., 1991, Geohydrology and water quality of Jefferson County, West Virginia, with emphasis on the carbonate area: *U.S. Geological Survey Water Resources Investigations Report* (open file).

Trainer, F.W. and Watkins, F.A., Jr., 1975, Geohydrologic reconnaissance of the upper Potomac River Basin: *U.S. Geological Survey Water-Supply Paper 2035*, 68 p.

White, W.B. and White, E.L., 1974, Base-level control of underground drainage in the Potomac River Basin, *in* Rauch, H.W. and Werner, E. (editors), *Proceedings of the 4th Conference on Karst Geology and Hydrology*: West Virginia Geological Survey, p. 41-53.

Wright, W.G., 1991, Groundwater hydrology and quality in the Valley and Ridge physiographic provinces of Clarke County, Virginia: *U.S. Geological Survey Water Resources Investigations Report 90-4134*, (open file).

Influence of Hydrogeologic Setting and Lineaments on Water-Well Yield in the Great Valley Karst Terrane of Eastern West Virginia

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ABSTRACT

A study of 419 water wells was done in the Cambrian-Ordovician carbonate rock terrane of Berkeley and Jefferson counties in West Virginia, to determine how hydrogeologic factors influence well yield. The study area covers about 280 square miles and is dominated by the Massanutten Synclinorium that contains many tight folds and faults. Drillers' well yields in gallons per minute were utilized from various sources.

Topography shows a very strong correlation with well yield, as expected. Yields from valley wells are the greatest (25 gpm, median yield), compared to slope and upland wells (15 gpm, median) and hilltop wells (8 gpm, median).

The carbonate terrane was subdivided into poorly cavernous (PC) and moderately cavernous (MC) rock types. Two cave indices were derived from lengths of caves over 50 ft long with natural entrances: the cave density (CD) index (total cave length / carbonate rock area), and the average cave length (ACL) index. PC rocks were judged to have an ACL index of less than 250 ft and a CD index of less than 75 ft/mi^2. MC rocks have higher index values. The median yield of PC rocks (15 gpm) is significantly higher than that for MC rocks (12 gpm).

MC rocks show no significant correlation between lineaments or faults and well yield. PC rocks exhibit strong well yield with short lineaments, averaging 33.5 gpm within 0.05 km of such lineaments, or 2.8 times the yield of more distant wells. These strong trends are evident for all topographic settings except hilltops. PC-rock wells, each located within 0.10 km of two short lineaments, average about 125 gpm, or about eight times greater than non-lineament wells. Wells in PC units were also found to have significantly higher yields within 0.20 km of a thrust fault. In summary, short lineaments are well suited for groundwater exploration in PC-carbonate aquifers.

Introduction

The main purpose of this research is to determine how hydrogeologic setting, and especially surficial lineaments, relate to water-well yield, with the intention of improving groundwater exploration in the study area.

The study area is located in the Valley and Ridge Province within the eastern panhandle of West Virginia, including all of Berkeley and Jefferson counties, as shown in Figure 1. The carbonate units range in age from the Cambrian Tomstown dolomite to the Devonian Helderberg limestones. The Cambrian-Ordovician carbonates, composed of limestone and dolostone units, represent approximately 95% of the total carbonate area and are situated within the Great Valley region, bounded by North Mountain on the west and the Blue Ridge on the east. The Great Valley is structurally dominated by the Massanutten Synclinorium, where bedrock is intensely faulted with many minor folds that trend approximately N 20 E.

The hydrology and hydrogeology within the study area have been described by Bieber (1961) and Hobba (1976, 1981). The carbonate units are dense, causing groundwater to flow primarily through secondary openings such as fractures, joints, bedding partings, and solutionally widened

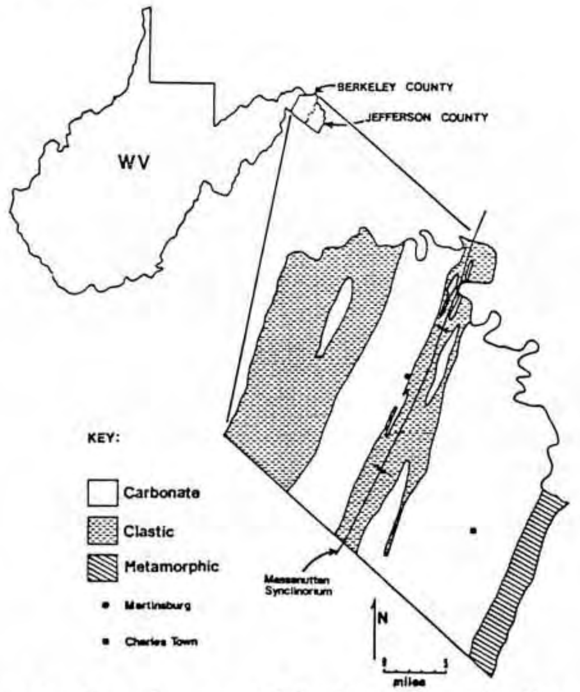

Figure 1: Location map of the Berkeley and Jefferson counties study area within West Virginia.

openings. These secondary features not only determine flow direction but also influence whether a well will produce adequate quantities of water.

Previous Investigations

The relationship between well yield and lineament proximity in carbonate rocks was originally described by Lattman and Parizek (1964) within the Valley and Ridge rocks of central Pennsylvania. They reported increased water-well yields near fracture traces (short lineaments) associated with enhanced solutional and fracture porosity and permeability along fracture zones. Siddiqui and Parizek (1971) elaborated on this lineament study, investigating hydrogeologic factors that influence well yields. They concluded that the controlling factors on well productivity are fracture traces, rock type, dip of the bedrock, topography, and structural setting, in order of decreasing importance.

LaRiccia and Rauch (1977) studied the effect of photolineaments on well yield in the carbonate rock of Frederick Valley, Maryland. They found significantly higher yields where wells are located within 200 ft and especially within 100 ft of a photolineament (short or long straight lineaments).

Rauch and Plitnik (1984) studied the effect of lineaments on well yield in the Cambrian-Ordovician rocks of Hagerstown Valley, Maryland, located adjacent to the present study area. Most carbonate units there are poorly cavernous (diffuse flow in character), and exhibit high yields to wells near lineaments, but one highly cavernous unit shows no relationship with well yield. Ogden (1976), Heller (1980), and LaRiccia and Rauch (1977) have also found that highly cavernous (conduit) aquifer systems are areas where lineaments are not useful as exploration tools for groundwater.

Methods

Lineaments were defined as being linear (straight lineament) or uniformly curved (curvilineament) features identified as stream segments, dry valleys, aligned sinkholes, aligned meander bends in channels, or tonal streaks in soil. Both short lineaments (0.5 - 2.0 km long) and long lineaments (>2.0 km long) were mapped. Lineament mapping proceeded in two phases; first lineaments were mapped based solely on topographic contours and streams shown on the 7.5-minute topographic quadrangles, and then lineaments were mapped from black-and-white aerial photos with a scale of 1:20,000. The LANDSAT lineaments used in this investigation were mapped by Hobba (1976, 1981), and have lengths ranging from 9 to 19 miles. Figure 2, part of a 7.5-minute topographic quadrangle that was used as a base map, shows examples of most types of mapped lineaments.

Well-yield data were obtained for 419 wells situated within the Cambrian-Ordovician units. About 20 percent of these data were gathered from the U.S. Geological Survey reports of Bieber (1961) and Hobba (1976, 1981), and about 80 percent of the well-yield data were derived from drillers' permit reports for 1984-1989 that are available at the county health-department offices. All well locations designated on the drillers' permits were first field checked for accuracy and then plotted on the topographic base maps. All well plotting was done only after all lineaments were identified, to eliminate biases in lineament mapping that could result from knowing locations of high-yielding wells beforehand.

The lateral distance from each well to the nearest lineament was measured directly on the base map with an accuracy of ±0.01 km. Graphical plots were constructed of well yield versus parameters of the hydrogeologic setting at the wells and of measured lineaments. Apparent visual trends in the data were statistically tested for strength using the nonparametric Mann-Whitney U test as described by Siegel (1956).

Hydrogeologic Influences on Well Yield

Topographic Influences

A topographic position of valley, slope, upland, or hilltop was designated for each well. Valley wells have a median yield of 25 gpm, compared to 15 gpm for slope and upland wells and 8 gpm for hilltop wells. Slope and upland wells were grouped together because their yields are not significantly different. Valley wells have significantly higher yields than do hilltop and slope/upland wells at the

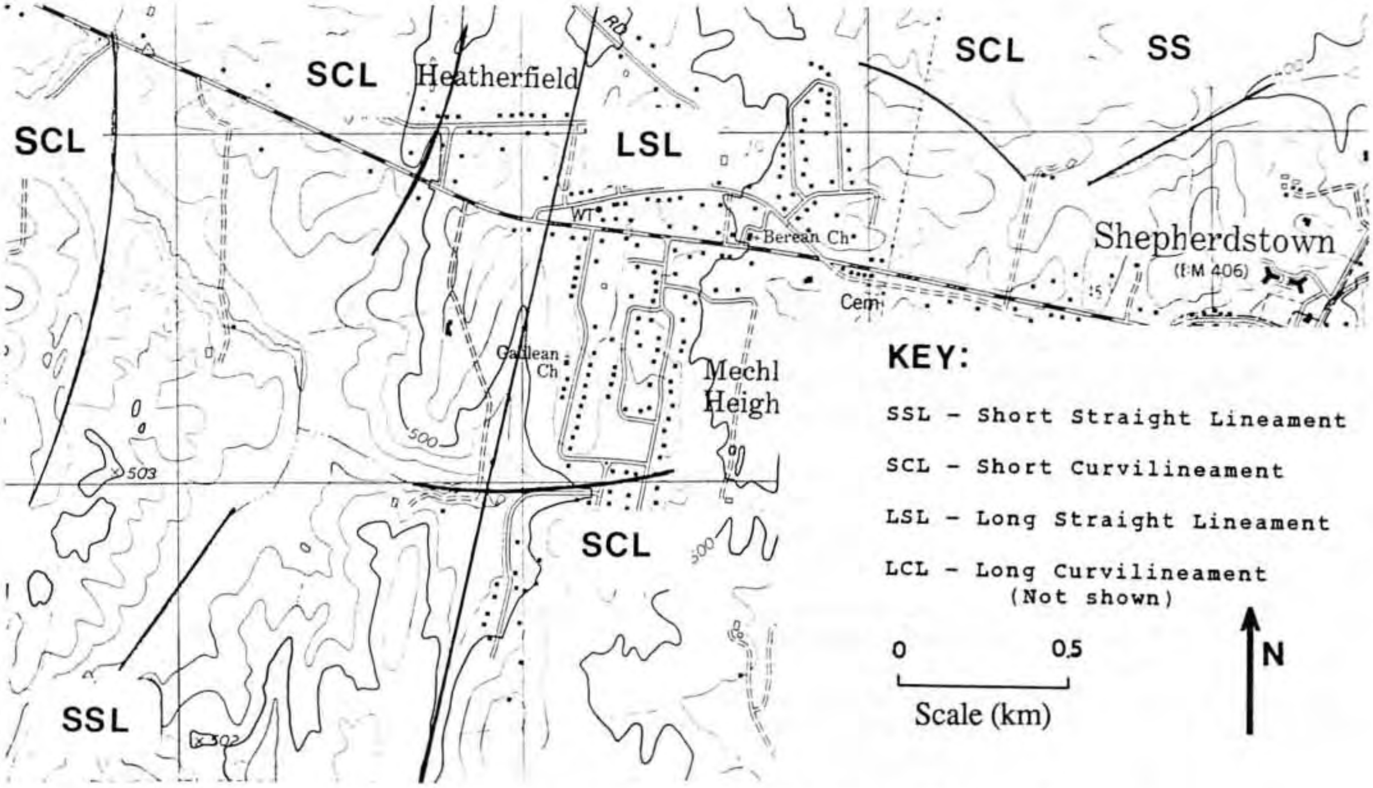

Figure 2: Part of the Shepherdstown 7.5-minute Quadrangle, showing typical types of mapped lineaments.

0.00005 and 0.0005 alpha probability levels, respectively, based on Mann-Whitney U statistical tests. Slope/upland wells have significantly higher yields than hilltop wells at the 0.05 alpha probability level.

It has been thoroughly documented from past studies that valley positions are favorable locations for higher-than-average yielding wells. There are several reasons for this. Valleys are areas of discharge where the water table is usually closer to the surface when compared to other topographic positions. Also, valleys preferentially form along areas of structural weakness such as fracture zones and thus promote differential erosion. These fractures would then be responsible for increased porosity and permeability associated with valleys, especially where cavities formed by dissolution are prevalent.

Well-Depth Influences

Well-depths, available for 401 analyzed wells located in carbonate terranes, were plotted against well yield. A notable drop-off in yield is apparent for wells deeper than 200 ft. Shallow wells (<200 ft deep) have a median yield of 20 gpm, compared to just 8 gpm for wells with depths of >200 ft. This trend is strongly significant at the 0.00005 alpha probability level. A secondary zone of moderate well production was observed between the 200-foot- and 300-foot-depth levels, where wells have a median yield of 15 gpm.

Higher well productivity at shallow depth probably represents an intensely weathered zone near the surface where well yields are high compared to the deeper, unweathered zone. Rock fractures tend to be more numerous and open, and solution cavities are more common within 200 ft of the surface. Valleys were also found to have a greater proportion of wells less than 200 ft deep than do other topographic settings.

Stratigraphic Influences

Wells in carbonate units have an overall median yield of 15 gpm with wells in individual carbonate beds having median yields ranging from 5 to 35 gpm. Table 1 shows a summary of the yields obtained from the Cambrian-Ordovician units. Lithologic descriptions and unit thicknesses were obtained from Dean and others (1987).

Karst Character Influences

The carbonate terrane was subdivided into poorly cavernous (PC) rock units and moderately cavernous (MC) rock units, based on inference from known caves as reported by Gulden and Johnson (1984). The lengths of those caves with natural entrances and with over 50 ft of passages were utilized to derive two cave indices, the cave density (CD) index and the average cave length (ACL) index. Our analysis was restricted to these caves because they probably best represent the larger population of all caves in the study area, including caves without entrances. The

Geol. Time Scale	Carbonate Rock Type	Range in Thickness (feet)	Lithologic Description	Average Cave Length Index+ (feet)	Cave Density Index+ ($feet/mile^2$)	Number of Test Wells	Median Well Yield (gpm)
ORDOVICIAN	Omc	540-860	Hard thinly bedded limestone with some pure limestone beds.	314	430.3	42	6
	Obps	0-500	Gray dolomite with calcite veins and chert nodules common.	*	*	3	35
	Obrr	2400-2750	Limestone and dolomite with dolomite predominant in the upper part.	168	42.3	62	20
	Obs	800	Limestone with argillaceous laminations. Silty laminations are common in the lower member.	*	*	30	12
CAMBRIAN	Cc	2200	Limestone with interbedded dolomite. Sandy dolomite at base.	56	4.3	167	12
	Ce	2000	Argillaceous, dolomitic limestone with shaly interbeds in the lower one-third.	856	84.6	87	15
	Cwy	1000	Dolomite and sandy limestone with sandstone and shale in upper part.	842	240.2	11	12
	Ct	1000	Massive dolomite and sandy dolomite with some interbeds of pure limestone.	786	164.6	13	5

Table 1: Descriptions and well yields for carbonate rock units in the study area.

Omc - Middle Ordovician Carbonates
Obps - Pinesburg Station Dolomite
Obrr - Rockdale Run Formation
Obs - Stonehenge Limestone

Cc - Conococheague Formation
Ce - Elbrook Formation
Cwy - Waynesboro Formation
Ct - Tomstown Dolomite

\+ - Based on only natural caves >50 feet long.
* - No natural caves > 50 feet in length are within these units.

PC rocks were judged to have an ACL index of less than 250 ft and a CD index of less than 75 ft/mi^2. The MC rocks were determined to have an ACL greater than 250 ft and a CD index greater than 75 ft/mi^2. None of the carbonate rock units were judged to be highly cavernous, as for example the Greenbrier Limestone in West Virginia. Table 1 shows the CD and ACL index for each Cambrian-Ordovician carbonate unit. Figure 3 shows the spatial distribution of PC and MC units. The MC units are located adjacent to areas of clastic rock at higher elevation where an increase in concentrated recharge owing to allogenic runoff may have led to increasing solutional development of these rocks, even though some are predominantly dolomitic.

The overall median yield of PC carbonate rocks is 15 gpm and the median yield for MC rocks is 12 gpm. This slight difference in well yield is statistically significant at the 0.10 but not at the 0.05 alpha probability level, indicating a moderate trend whereby the least cavernous rocks have higher well yields on average. It is believed that the PC rocks represent diffuse-flow aquifers with more uniformly spaced fractures and small cavities that are more conducive to high yields on average.

Structural Influences

Wells positioned on synclines in carbonate terrane show a slightly higher median yield (15 gpm) than wells positioned on anticlines (10 gpm) (based on the 136 wells located within 0.50 km of a structural fold axis). However, statistical tests show no significant difference in well yield between synclines and anticlines, at the 0.10 alpha probability level.

Carbonate units as a whole have a significant correlation between well yield and thrust faults. These major faults are usually oriented parallel to stratigraphic strike. Carbonate wells within 0.25 km of a thrust fault were found to have significantly higher yields than more distant

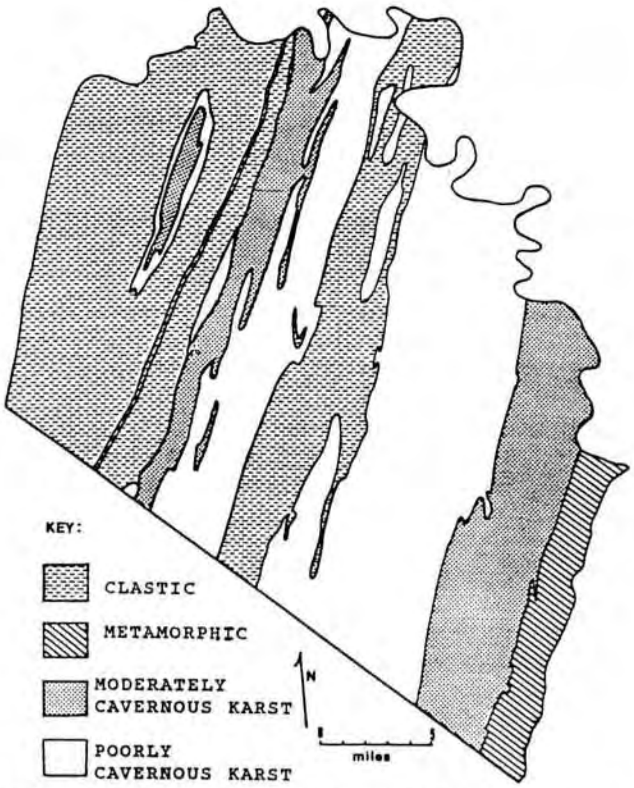

Figure 3: Spatial distribution of poorly cavernous and moderately cavernous carbonate-aquifer units within the study area.

wells at 0.05 alpha. This trend was re-analyzed, breaking the carbonate units into PC and MC units. MC units show no graphical well-yield trend with thrust-fault proximity, but PC units were found to have significantly higher yields (at 0.05 alpha) when located within 0.20 km of a thrust fault. For PC units, wells near such faults have a median well yield of 25 gpm compared to 10 gpm for wells >0.20 km from such faults. Thrust faults are probably accompanied by permeable fracture zones that enhance well yield.

Transverse faults and normal faults, usually oriented nearly perpendicular to stratigraphic strike, are also apparently associated with higher well yields. However, this trend is not statistically significant at 0.10 alpha.

Lineament Associations with Well Yield

LANDSAT Lineaments

A graphical plot of well yield versus distance to the nearest LANDSAT lineament showed that wells within 0.10 km of a LANDSAT lineament have significantly higher yields at 0.10 alpha than wells at greater distances (0.10 km to 0.50 km). When subdividing the well data into PC and MC units, PC units maintain and strengthen the well-yield correlation within 0.10 km that is statistically significant at 0.05 alpha, but no LANDSAT lineament correlation was observed for yields of wells in MC units.

Long Lineaments

A graphical plot for long, straight lineaments (LSL's) showed a well-yield trend with lower yields associated with wells located <0.20 km from a LSL. This trend is only slightly significant at 0.10 alpha. The statistical results indicate that long, straight lineaments do not strongly relate to well yield. No well-yield trend was found to be associated with proximal distance to long curvilineaments (LCL). Because only one out of 23 wells located within 0.20 km of a long lineament (LSL or LCL) is situated in a MC unit, these long-lineament results are indicative of PC units.

In an attempt to search for overlooked long-lineament trends, topography was held constant using the same data, but no significant well-yield correlations with long lineaments were apparent at 0.10 alpha. These data were again re-analyzed keeping well depth constant. Wells <200 ft in depth showed no well-yield relationships with long lineament proximal distance. Deep wells (>200 ft in depth) were found to have significantly lower yields when located within 0.20 km of a long lineament, at 0.05 alpha. Also, no significant well-yield associations were found with either long-lineament length or orientation for wells within 0.20 km of a long lineament. In general, therefore, long lineaments should be avoided in groundwater exploration.

Short Straight Lineaments

Short straight lineaments (SSL's) were next examined for well-yield associations. The median yield of wells within 0.05 km of a SSL (30 gpm) is over twice that of wells located from 0.05 km to 0.50 km from a SSL, and this trend is significant at 0.05 alpha. Also, wells located within 0.30 km of a SSL have significantly higher yields (15 gpm, median) at 0.01 alpha than wells located at greater distances (0.30 km to 0.50 km). These two results indicate that short straight lineaments are strongly associated with high well yields and that average well yield becomes progressively higher as such lineaments are approached within 0.30 km.

Topography was then held constant using the same data. Wells positioned as slope/upland and hilltop show significantly higher yields at 0.05 alpha when located within 0.20 km and 0.30 km of a SSL, respectively. However, valley wells show no significant correlation between yield and proximal distance to SSL's at 0.05 alpha.

Well depth was then held constant. Wells <200 ft deep have slightly significant higher yields at 0.10 alpha when located within 0.10 km of a SSL, and wells >200 ft deep show significant increases in well yield at 0.01 alpha if located within 0.30 km of a SSL lineament.

Other SSL parameter associations with well yield were then analyzed for wells within 0.30 km of a SSL. Statistical tests indicated no strong relationships between well yield and SSL length or SSL orientation.

Short Curvilineaments

Short curvilineaments were investigated for well-yield associations. A median well yield of 20 gpm was found for wells within 0.20 km of a short curvilineament (SCL) compared to a median yield of 13 gpm for wells located at greater distances (0.20 km to 0.50 km away). This trend is statistically significant at the 0.005 alpha level.

These data were then re-analyzed holding topography constant. Valley wells and slope/upland wells have an apparent well-yield correlation with SCL proximal distance, indicating increased yields within 0.20 km of a SCL lineament; these well-yield trends are statistically significant at 0.10 alpha and 0.05 alpha, respectively. Hilltop wells show no well-yield relationships with SCL proximal distance at 0.10 alpha.

Well depth was then held constant, subdividing the data into two groups. Deep wells (>200 ft deep) and shallow wells (<200 ft deep) were both found to have high yields if located within 0.20 km of a SCL, at 0.05 and 0.01 alpha, respectively. The deep-well trends found with SSL's and SCL's support the idea that short lineament fractures in the carbonate units penetrate deeper than 200 ft and influence well yields.

Lineament parameters were next investigated for wells within 0.20 km of a SCL. A SCL orientation of 300 to 330 degrees (across-strike) was found to be associated with significantly higher well yields at 0.05 alpha. SCL's with a radius of curvature of 0.5 km to 1.0 km were also found to be associated with increased well yields compared to nearly straight SCL's, but this trend is only significant at 0.10 alpha. SCL length was not found to be associated with well yield.

Short Lineaments

Because short, straight lineaments and short curvilineaments have many similar associations with well yield, their data sets were combined for comparisons with relative degree of cavern development.

The influence of lineament intersections could not be analyzed because too few exist; however, the only two carbonate wells located in close proximity (<0.10 km) of two short lineaments have yields of 150 gpm and 100 gpm. Both wells are located within a poorly cavernous (PC) unit, indicating that lineament density is an important factor determining high yield in such units.

Well-proximity data regarding short lineaments (SCL's and SSL's) were next grouped together and examined. A graphical plot of well yield versus proximal distance to the nearest short lineament for moderately cavernous (MC) units showed no apparent well-yield trend. A similar graphical plot was made for PC unit data, (Figure 4). This figure shows that wells in PC rocks exhibit strong short-lineament trends with higher yields in wells near such lineaments. A 33.5-gpm-median yield is evident within 0.05 km of a short lineament and a 17.5-gpm-median yield occurs within 0.20 km of a short lineament; these both indicate significant correlations at 0.05 and 0.001 alpha, respectively.

Data for well yield versus short lineament proximal distance were then analyzed for each of the carbonate units listed in Table 1. All of the PC units that could be statistically tested were found to have significant well-yield correlations with short lineaments at 0.01 to 0.10 alpha, whereas three of the four MC units showed no well-yield relationship with short lineament proximity at 0.10 alpha. Therefore, moderately cavernous units, that should exhibit somewhat less diffuse-flow character than poorly cavernous units, are not to be recommended for using short-lineament locations in groundwater exploration. When holding topography constant, valley and slope/upland wells have significantly higher yields within 0.20 km of a short linea-ment at 0.05 alpha. However, hilltop wells show no well-yield

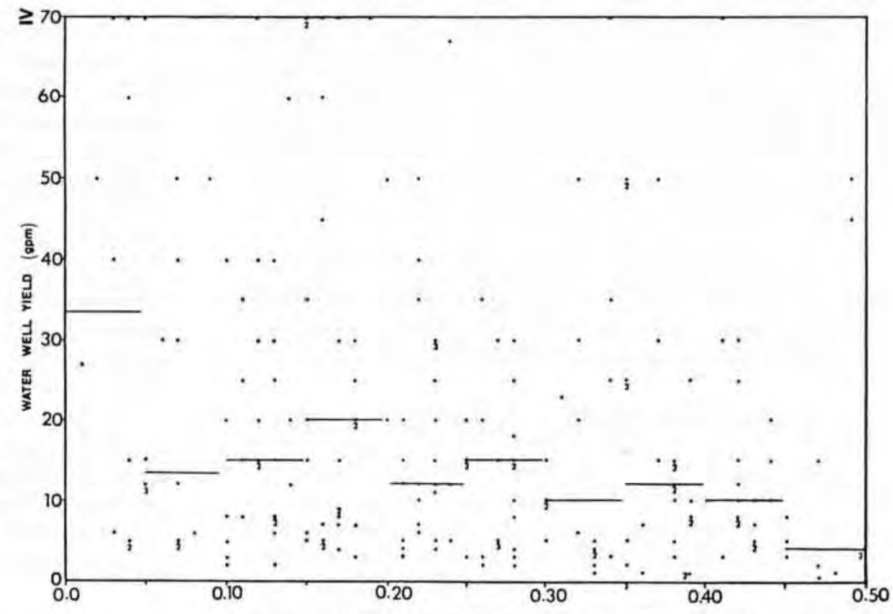

Figure 4: Plot of well yield (gpm) versus proximal distance (km) to the nearest short lineament for wells in poorly cavernous carbonate aquifers. Solid horizontal lines represent the median yield for each 0.05 km class of proximal distance to a short lineament.

correlation with short lineament proximal distance.

Lineament orientation was then analyzed with respect to well yield for wells within 0.20 km of a SSL or SCL. MC units show no well-yield association with short-lineament orientation. PC units show no well-yield relationship with SSL orientation, but have a statistically significant well-yield correlation with SCL orientation at the 0.05 alpha level. SCL's with an orientation of 300 to 360 degrees (across-strike) are associated with higher yields than SCL's at other orientations.

Lineament Expression

Dry-valley lineaments were compared with stream lineaments for their influence on well yield in MC and PC units. Only wells within 0.10 km of a short lineament were considered in the analysis. MC units have significantly higher well yields (20 gpm, median) near lineaments representing dry valleys than near lineaments representing streams, at 0.05 alpha. However, PC units have higher well yields associated with stream lineaments than with dry-valley lineaments, but not significantly higher at 0.05 alpha. Yields of wells from PC units and located near (<0.20 km of) stream lineaments are not significantly different from the yields of wells near streams without lineaments. PC unit wells located within 0.10 km of a stream have a median yield nearly three times the overall median yield for PC units. Thus, streams are important locations for high yielding wells in PC units, whether or not the stream is a mapped lineament.

A more detailed investigation of these wells within 0.10 km of short lineaments was also made. Wells in close proximity to perennial-stream lineaments appear to be higher producing than wells near intermittent-stream lineaments (not dry valleys), for both MC and PC rock units. The three wells near perennial-stream lineaments have yields of 30, 100, and 150 gpm. In contrast, two wells in PC units and located at the head of intermittent streams have yields of just 5 and 6 gpm, possibly indicating that these are poor producing areas.

Primary Conclusions for the Study Area

1. Topography is an extremely important factor impacting well yield. Valley wells have the highest median yield (25 gpm), followed by slope/upland wells (15 gpm) and hilltop wells (8 gpm).

2. The optimum depth in regard to well yield is less than 200 ft (20 gpm, median), with moderate yields for depths between 200 and 300 ft (15 gpm, median) and low yields for depths >300 ft (8 gpm, median). The optimum zone of enhanced fracturing and secondary permeability therefore is within 200 ft of the surface.

3. Poorly cavernous (PC) rocks have significantly higher yields than moderately cavernous (MC) rocks. Wells within PC rocks probably draw water from more numerous and better integrated fractures and bed partings than do wells in MC rock.

4. Wells in PC rocks have significantly increased yields within 0.20 km of a thrust fault (25 gpm, median), within 0.10 km of a LANDSAT lineament (20 gpm, median), and within 0.20 km and especially within 0.05 km of a short lineament. The very best PC rock wells are located within 0.10 km of two short lineaments, where eight times higher yield (125 gpm) may be expected. Apparently PC rocks are excellent diffuse-flow carbonate aquifers that have well integrated networks of fractures and small solution cavities along lineament/fracture zones.

5. No positive correlations between well yield and lineaments were found to be associated with MC rocks. Cavernous rocks may possibly be associated with narrow fracture zones, and widely spaced joints and bedding partings as noted by Rauch and Plitnik (1984). Also, such rocks may possibly be non-conducive to high well yield because of lowered water tables near conduits underlying the lineaments, where any fracture zones could be well drained, especially during times of low flow.

6. Wells located within 0.10 km of a stream in PC units have a median yield nearly three times the overall median yield for PC units, indicating that streams make good locations for wells in such rocks.

References

Bieber, P.P., 1961, Ground-water features of Berkeley and Jefferson counties, West Virginia: *West Virginia Geological and Economic Survey Bulletin 21*, 79 p.

Dean, S.L.; Kulander, B.R.; and Lessing, P., 1987, Geology of the Hedgesville, Keedysville, Martinsburg, Shepherdstown, and Williamsport quadrangles, Berkeley and Jefferson counties, West Virginia: *West Virginia Geological and Economic Survey, WV Map-WV31.*

Gulden, R. and Johnson, M., 1984, Caves of the eastern panhandle of West Virginia: *West Virginia Speleological Survey Bulletin 8*, 135 p.

Heller, S., 1980, *A Hydrogeologic Study of the Greenbrier Limestone Karst of Central Greenbrier County, West Virginia*: Ph.D. dissertation (unpublished), Department of Geology and Geography, West Virginia University, 167 p.

Hobba, W.A., Jr., 1976, Ground water hydrology of Berkeley County, West Virginia: *West Virginia Geological Survey Environmental Geologic Bulletin 13*, 21 p.

Hobba, W.A., Jr., 1981, Ground water hydrology of Jefferson County, West Virginia: *West Virginia Geological Survey Environmental Geologic Bulletin 16*, 21 p.

LaRiccia, M.P. and Rauch, H.W., 1977, Water well productivity related to photo-lineaments in carbonates of Frederick Valley, Maryland, *in* Dilamarter, R.R. and Csallany, S.C. (editors), *Hydrologic Problems in Karst Regions*, Western Kentucky University, Bowling Green, Kentucky, p. 228-234.

Lattman, L.H. and Parizek, R.R., 1964, Relations between fracture traces and the occurrence of ground water in carbonate rocks: *Journal of Hydrology*, v. 2, p. 73-91.

Ogden, A., 1976, *The Hydrogeology of the Central Monroe County Karst, West Virginia*: Ph.D. dissertation (unpublished), Department of Geology and Geography, West Virginia University, 263 p.

Rauch, H.W. and Plitnik, M., 1984, Use of lineaments in exploration for ground water in karst terrain of the Hagerstown Valley, Maryland, *in Geologic and Geotechnical Problems in Karstic Limestone of the Northeastern U.S.*: Association of Engineering Geologists and American Society of Civil Engineers, 18 p.

Siddiqui, S.H. and Parizek, R.R., 1971, Hydrogeologic factors influencing well-yields in folded and faulted carbonate rocks in central Pennsylvania: *Water Resources Research*, v. 7, p. 1295-1312.

Siegel, S., 1956, *Nonparametric Statistics for the Behavioral Sciences*: McGraw-Hill Book Co., New York, NY, p. 116-127.

On Calculating the Risk of Sinkhole Collapse

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ABSTRACT

It is often important to be able to calculate the statistical risk of sinkhole collapse for a specific site--for insurance purposes, for pre-purchase evaluation, and for engineering the development of the site. Most damaging sinkholes that occur today are cover-collapse or cover-subsidence sinkholes: the results of the downward movement of unconsolidated mantling sediment into dissolved voids in the karstic host-rock. The process of sinkhole development is continual and often repeated. Individual events of cover collapse or subsidence are stages in the formation of the larger landform termed a doline (or simply, a large, ancient sinkhole). It is possible to have a doline-dotted landscape where sinkholes are not collapsing today, or to have sinkholes collapsing (or subsiding) in an area where large surficial dolines are not present. Therefore, the calculation of the risk of sinkhole collapse must be based on statistics of recent events of collapse (or subsidence), not on the number of dolines that are detectable on topographic maps or aerial photographs. Obviously this requires the compilation of an exhaustive data base on recent collapse--from local residents, governmental agencies, newspaper reports, insurance reports, consulting engineers' records, etc.

Such statistics must be in the form of the number of collapses over an interval of time over some area of land: *e.g.* 23 collapses/30 years/25 square miles. In order to use a significant number of data points in the calculation, the risk factor will usually need to be calculated over a large area--16 or 25 square miles, for example. Because the number of sinkholes that develop will vary based on the lithology of the soluble rock, the statistical area must all be underlain by the same host-rocks. Moreover, the number of sinkholes that develop will also vary with the surficial geomorphology; therefore, the area of analysis must also lie within a single geomorphic setting. When the area of statistical record is selected appropriately, the average number of collapses per square mile per year can be calculated. A risk factor for the proposed site may then be calculated by multiplying by the area of the site and by the number of years of anticipated life of the property.

A 1,500-square-foot ranch-style home covers 0.00005 square miles, so the risk of sinkhole collapse affecting an individual home is usually very small. However, a 170 acre industrial park in the same area has a 5,000-times greater risk. If the final risk factor appears to have significant impact on the proposed use of the site, and if cost-effective engineering modifications are applicable to avoid the damage, then it is pertinent to conduct a geologic/geophysical/geotechnical investigation in order to delineate the specific areas that are most hazardous. The simplest and most effective method of mitigating the hazard is to redesign the project to avoid the hazardous area.

Introduction

Geological/geotechnical investigations to evaluate sinkhole hazards are often requested as part of the planning phase of a site evaluation. Such investigations may also be required as part of a remedial action, in which case the investigation is considerably more focused because damage has already occurred at a specific location. However, in the planning phase it is necessary to examine large tracts of land in a cost-effective manner.

To be cost effective, it is important that the evaluation take place in a hierarchical manner. *First, it must be determined if sinkholes are a significant hazard in this area*, for the specific intended use of the property. Second, cost-effective measures that could be taken to avoid the hazard must be identified. If the sinkhole hazard is significant, and if there are appropriate measures by which the hazard may be minimized or avoided, then, and only then, is it warranted to conduct a detailed, site-specific investigation. *Thus, the first step in evaluating the risk of sinkhole*

damage at a site planned for development, is a statistical analysis of the risk of sinkhole collapse.

Almost All Sinkholes that Collapse and Cause Damage Today Are Subsidence Sinkholes, Either Cover Collapse or Cover Subsidence

Geologically, a sinkhole is a closed depression in the land surface, a basin; generally of moderate dimensions, measured in ten's and even hundred's of feet in diameter. It functions as a drainage basin, funneling water into the subsurface. It owes its origin to the fact that the underlying bedrock dissolves in groundwater more rapidly than most rocks. Most sinkholes are formed on limestone or dolostone, although a few less common rocks are also soluble.

Although there are generally four or five different identified types of sinkholes (Beck, 1988; White, 1988; Culshaw and Waltham, 1987), these are the result of only two different processes: the transport of surficial material downward along solutionally-enlarged channels, or collapse of the rock roofs over large bedrock cavities (Figure 1).

Williams (1985) explains how deep solution pipes develop as master drainage channels connecting the shallow epikarstic zone to the deeper true karst groundwater. In a terrane where limestone is exposed at the ground surface, water flows over the limestone surface toward these dissolved "pipes" and then downward. Because water converges on these pipes, the limestone around them is more rapidly removed by solutional corrossion and erosion, resulting in a bowl-shaped depression. This is the classic *solution sinkhole* (Figure 1A). It develops imperceptibly and is not generally an engineering hazard except for the fact that additional voids may underlie the solution pipe.

Insoluble residue from the limestone may be left behind on the surface as a thin soil that tends to accumulate in the bottom of these sinkholes. If the residual soil becomes thick enough, the terrane is termed a subsoil karst. However, if the limestone is mantled by sediments of an outside origin (for example, marine sands or glacial drift), the terrane is termed a mantled karst. With respect to processes that form sinkholes, the two situations are the same. The overlying, unconsolidated sediment may simply be termed "cover" and the resulting sinkholes are cover-collapse or cover-subsidence sinkholes (*see* Figure 1 for examples). Conditions within the limestone are still the same as those described above, but the basin and the solution pipe that drains it are covered and infilled with loose sediment.

Precipitation will now infiltrate through the sediment and seep down to the limestone surface. There it will migrate along the bedrock surface to the solution pipe that drains the area, and through that, deeper into the limestone aquifer. The loose sediment directly above the solution pipe may gradually erode and be transported down the pipe, with the aid of infiltrating water, leaving a soil cavity over the pipe. If the sediment is somewhat cohesive, this void may grow larger and larger over time. More cohesive strata within the overburden sediment may cause the cavity to grow laterally, with a flat roof. Eventually the upward growth of the void may leave only a thin roof of soil that is not strong enough to support its own weight, and the ground surface collapses. At that moment, a large, gaping sinkhole suddenly appears: *a cover-collapse sinkhole* (Figure 1F). The limestone and the pipe may be visible in the bottom of the hole.

On the other hand, if the mantling sediment is relatively non-cohesive, the soil cavity may erode upward rapidly, without growing very large. In fact, as the roof of the cavity "crumbles" and is deposited on the floor of the cavity, the cavity simply migrates upward without growing larger, like a bubble rising through a liquid. When this reaches the surface, a small hole suddenly appears; this is also a cover-collapse sinkhole. In this case all that is visible in the hole is loose overburden sediment.

The engineering implications and hazards to man's structures resulting from cover-collapse sinkholes are obvious and may be catastrophic. It is important to understand that the position of the solution pipe and the processes operating in the subsurface have not changed and that this erosion process will continue in the future, with further potential collapses. However, the time frame is variable, sporadic, and unpredictable.

If the sediment overlying the limestone is very loose, the process of downward erosion and surface subsidence may take place gradually and continuously. As sand is transported down the pipe grain by grain, sand from above settles down to take its place. A roofed cavity is never formed, but the sediment is slowly eroded from beneath the ground surface causing a slow, gradual settling. The rate of subsidence may be only inches per year, but it is still a sinkhole: *a cover-subsidence sinkhole* (Figure 1E). Such sinkholes can cause cracking of rigid structures built over the subsiding area and have caused buildings, including homes, to be condemned and declared a total loss.

Such gradual sinkhole development may also occur where a clay stratum overlies limestone and is below the water table. As support is eroded from beneath the clay by erosion down the solution pipe, the clay becomes looser and may absorb more water. This loose, very plastic, water-saturated clay may gradually "sag" under the weight of the overlying sediment or structure. Because the basic cause of this subsidence is erosion of material downward into the dissolved voids in the limestone, this is a sinkhole. Because the cover sediment gradually subsides, this too is a cover-subsidence sinkhole.

The sudden failure of a cover-collapse sinkhole creates a bowl-shaped, funnel-shaped, or cylindrical depression in the ground surface. However, after the sudden development

of this basin, natural processes will begin to fill it. The sides will erode inward creating a broader, shallower depression. The depressed area may be wet and swampy, possibly even forming a small lake, where organic growth will infill the basin over time. These processes, and others, may infill the sinkhole completely, eventually leaving no sign of its previous existence. Such a feature is now termed a *buried sinkhole* (Figure 1D).

Buried sinkholes have serious engineering implications. Obviously, the original sinkhole is subject to reactivation if the sediment filling the solution pipe is eroded further down into the limestone. That is, it may collapse again.

The continued, repeated collapse of a cover-collapse sinkhole will sporadically, but inexorably, erode surface

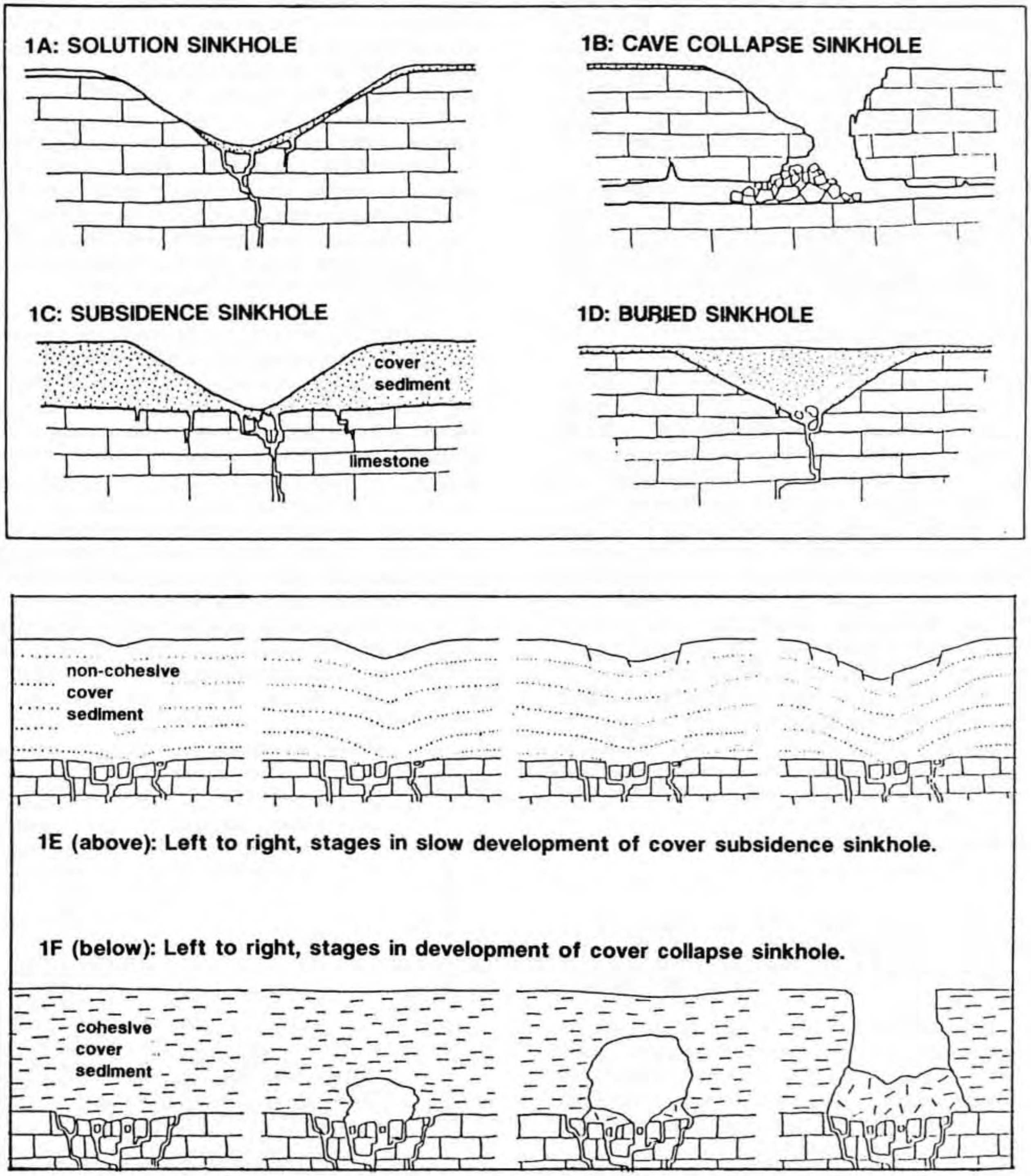

Figure 1: Types of sinkholes. E and F are subtypes of C. The processes of origin are discussed in the text. (From Culshaw and Waltham, 1987.)

sediments. The sinkhole cannot get significantly deeper, because the slowly dissolving limestone underlies the bottom, but it will gradually grow wider as material from the side slopes is eroded into the basin and then down the pipe into the subsurface. Thus, the very large sinkhole basins that we see today on a mantled karst landscape are probably not the result of a single collapse, but rather of a slow, sporadic erosion of the ground surface over thousands of years. As these sinkholes grow they may eventually merge due to cover erosion, forming even larger, irregular, compound basins. Of course, the long-term landforms are not pertinent to our discussion herein. However, *it is critical to understand that these large sinkhole basins develop through repeated collapse, on the same feature.*

Cover-collapse and cover-subsidence sinkholes are together termed "subsidence sinkholes," because the basic mode of origin is subsidence of the overburden, whether slow or rapid. Engineers have termed this upward erosion process "ravelling" and refer to these sinkholes as ravelling sinkholes. *Almost all of the damage to man's structures by sinkholes is from the subsidence variety.*

All four of the aforementioned types of sinkholes--solution, cover collapse, cover subsidence, and buried--are the result of downward movement of material through the limestone along solutionally enlarged channels. However, dissolution of the limestone may also produce caves at depth. These processes may result in a situation where a large cave is close to the limestone surface with only a thin rock roof over the void. If this rock roof suddenly collapses into the cave, a *bedrock-collapse or cave-collapse sinkhole* is formed (Figure 1B). However, White (1988, p. 359) states "because the rate of surface downwasting is extremely slow on a human time scale, [bedrock] collapse dolines are rarely observed in the act of collapsing." (Doline is the technical synonym for sinkhole.) Waltham (1989, p. 17) says that bedrock-collapse sinkholes are "rare, except on a geologic time scale." The Florida Sinkhole Research Institute (FSRI) has reports on more than 1,700 sinkholes that have developed in Florida in approximately the last 25 years, and not one of these was definitely a bedrock-collapse sinkhole. *All of them were cover-collapse or cover-subsidence sinkholes.*

To Calculate the Risk of Sinkhole Collapse It Is First Necessary to Have a Valid Data Base on Modern Collapses

A statistical risk assessment for sinkhole damage, either by collapse or by subsidence, must be conducted specifically for the site under evaluation. It is critical that the data base used for this evaluation be appropriate for the intended purpose. The data base should contain pertinent information on all, or as many as possible of, the sinkholes that have developed recently. In order to establish an appropriate time frame, it is necessary that the data base include the date of collapse, or at least the year of collapse (damaging cover-subsidence sinkholes should also be included in the data base). It is also critical that the data base *not* contain extraneous entries so old that the data collection cannot be complete for that time.

Figure 2 shows FSRI data on sinkhole collapses in the Orlando, Florida, area (Wilson and others, 1987). From 1981 to 1986, there was an average of 10.8 sinkhole collapses per year. However, from 1974 through 1986, the average is only 5.8 sinkholes per year. The data prior to 1981 is apparently much less complete, even though an effort was made to discover all old reports from newspaper files, insurance records, engineers' reports, government-agencies' reports, etc. Ketelle and others (1988) and Beck and others (1991) found similar results. Even when an exhaustive effort is made to discover past data on sinkhole collapses, the memory and records of the general public only appear to be reliable for three to five years. Obviously the best method of compiling a valid data base is a continuing data-collection effort over many years.

For risk-analysis purposes, it is only valid to use data for that period of time over which the data collection appears to be uniformly effective. However, when plotting the geographic or geologic distribution of sinkholes, the older data is still useful, although it must be recognized that it is probably biased toward the larger sinkholes because these are the most noteworthy and memorable.

It is important to note that *map or air-photo data on the locations of large, ancient sinkholes is not appropriate for this data base.* First, there is no year of collapse associated with these features, so they cannot be used in a temporal analysis. Further, as pointed out in the discussion above, these large basins are probably the result of repeated collapses over many hundreds and thousands of years.

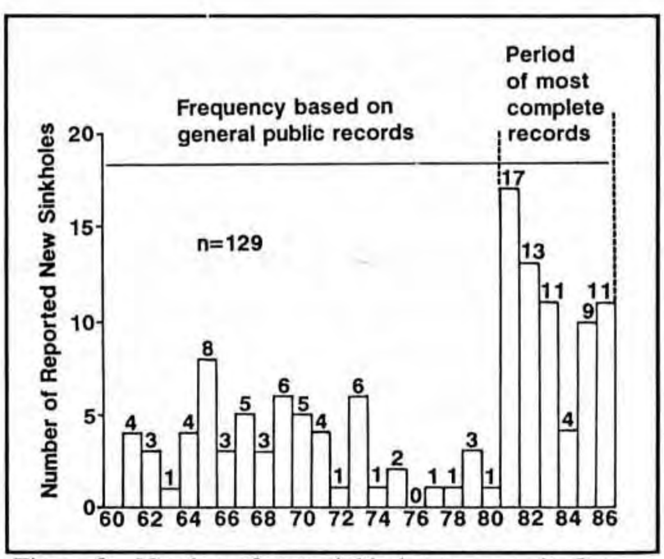

Figure 2: Number of new sinkholes per year in Orange and Seminole counties, Florida, for the years 1961-1986.

Moreover, several recent investigations have shown that these *large, ancient sinkholes cannot even be used as indicators of areas where sinkholes are now likely to form.* "In areas of covered karst..., the distributions of ancient sinkholes do not predict modern sinks" (Upchurch and Littlefield, 1987, p. 363). Whether this is the result of changing conditions--such as rainfall, sea level, or ground-water level--or simply because the collapses are so infrequent that they are not noted during the normal, brief (circa twenty-five years) period of record, is a moot point. *In mantled karst "....new sinkholes are appearing most commonly in the areas that do not correspond with areas of high old sinkhole density"* (Bahtijarevic, 1989, p. 80).

For the purpose of long-term planning, such as for the design life of a building, *it is important that the sinkhole data base encompass a sufficient time span to include climatic extremes of rainfall and drought.* These extremes may act as a "threshold" to totally modify the sinkhole incidence in an area.

For example, in the vicinity of Chiefland on the central western coast of Florida, the data base compiled by the FSRI contained records on less than ten sinkholes from the late 1960's through 1990. This is probably somewhat low due to incomplete reporting. In late February, 1991 the Chiefland area received more than 10 inches of rain in one weekend. At that time, more than a hundred sinkholes collapsed in the immediate Chiefland area. Obviously, the incidence of sinkholes calculated for the Chiefland area as of December, 1990, would have been far different than it would be if calculated now.

In western-coastal Florida, annual rainfall was unusually high during the late 1950's--significantly greater than it has been during the period for which the FSRI has been collecting data on sinkhole collapses. Furthermore, just prior to this time, in the mid-1950's, a three-year drought occurred, ending with the driest year since 1900. The records on sinkhole collapse compiled by the FSRI are not valid for either of these climatic extremes, yet the possibility of such conditions occurring again is obviously significant. Thus, even though the data that have been compiled in the Institute's data base are the most complete available and as thorough as professionally possible, they may not be totally indicative of long-term sinkhole risk in the area. For this reason, and other reasons discussed previously, *statistics on sinkhole collapses must generally be accepted only as a minimum.*

Risk of Sinkhole Occurrence Is a Function of the Number of Sinkholes that Develop Per Unit Area Per Unit Time

A valid data base on recent sinkhole occurrence, such as discussed above, may be used to calculate the minimum occurrence rate of sinkholes. However, the risk factor must be calculated in a valid fashion in order for this to be applicable to a specific site being evaluated.

Ketelle and others (1988), studying sinkhole collapses in eastern Tennessee, found a maximum occurrence rate of 1.69 sinkholes/mi^2/year in Hamblin County. However, the more commonly encountered "high" incidence rate ranged between 0.1 and 0.7 sinkholes/mi^2/year. In the Orlando, Florida, area, Wilson and others (1987) found the highest sinkhole rate to be approximately 1 sinkhole/mi^2/ 25 years or 0.04 sinkholes/mi^2/year. However, as noted previously, the older part of the data base was inapplicable and the Orlando rate may be approximately 0.1 sinkholes/ mi^2/year, based on the more up-to-date part of the data.

In order to have a statistically valid estimate of sinkhole-occurrence rate, a significant number of sinkholes must be counted. Assuming at least ten, and preferably more, sinkholes and an occurrence rate of 0.1 sinkholes/ mi^2/year, it is suggested that *the product of the area over which the sinkholes are counted (in square miles) times the number of years of record, should be at least 100,* as a general rule of thumb. Obviously, most sites for development are only fractions of a square mile. Therefore, in order to collect statistically valid data, it is necessary to select a larger study area of several square miles surrounding the site, that will be presumed to be representative of the site.

However, a larger sampling area cannot be selected at random, for example by drawing a radial distance around the site. *The statistical sampling area must be geologically consistent, i.e.* the bedrock geology underlying the site must be similar for the entire sampling area. Ketelle and others (1988) show that there are distinctly different rates of sinkhole occurrence in karst developed on different limestone formations. In eastern Tennessee, the Conasauga Group has a high rate of 0.29, the Knox Group 0.13, and the Chickamauga Group 0.04, sinkholes/mi^2/year. It seems logical that geological consistency must be stringent to be valid. For example, a given formation may contain a carbonate member and a shaley member. Obviously, the incidence of sinkhole occurrence will be different for each.

Furthermore, *the statistical sampling area must be geomorphically and hydrologically consistent.* Wilson and others (1987) found that the occurrence of sinkholes in the Orlando, Florida area correlated with geomorphic ridges, that were also high recharge sites. The entire region is underlain by similar bedrock geology. Therefore, in the Orlando region, a statistical sampling area should be either all located on the ridges, or entirely on the inter-ridge lowlands. A sampling area which spans both provinces would average "apples and oranges" and would arrive at a meaningless result. Other geomorphic and hydrogeologic constraints may modify the occurrence rate of sinkholes and these must be evaluated carefully before selecting a sampling area applicable to the site under evaluation.

Selection of an Acceptable Level of Risk Must Be Tailored to the Intended Use of the Property and to the Cost of Avoiding Potential Damage

Note that the impact of a sinkhole on a large shopping center or an industrial park may be minimal. As a first consideration, the odds are that it will only impact the parking area, where the chances for a serious accident are minimal. Secondly, assuming no unusual human tragedy, the cost incurred in repairs and lost revenue will probably be minor compared to the overall cost of the project. However, if the site is to be used for an airport, or a nursing home, the impact could be far different. A sinkhole in an airport runway could cause a major tragedy, and the extensive area of major airport runways raises the odds of occurrence. On a nursing home property, a sinkhole would probably not cause extensive damage *per se*, but the loss of confidence from the residents and their families could cause major loss of revenue.

If the statistics on sinkhole collapse indicate that the site being considered has an unacceptable risk, it is important to evaluate whether preventative measures are possible and what the cost will be, before investing in a site specific survey to pinpoint areas of potential sinkhole activity. *One must balance the cost of site investigation plus the cost of engineering modifications to avoid sinkhole damage, against the potential cost and impact of the sinkhole.*

If the odds are 1:100 that a sinkhole will collapse on site and if the potential damage is $50,000, then it is only justified to spend $500 to avoid the damage. The probability of sinkhole damage to a 2,000-square-foot suburban home in northern Pinellas County, Florida, is 1:5,555 (Beck and others, 1991). Therefore, it is only warranted to invest $18.00 to avoid complete destruction of the $100,000 home, assuming a 50-year planned life. In such cases, insurance is the only practical solution; engineering modifications are simply not warranted. However, as the size of the site being developed increases and the potential cost of damage increases, it may become cost effective to attempt to avoid sinkhole damage *a priori*.

This is particularly true in the planning phase of the development of large sites, which may have significant odds of a sinkhole collapsing somewhere on the site. *The most cost-effective method of avoiding a specific sinkhole hazard is simply to resite the structure that is at risk.* This assumes a geologic investigation has identified a specifically hazardous area. Plans should place a building away from a hazardous area, that may instead be used for a parking lot or decorative landscaping. If it is not possible to redesign plans to avoid a hazardous site, there are obviously many different ways in which engineers could render a structure "sinkhole proof"--everything from deep pilings to a rigid foundation that would bridge over any sinkhole void. However, the cost of the modifications must be balanced against the risk of sinkhole occurrence and potential cost of damages.

References

Bahtijarevic, Aida, 1989, Sinkhole density of the Forest City Quadrangle, *in* Beck, B.F. (editor), *Engineering and Environmental Impacts of Sinkholes and Karst: Proceedings of the Third Multidisciplinary Conference on Sinkholes and the Engineering and Environmental Impacts of Karst, St Petersburg Beach, Florida, 2-4 October 1989*: A.A. Balkema, Rotterdam and Brookfield, p. 75-82.

Beck, B.F., 1988, Environmental and engineering effects of sinkholes--The processes behind the problems: *Environmental Geology and Water Science*, v. 12, p. 71-78.

Beck, B.F.; Estes, E.; and Sayed, S., 1991, The sinkhole hazard in Pinellas County: A geologic summary for planning purposes: *Florida Sinkhole Research Institute, University of Central Florida, Report 90-91-1*, 55 p.

Culshaw, M.G. and Waltham, A.C., 1987, Natural and artificial cavities as ground engineering hazards: *Quarterly Journal of Engineering Geology*, v. 20, p. 139-150.

Ketelle, R.H.; Newton, J.G.; and Tanner, J.M., 1988, Karst subsidence in East Tennessee, *in Proceedings 2nd Conference on Environmental Problems in Karst Terranes and Their Solutions*: National Water Well Association, Dublin, Ohio, p. 51-65.

Upchurch, S.B. and Littlefield, J.R. Jr., 1987, Evaluation of data for sinkhole-development risk models, *in* Beck, B.F. and Wilson, W.L. (editors), *Karst Hydrogeology: Engineering and Environmental Applications: Proceedings of the Second Multidisciplinary Conference on Sinkholes and the Environmental Impacts of Karst, Orlando, Florida, 9-11 February 1987*: A.A. Balkema, Rotterdam and Boston, p. 359-364.

Waltham, A.C., 1989, *Ground Subsidence*: Blackie, London, 202 p.

White, W.B., 1988, *Geomorphology and Hydrology of Karst Terrains*: Oxford University Press, Oxford, 464 p.

Williams, P.W., 1985, Subcutaneous hydrology and the development of doline and cockpit karst: *Zeitschrift für Geomorphologie, N.F.*, v. 29, p. 463-482.

Wilson, W.L.; McDonald, K.M.; Barfus, B.L.; and Beck, B.F., 1987, Hydrogeologic factors associated with recent sinkhole development in the Orlando area, Florida: *Florida Sinkhole Research Institute, University of Central Florida, Report 87-88-4*, 109 p.

A Suggested Strategy for Characterizing the Hydrogeologic Regime of Karst Terranes in the Valley and Ridge Province

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ABSTRACT

The Valley and Ridge Province is relatively close to the East Coast Megalopolis and many areas of it are coming under increased pressure from land developers. Growth will increase the likelihood of mishaps involving hazardous wastes and hazardous materials, and will also increase the need for municipal solid-waste landfills, sewage-treatment systems, and other solid-waste handling systems. The ability to characterize the hydrogeology of a site planned for development is important, and for some types of facilities, such as most generators or handlers of hazardous wastes, hydrogeologic studies are required under RCRA regulations. An understanding of groundwater flow paths and flow rates is critical in the early detection of released contaminants. This paper discusses a strategy for characterizing the hydrogeologic regime of karst terranes. The authors call for a comprehensive strategy that incorporates the best available combination of investigative tools.

Introduction

Karst terranes in the Valley and Ridge Province are more vulnerable to contamination than many other groundwater systems because weathering has produced solutionally enlarged conduits in carbonate rock along which movement of contaminants occurs very rapidly and efficiently. Many karst terranes in the region that are under the most pressure from developers - primarily in the Great Valley, developed on the Cambrian-Ordovician carbonate sequence - are characterized primarily by a diffuse-flow karstic aquifer. In such terranes, traditional strategies for groundwater monitoring, based on flow through granular media, are inappropriate. At the same time, however, a monitoring strategy based solely on tracer studies, as advocated by some geologists, may also be inappropriate because tracing materials usually become dispersed. A tracer such as dye may take weeks or months to be recovered, and then it may turn up at random, at all, some, or none of the monitoring stations. Thus, studies using only this method for establishing flow routes are often inconclusive at best.

Owing to the diversity of flow regimes in the region, it is our opinion that a comprehensive groundwater monitoring strategy should be used to determine the degree to which a given carbonate aquifer exhibits the characteristics of a karst aquifer. Such a strategy would involve some combination of remote-sensing techniques (both photogeologic and geophysical), groundwater tracing studies combined with cave mapping, and geochemical analyses of karst springs.

Discussion

During the initial phase of a characterization process, for any carbonate aquifer, photogeologic techniques and field surveying should be used to produce a detailed map of sources of recharge and points of discharge. At this phase it should be obvious if these features are readily visible (*e.g.* sinkholes/swallets, sinking streams, or springs) or if such features are muted or completely masked by alluvium or colluvium. Once the map has been prepared, it can be used for planning of geophysical investigations, geochemical analyses of spring waters, tracer studies or a combination of these. The most important point is that, in many

cases, no *single* investigative technique will be suitable for characterizing the full range of possible conditions (*i.e.* the spectrum from diffuse flow to conduit flow) that may be present within a given carbonate aquifer.

For example, in carbonate aquifers that are overlain by thick (10 m or greater) sequences of alluvial or colluvial materials, the water table is commonly above the upper surface of the bedrock. Such aquifers occur in many areas along the western slope of the Blue Ridge Mountains in Virginia, but particularly from Buena Vista north to Front Royal, where colluvium and alluvium derived from the mass wasting of the Blue Ridge massif cover large areas along the eastern edge of the Shenandoah Valley that are underlain by carbonate rocks (*cf.* Gathright and others, 1978a,b). Because it is commonly the hydrogeology of the overburden rather than of the carbonate bedrock that determines how fast groundwater travels through the subsurface in such aquifers, the flow will normally be diffuse. Thus, even though a carbonate aquifer is present at depth, the carbonate strata are almost always completely within the phreatic zone and the carbonate aquifer has little or no influence on the hydrogeologic behavior of the entire aquifer. In fact, because the overburden is a granular medium, the aquifer may be more readily characterized using traditional concepts of Darcian flow. It is very important to realize that tracer studies (*e.g.* Quinlan and Ewers, 1985) may be of only limited value. A monitoring strategy based solely on tracer studies is applicable only in areas with carbonate aquifers that have little overburden and that are characterized primarily by conduit flow in the vadose zone. In such aquifers tracer recovery usually occurs in a few days to, at most, three to four weeks. In areas where carbonate aquifers are overlain by relatively great thicknesses of overburden, however, tracer studies alone will commonly produce results that cannot be interpreted without much ambiguity owing to extended recovery times for tracers (several months to years) and high rates of dispersion and diffusion that occur in such aquifers.

In these combination overburden/carbonate-rock aquifers, the use of spring-water geochemistry to characterize the aquifer (Shuster and White, 1971) will probably also be of less value as an investigative tool. This is because the residence time of the groundwater is relatively great, allowing for greater geochemical interaction, not just between the groundwater and the carbonate rock, but between the groundwater and sediments in the overburden. As a result, even though large springs can and do occur locally, they are most likely discharging water that has traveled primarily through the overburden and has had little or no contact with the carbonate bedrock. Therefore, in the context of the investigative methods of Shuster and White (1971), it may not be possible to interpret the results obtained from analysis of spring waters without ambiguity.

Certain geophysical methods (*e.g.* ground-penetrating radar, seismic refraction or reflection) can be helpful in locating subsurface voids and in determining the thickness of the overburden in such areas without resorting to drilling, the most costly of all investigative techniques even though it provides a direct "look" at the subsurface. Results of geophysical surveys must be interpreted carefully, because the results will quite commonly be ambiguous (Greenfield, 1979).

As mentioned above, photogeologic techniques such as aerial-photo interpretation can be effective tools for identifying geomorphic features that may aid in investigation of the hydrogeologic regime of a carbonate aquifer. Fracture-trace analysis, a type of aerial photographic analysis, can be very helpful in identifying potential pathways of groundwater migration in carbonate aquifers (Mata and others, 1986). *Note* that with all remote-sensing techniques it is imperative that the findings be field checked.

Conclusions

With the use of a comprehensive strategy that takes advantage of all available investigative methods and tools, it is possible to gain a better understanding of the hydrogeologic character of a given carbonate aquifer, no matter whether that aquifer is a "classic" karst aquifer (*i.e.* one that is overlain by minimal overburden and is characterized principally by conduit flow in the vadose zone), or whether the aquifer is buried beneath relatively great thicknesses of overburden and is a diffuse-flow aquifer that is primarily in the phreatic zone. Making this determination is critical in the choice of an appropriate groundwater monitoring strategy at any regulated waste-handling facility located above a carbonate aquifer in the Valley and Ridge Province, but especially in those areas where development is occurring most rapidly.

References

Gathright, T.M., II; Henika, W.S.; and Sullivan, J.L., III, 1978a, Geology of the Grottoes quadrangle, Virginia: *Virginia Division of Mineral Resources Publication 10*, text and 1:24,000 scale map.

Gathright, T.M., II; Henika, W.S.; and Sullivan, J.L., III, 1978b, Geology of the Crimora quadrangle, Virginia: *Virginia Division of Mineral Resources Publication 13*, text and 1:24,000 scale map.

Greenfield, R.J., 1979, Review of geophysical approaches to the detection of karst: *Bulletin of the Association of Engineering Geologists*, v. 16, p. 393-408.

Mata, L.; Fauss, L.M.; and May, I.P., 1986, Use of fracture trace analysis in karst terranes for environmental remedial action planning - A case study, *in Proceedings of the Environmental Problems in Karst Terranes and Their Solutions Conference, Bowling Green, Kentucky*: National Water Well Association, Dublin, Ohio, p. 263-277.

Quinlan, J.F. and Ewers, R.O., 1985, Ground water flow in limestone terranes: Strategy, rationale, and procedure for reliable, efficient monitoring of ground water quality in karst areas, *in Proceedings of the 5th National Symposium and Exposition on Aquifer Restoration and Ground Water Monitoring, Columbus, Ohio*: National Water Well Association, Worthington, Ohio, p. 197-234.

Shuster, E.T. and White, W.B., 1971, Seasonal fluctuations in the chemistry of limestone springs: A possible means for characterizing carbonate aquifers: *Journal of Hydrology*, v. 14, p. 93-128.