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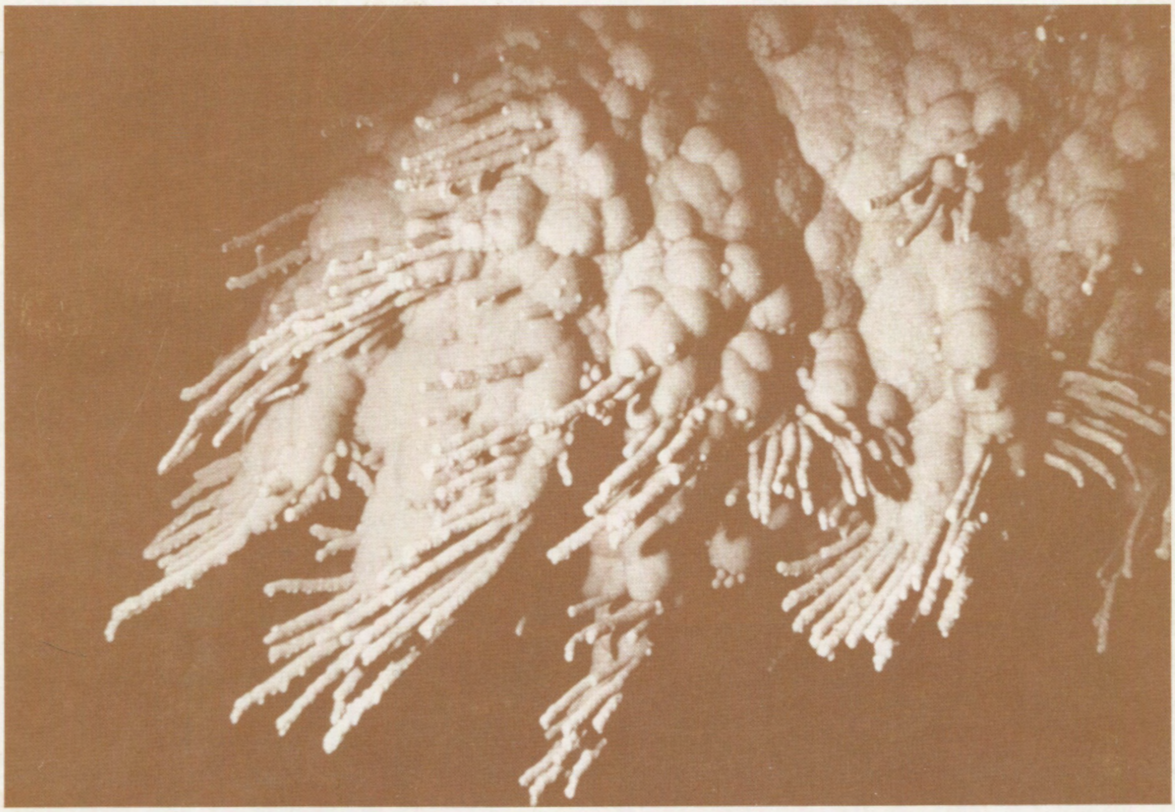

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LITHOLOGIC CONTROL OF SHALLOW KARST GROUNDWATER FLOW ON THE SINKHOLE PLAIN OF KENTUCKY

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The Lost River Groundwater Basin in Warren County, Kentucky is a karst drainage system encompassing 143 km² developed within the Mississippian St. Louis and Ste. Genevieve Limestones. Near the contact between these formations are two bedded chert units, the Lost River Chert within the Ste. Genevieve and The Corydon Chert Member of the St. Louis, which appear to be perching layers to shallow karst groundwater flow. Groundwater may be seen flowing at or near the tops of these beds in various cave streams and at swallets and springs within the basin.

In order to compare the vertical positions of these chert intervals to shallow karst groundwater flow, geologic structure maps of the Lost River Chert and the Corydon Chert Member were prepared for the basin. A contour map of the water table (at or near which shallow karst groundwater flow is assumed to take place) was constructed over the same area. These surfaces were digitized, then contoured and compared using SURFACE II and DISSPLA computer graphics systems. Correlation was accepted for points where the water table is either 6 m above or below the tops of the two chert layers. The water table (at baseflow conditions) was found to correlate with the Lost River Chert over 42.6% of the basin, as well as 40.7% for the Corydon Member. Shallow karst groundwater flow is found to correlate with bedded cherts over 83.3% of the study area, and therefore it is concluded that chert layers are concordant to shallow groundwater flow. This concordance, in conjunction with other observations of the relationship between groundwater flow and the cherts, suggests that the cherts strongly affect shallow karst groundwater flow within the Lost River Groundwater Basin.

INTRODUCTION

The Lost River Groundwater Drainage Basin is a karst drainage system encompassing 143 km² located on the Pennyroyal Plateau in southern Warren County, Kentucky (Fig. 1). As in other parts of the plateau, most drainage occurs through solutionally enlarged subsurface conduits within the Mississippian St. Louis and Ste. Genevieve Limestones (Fig. 2). Surface drainage is rare, and usually is intermittent where it does occur. Because most groundwater recharge to the system occurs quickly through swallet input at discrete points, rather than through diffuse recharge, the system is well into the conduit flow portion of the diffuse flow-conduit flow spectrum for karst aquifers (Smith, Atkinson, and Drew, 1976). The aquifer may be characterized as an open, free flow system in the terminology of White (1969).

The headwaters of the Lost River are located in the

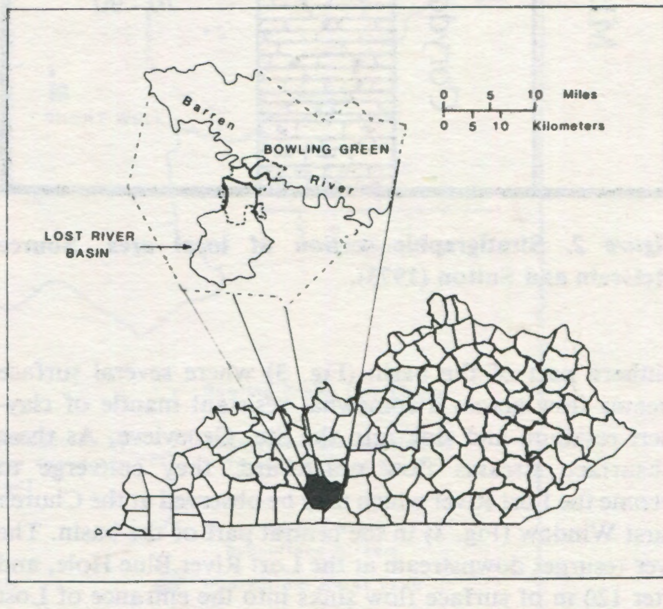

Figure 1. **Location of the Lost River Groundwater Basin.**

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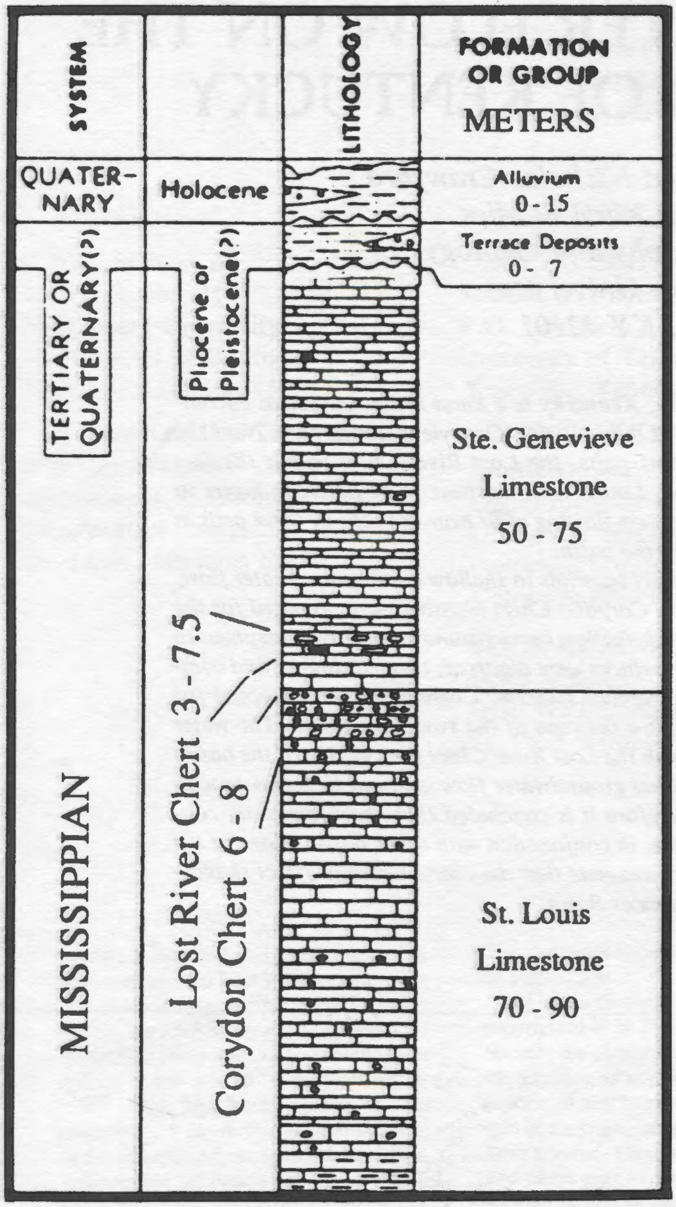

Figure 2. **Stratigraphic section of local area. Source: McGrain and Sutton (1973).**

southern part of the basin (Fig. 3) where several surface streams flow across a somewhat resistant mantle of clay-chert residium and sink into the Ste. Genevieve. As these subsurface streams flow northward, they converge to become the Lost River which may be observed at the Church Karst Window (Fig. 3) in the central part of the basin. The river resurges downstream at the Lost River Blue Hole, and after 120 m of surface flow sinks into the entrance of Lost River Cave. The river then flows beneath the City of Bowling Green to eventually resurface at the Lost River Rise and flow on the surface to the Barren River, the major base level stream for the area.

Although groundwater flows under an area of 143 km^2 within the basin, most flow (and associated cavern development) is concentrated within the lowermost 25 m of the St. Genevieve Limestone. Within, and at the base of this zone are two bedded chert units which appear to influence groundwater flow and cavern development (Crawford, 1982). The Corydon Member of the St. Louis Limestone (Woodson, 1981a) provides a base for the zone and consists of a limestone matrix packed with irregularly shaped chert nodules (Fig. 2). The Lost River Chert (Elrod, 1899), the top of which is about 12 m above the top of the Corydon Member, is a 3 to 7.5 m thick fossiliferous bed that ranges in composition from chert to partially silicified limestone to limestone.

Within the Lost River Groundwater Basin the cherts can be seen in various cave stream passages and appear to have a strong influence, at least locally, on groundwater flow. Outcrops of the two cherts are often visible at springs and swallets, and are very difficult to find at other locations within the basin.

The importance of lithologic heterogeneity on stratigraphic control of groundwater flow and landform development in karst areas (including the roles of the Lost River Chert and Corydon Member where they occur) is not completely understood. Howard (1968) found evidence of a high degree of stratigraphic landform control on the Pennyroyal Plateau which he associated with bedded cherts at or near the St. Louis-Ste. Genevieve contact. Other investigators have assigned a very minor role to their importance (Palmer, 1981; Quinlan and Ewers, 1981; Wells, 1976). The major objective of this study was to examine these relationships within the Lost River Basin.

THE STUDY AREA

The Lost River Groundwater Basin extends from the town of Woodburn northward for about 19 km to Bowling Green where the outlet for the basin, the Lost River Rise, resurges into Jennings Creek. At its widest, the basin is about 14 km wide. The study area comprises parts of the Bowling Green North, Bowling Green South, Woodburn, Drake, and Rockfield U.S.G.S. 7.5 minute topographic maps of Kentucky.

The basin lies within the Pennyroyal Plateau, which (along with the adjacent Mitchell Plain in Indiana and Highland Rim in Tennessee) is a classic sinkhole plain with gently rolling topography dominated by an abundance of sinkholes and other karst features. This landscape has developed on the nearly horizontal Late Mississippian limestones of the St. Louis, Ste. Genevieve, and Girkin formations. These units have a gentle northwesterly regional dip of about 6

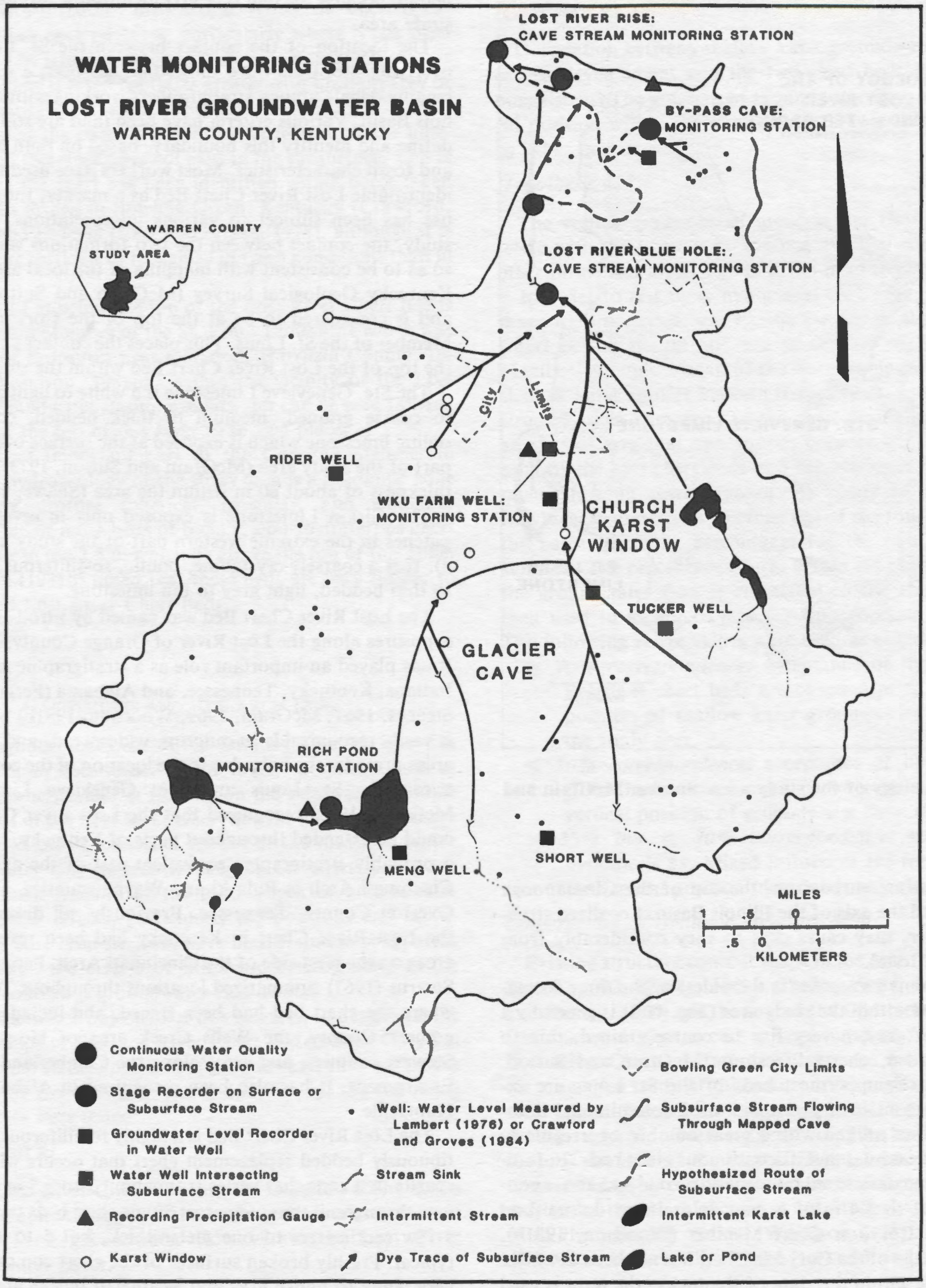

Figure 3. **The Lost River Groundwater Basin. Source: Crawford et al. (1987).**

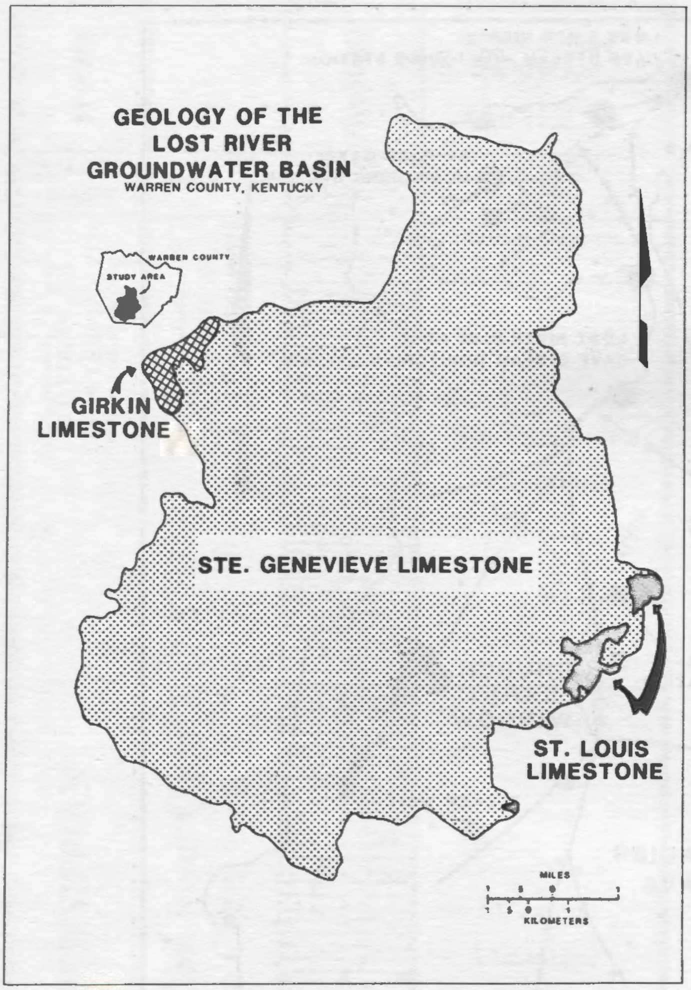

Figure 4. **Geology of the study area. Source: McGrain and Sutton (1973).**

m/km, when measured from the top of the Chattanooga Shale, toward the axis of the Illinois Basin. Localized structure, however, may cause dips to vary considerably from this regional trend.

The St. Louis Limestone is the oldest of the three formations exposed within the study area (Fig. 4). It is generally a light to dark grey, a very fine to coarse grained, thin to medium bedded, cherty limestone (McGrain and Sutton, 1973). Only the uppermost beds of the St. Louis are exposed; these consist of perhaps 6 m of dolomite and dolomitic limestone packed with a great number of irregularly shaped chert nodules and discontinuous chert beds. In Indiana it has been designated (along with a zone just above consisting of 1.2 to 2.4 m of brown dolomite or dolomitized micrite) the Corydon Chert Member (Woodson, 1981b). Although usage of the Corydon Chert has not been extended to these beds within the state of Kentucky, the term is used herein, because the unit is clearly identifiable throughout the study area.

The location of the contact between the St. Louis and overlying Ste. Genevieve Limestone has been a subject of lengthy debate among stratigraphers working within the Illinois Basin. Various criteria have been (and are still) used to define and identify this boundary, based on both lithologic and fossil characteristics. Most workers have used the easily identifiable Lost River Chert Bed as a marker, though their use has been subject to various interpretations. For this study, the contact between the two formations was chosen so as to be consistent with mapping of the local area by the Kentucky Geological Survey (McGrain and Sutton, 1973) and is considered to be at the top of the Corydon Chert Member of the St. Louis. This places the contact 12 m below the top of the Lost River Chert Bed within the study area.

The Ste. Genevieve Limestone is a white to light grey, fine to coarse grained, medium to thick bedded, commonly oolitic limestone which is exposed at the surface over a large part of the study area (McGrain and Sutton, 1973). It has a thickness of about 60 m within the area (Shawe, 1963a).

The Girkin Limestone is exposed only in several small patches in the extreme western part of the study area (Fig. 4). It is a coarsely crystalline, oolitic, fossiliferous, massive to thin bedded, light grey to tan limestone.

The Lost River Chert Bed was named by Elrod (1899) for exposures along the Lost River of Orange County, Indiana. It has played an important role as a stratigraphic marker in Indiana, Kentucky, Tennessee, and Alabama (Ferguson and Stearns, 1967; McGrain, 1969; Woodson, 1981b) because it is easily recognizable in outcrop, widespread, and in many areas provides the only clue to the location of the contact between the St. Louis and Ste. Genevieve Limestones. McGrain (1969) recognized that the Lost River Chert Bed could be extended throughout parts of Kentucky, including a probably stratigraphic equivalent east of the axis of the Cincinnati Arch in Pulaski and Wayne counties, as well as Overton County, Tennessee. Previously, all discussion of the Lost River Chert in Kentucky had been restricted to areas on the west side of the Cincinnati Arch. Ferguson and Stearns (1967) summarized locations throughout Tennessee where the chert bed had been traced, and included Montgomery County, the Wells Creek area of Houston and Stewart counties, and areas along the Cumberland Plateau Escarpment. It has also been recognized in Alabama near Huntsville.

The Lost River Chert Bed is a highly fossiliferous, discontinuously bedded replacement chert that occurs with white sparite in a zone that varies from about 3 to 7.5 m in thickness throughout the study area. Single chert beds range from a few centimeters to one meter thick, but 5 to 15 cm is typical. Freshly broken surfaces of the chert commonly are very white to bluish grey and weathered blocks are usually stained with a reddish brown tint. Where the chert is exposed within cave passages, particularly where it has been weathered by stream erosion, a dark brown or black color is common, and it often contrasts greatly with the light grey or white sparite host rock. Ferguson and Stearns (1967) claimed that the chert is a weathering feature of the limestones and therefore does not exist in the subsurface. This is not so, as the chert has been identified within cores taken from wells drilled in Bowling Green, Kentucky (Crawford, 1985b), and Indiana (Carr and others, 1978), although it is much more difficult to recognize in fresh conditions than in outcrop.

About 12 m below the top of the Lost River Chert Bed is the Corydon Member of the St. Louis Limestone which was named for exposures near Corydon, Harrison County, Indiana (Woodson, 1981b). Within the vicinity of the Lost River Groundwater Basin, groundwater can be seen flowing at or near the top of this layer in several places.

The most important part of the Corydon Member in terms of hydrogeology of the study area is a dolomitic zone containing profuse spherical, irregular, and lenticular cherts. One or more beds of blocky, non-fossiliferous chert also occur in the 2.5 m which lie above the top of this zone. Woodson (1981b) noted that these beds are very similar to weathered Lost River Chert Beds.

HYPOTHESES AND RESEARCH DESIGN

Throughout the Lost River Drainage Basin, one can observe many locations—within caves, at springs, and at swallets, where water is flowing on the relatively insoluble chert beds of the Lost River Chert and the Corydon Member of the St. Louis Limestone. However, the locations where groundwater may be seen flowing on the cherts constitute only a small fraction of the total length of groundwater paths. Other effects, particularly the positions and movements of base level streams, have been found in some areas to outweigh the effects of lithologic heterogeneity in the vertical control of groundwater flow routes.

It was a major objective of this research to determine the relationships of the Lost River Chert Bed and the Corydon Chert to the vertical position of groundwater flow within the Lost River Groundwater Basin. Specifically, the following hypotheses were tested:

Hypothesis #1

The Lost River Chert Bed and the Corydon Chert act as perching units throughout the basin and are concordant to shallow karst groundwater flow within the Lost River Groundwater Basin.

Hypothesis #2

Correlation between shallow karst groundwater flow and cavern development and the two chert beds is limited or nonexistent. The gradient of groundwater flow and its vertical position within the stratigraphic section are influenced by other factors.

Hypothesis #3

The vertical positions of groundwater flow routes correlate with the two cherts in some areas of the basin, but other influences affect this position in other areas.

In order to test these hypotheses, two structure contour maps were prepared: one having the top of the Lost River Chert Bed as datum and one of the top of the Corydon Member. A contour map of the water table surface for the shallow karst aquifer beneath the study area was also constructed, then compared to the chert structure maps. The amount of area that corresponds between the elevations of each of the two chert beds and the elevation of the water table (showing correlation of the cherts to groundwater flow) was expressed as a percentage of the total area within the basin, then the percentages for the two cherts were summed. The percentage of area within the basin over which the groundwater flow is correlated to the chert beds was then used to accept or reject the appropriate hypotheses. The following range values were used as criteria:

- $\geq$ *70% correspondence:* acceptance of hypothesis #1. Resistant chert beds are concordant to the vertical position of shallow karst groundwater flow within the study area.
- $\leq$ *30% correspondence:* acceptance of hypothesis #2. Chert beds have limited or no relationship to the vertical position of groundwater flow.
- $>$ *30% but* $<$ *70% correspondence:* acceptance of hypothesis #3. Mixed influences are responsible for the vertical position of groundwater flow.

Construction of Geologic Structure Maps

Existing structure contour maps in the area (McGrain and Sutton, 1973; Shawe, 1963a, 1963b, 1963c; Moore, 1963; and Rainey, 1964) use the top of the Chattanooga Shale, a Devonian unit that occurs several hundred feet below the Lost River Chert, as datum. These maps are not adequate for a detailed study of this nature. Although the Ste. Genevieve Limestone and the Chattanooga Shale are affected by similar regional influences (gently dipping to the northwest towards the axis of the Illinois Basin), considerable variations in thickness of the three formations between the units may mean that the structure is not precisely translated upward through the section. Woodson (1981b) found that differences between dips of the Devonian and Mississippian strata caused discrepancies as much as 18 m between inferred (from the Chattanooga Shale elevation) and actual elevations of the Lost River Chert in Southern Kentucky. Gentle structural elements within the Ste. Genevieve that may have significant influence on groundwater flow within the basin could certainly be absent on a structure map of the Chattanooga Shale.

Constructing such a map within the study area was difficult because the Pennyroyal Sinkhole Plain typically offers few rock outcrops at which to gather elevation data. The following sources of data were utilized:

1) Surface exposures.

Although not plentiful, outcrops of the cherts are located throughout the Lost River Basin. The first phase of this research was to identify new exposures through field checking of parts of the basin (as well as areas around the outside of the basin) that had not been thoroughly investigated, and to make an inventory of known chert outcrops from previous fieldwork. Outcrops of the cherts were found at springs, swallets and karst windows within the study area, as well as roadcuts, quarries and construction sites.

Outcrops of the Lost River Chert were more common than those of the Corydon Member within the study area. Fortunately, the top of the Corydon lies consistently 12 m below the top of the Lost River Chert (Woodson, 1981b) so the elevation of one of the chert beds can be used to infer the elevation of the other at that point. Although only a few locations were found within the study area to check this relationship, the assumption was supported.

Elevations of the tops of the chert beds at outcrops throughout the study area were found through transit leveling from the nearest point of known elevation to each outcrop, then back to the point of known elevation (to check accuracy) using standard procedures. Additional chert elevations at outcrops were determined using an altimeter.

2) Exposures within cave passages.

In a soil-covered sinkhole plain, caves often contain the best exposures of strata, although much of their true appearance may be masked by water and sediment. In this study, eight of the chert exposures were found within the caves (Fig. 5). Elevations of the cherts were found by transit leveling to a point near a cave's entrance and then continuing inward using a Suunto hand held clinometer and standard cave surveying techniques. Clinometer readings and backsites were estimated to one-half of one degree, and were required to agree within one degree.

Some of the clearest exposures of the Lost River Chert found within the study area occur underground, where the chert stands out conspicuously from passage walls. Where the chert is visible within caves throughout the study area it is most often seen on the floor or the lowest few meters of the passage walls where the stream has cut downward into this unit. In several local caves, streams flowing atop the chert breach it at joint related canyons, forming waterfalls.

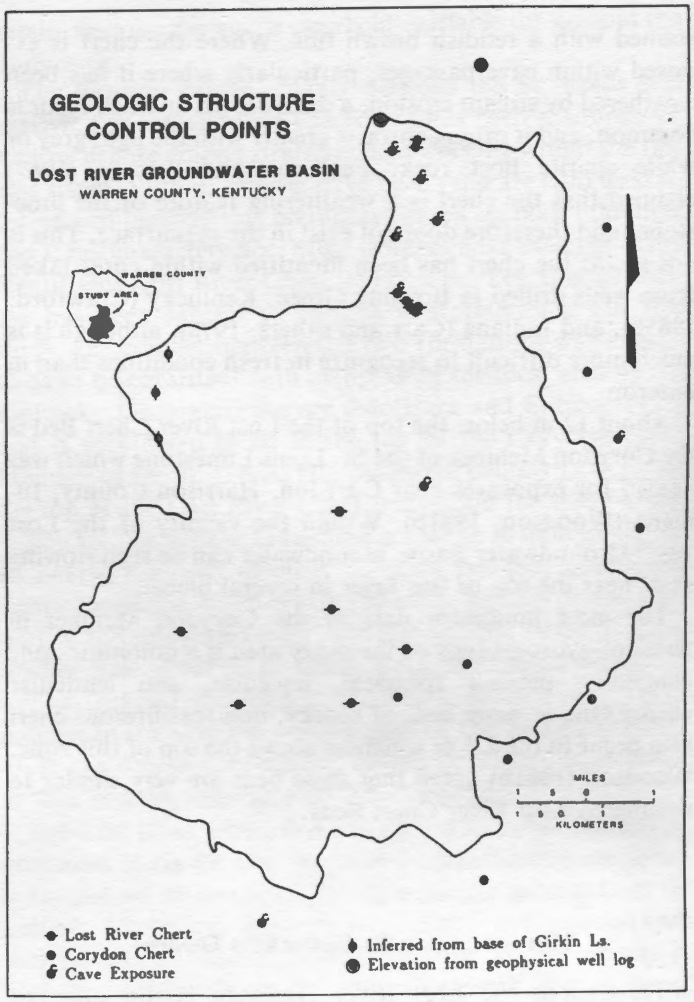

Figure 5. **Geologic structure control points.**

3) Existing well log information.

Warren County has a varied history of oil and gas production stretching back to the early part of this century. Dilamarter (1985) noted that before encountering oil bearing strata, drillholes are likely to penetrate shallow cavernous zones, including the St. Louis, Ste. Genevieve and Girkin Limestones. Records on wells that have been logged are available at the Kentucky Geological Survey in Lexington, and these were investigated for additional information on chert elevation. Well logs were located for 396 wells in or adjacent to the study area; however, most were logged only from the Chattanooga Shale (an important layer to oil drillers) downwards, and the ones that began at the surface were not in sufficient detail, because most wells were logged from cuttings. This development was particularly frustrating because most of the wells, particularly in the western sections of the basin, were started in the Ste. Genevieve or higher, and therefore penetrated both cherts.

4) Geophysical well logging.

Many wells that have been drilled in the vicinity of the study area are lacking stratigraphic logs. Over 450 storm water drainage wells have been drilled or dug, mostly within the urban parts of Bowling Green and the rapidly growing area to the south of town (Crawford and Groves, 1984). Many water supply wells also exist in the more rural parts of the basin. The Dresser-Atlas Company of Henderson, Kentucky, was hired to attempt to find chert elevations using bulk density logging equipment. The Lost River Chert has a bulk density of about 2.45 g/cm³ (Carr and others, 1978) and that of calcite (which makes up the great bulk of the Ste. Genevieve Limestone surrounding the chert) is 2.7 g/cm³ (Pough, 1976). The well logging program was only partially successful. Two test wells were run where the elevation of the Lost River Chert Bed was known, and clear negative density anomalies occurred at the predicted elevations in both cases. The chert was likely found in most of the other wells, but since small voids also produced negative density anomalies, these produced similar readings which often masked the chert. Of nine wells that were logged (including the two test wells), new chert elevations were accepted from only two wells. The wells accepted were in areas of better control, where there was a good idea of the approximate elevation of the chert. Interpretation was also aided by simultaneous caliper logging, which identified some of the voids that would produce non-chert negative density anomalies. Some wells in areas of poorer control produced possible chert readings, but interpretations were not considered reliable enough to use the data as control points for the structure maps.

5) Existing geologic maps.

Additional chert elevations were taken from the contact between the St. Louis and Ste. Genevieve on the geologic map of the Drake quadrangle by Moore (1963) who draws this contact at the top of the Lost River Chert. This procedure added points to the area east of the drainage basin. Data for the far western edge of the basin were taken from the Bowling Green North geologic map (Shawe, 1963a).

Construction of Geologic Structure Maps

Once the field data were collected, the surfaces were contoured using a combination of SURFACE II (Sampson, 1978) and DISSPLA (Integrated Software Systems Corporation, 1984) computer graphics systems. In order to input the control points into the system, they first required numerical encoding. This was done by plotting the points on a map of the study area, then measuring (from an arbitrary point chosen to the southwest of the study area) northward and eastward to each point. This gave a unique "x" and "y" coordinate for each control point, and the elevation then supplied the "z" value. Once a computer file was created for the points, SURFACE II took the unevenly distributed points and created a grid of evenly spaced "nodes" over the study area, estimating the elevation value at each node. For this project, a local fit procedure was used to estimate node elevations, utilizing the six nearest neighbors from each point. The surfaces were then described as x, y, and z values at 88 evenly spaced nodes, and the computer drew in contour lines at the appropriate locations. The Lost River Chert surface was contoured (Fig. 6), then SURFACE II contoured the Corydon Chert Member by lowering the Lost River Chert map by 12 m in elevation (Fig. 7). This provided a good model for the Corydon structure in an area where few outcrops are available. Thirty-three control points were used in the construction of these maps (Fig. 5, Table 1).

Table 1. Locations and elevations of control points used in construction of geologic structure maps.

Location	Elevation (m)	Longitude-Latitude	Chert Unit
Well C6–3	131.0	36°59'29"N 86°28'30"W	L.R.
Well F3–2	131.2	37°00'38"N 86°26'48"W	L.R.
Harvestman Haven Cave	128.8	*	L.R.
South Sunrise Well No. 1	129.9	36°59'08"N 86°28'06"W	L.R.
South Sunrise Well No. 2	131.7	36°59'12"N 86°28'06"W	L.R.
Sullivan Cave	136.6	*	L.R.
Robinson Cave	131.3	*	L.R.
Big Bertha Cave	134.0	*	L.R.
Lost River Cave	141.1	*	L.R.
L.R. Uvala Outcrop No. 1	144.0	36°57'08"N 86°28'15"W	L.R.
L.R. Uvala Outcrop No. 2	144.6	36°57'00"N 86°28'02"W	L.R.
L.R. Uvala Outcrop No. 3	145.3	36°56'57"N 86°27'53"W	L.R.
L.R. Uvala Outcrop No. 4	145.4	36°56'51"N 86°27'52"W	L.R.
Church Window Cave	154.6	*	Corydon
Sunken Spring	170.8	36°54'17"N 86°29'28"W	L.R.

(Continued on next page)

Table 1. Locations and elevations of control points used in construction of geologic structure maps—continued.

Location	Elevation (m)	Longitude-Latitude	Chert Unit
Big Sinking Creek	173.2	36°52'08"N 86°29'08"W	L.R.
Woodburn Cave	166.7	*	Corydon
Scottsville Road Retention Basin	160.1	36°56'35"N 86°25'16"W	Corydon
Chaney Lake	177.2	36°53'05"N 86°31'47"W	L.R.
Greenwood Cave	161.7	*	Corydon
Rich Pond	173.1	36°51'52"N 86°30'56"W	L.R.
Farm Creek	171.0	36°53'32"N 86°29'32"W	L.R.
Shawe No. 1	118	36°56'58"N 86°31'40"W	L.R.**
Shawe No. 2	121	36°56'22"N 86°32'05"W	L.R.**
Shawe No. 3	124	36°55'51"N 86°32'32"W	L.R.**
Matlock Road	166.1	36°52'09"N 86°27'23"W	Corydon
Plano Road	163.9	36°53'00"N 86°28'12"W	Corydon
Lively Road Sinkhole	175.2	36°52'10"N 86°28'28"W	L.R.
Moore No. 1	166	36°54'08"N 86°24'40"W	Corydon ***
Moore No. 2	174	36°52'00"N 86°25'18"W	Corydon ***
Moore No. 3	171	36°51'22"N 86°26'30"W	Corydon ***
Moore No. 4	176	36°49'50"N 86°27'05"W	Corydon ***

*Cave entrance location information available from the Central Kentucky Cave Survey, Dept. of Geography and Geology, Western Kentucky University.

**Lost River Chert elevation inferred from Ste. Genevieve-Girken contact of Shawe (1963a).

***Corydon Chert elevation inferred from St. Louis-Ste. Genevieve contact of Moore (1963).

Construction of the Potentiometric Surface Map

Several potentiometric surface maps have been constructed within the study area (Groves, 1983; Crawford, 1985a) and surrounding region (LaValle, 1967; Lambert, 1976; Quinlan and Ray, 1981; Plebuch et al., 1985). These maps are useful for understanding general directions of groundwater flow, although carbonate aquifers tend to be highly anisotropic. Groundwater flow routes at (or just below) the water table are contained within the surface shown on such a map.

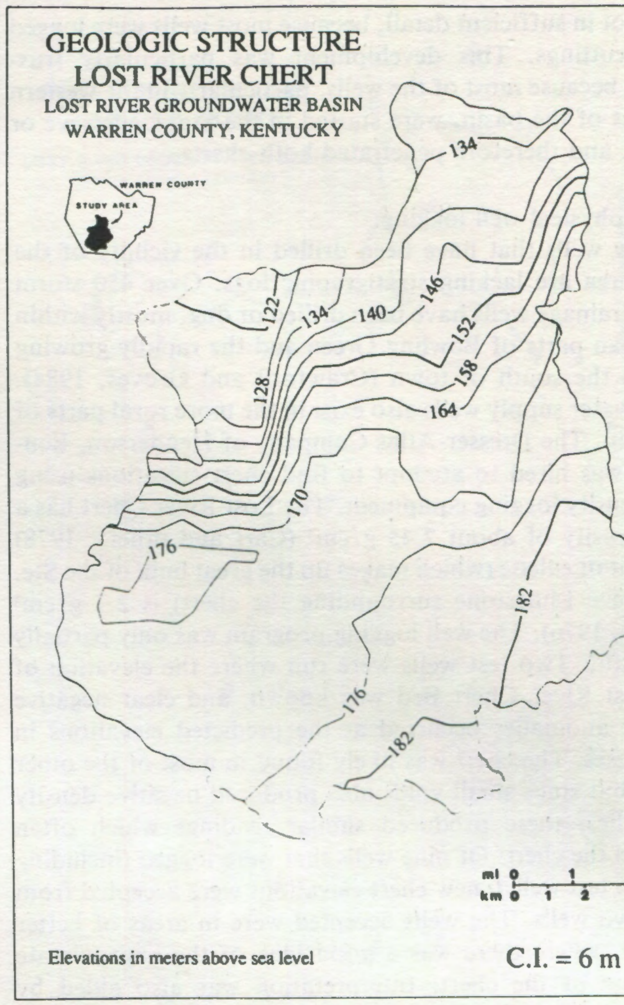

Figure 6. **Geologic structure, Lost River Chert.**

The map constructed for this study contains 283 control points (Fig. 8), and is the first one within the study area to be contoured by computer. Control points were gathered largely from existing data sources (Lambert, 1976; Crawford et al., 1984) and a few were added by the authors. An inventory of storm water wells done by Crawford et al. (1984) for the U.S. Environmental Protection Agency lists the locations (and water elevations) for 444 wells in the area. Unfortunately, these measurements were taken at a variety of antecedent moisture conditions and not all represent a probable base level condition, which was assumed for construction of this map. In order to utilize these data, a filtering system was established: the date of each measurement was checked in the records at the College Heights Weather Station in Bowling Green (on the Western Kentucky University campus in the northern part of the basin),

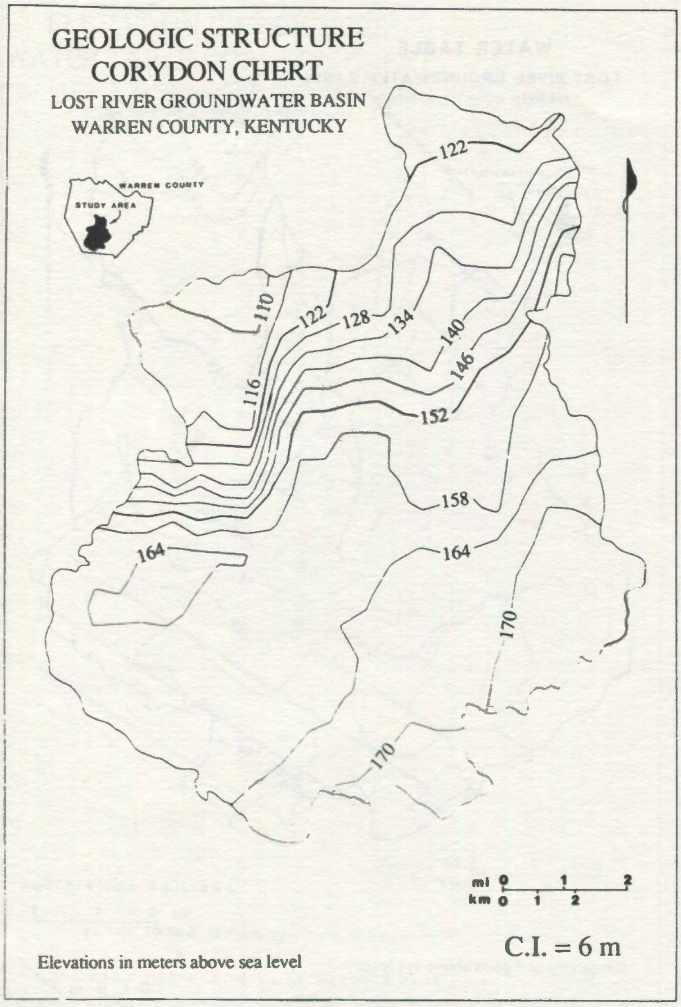

Figure 7. **Geologic structure, Corydon Chert.**

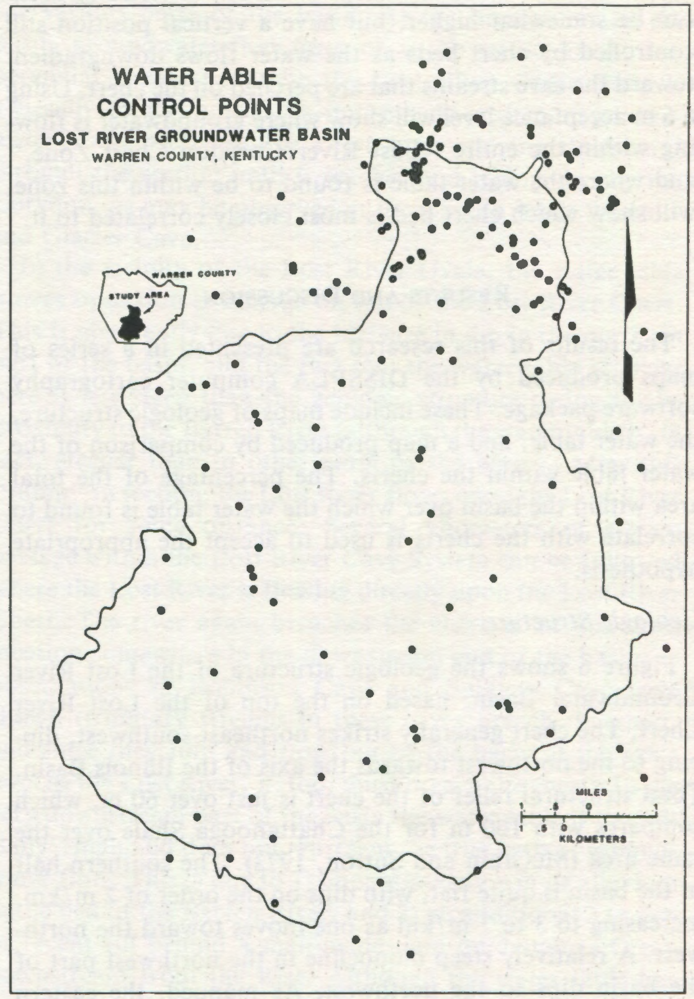

Figure 8. **Water table control points.**

and the measurement was not used if 1.2 cm of rain had fallen in the two days previous to measurement or if 5 cm had fallen in the previous week. In addition, a minimum well depth of 9 m was required along with a minimum water depth of 1 m. Wells that were found to be unacceptable were remeasured during dry conditions. Later water level measurements were concentrated during dry periods to prevent this problem.

The water table control points were digitized and the surface contoured by SURFACE II in a manner similar to the construction of the geologic structure maps.

Data Analysis

The procedures for analysis are described below for comparison of the Lost River Chert surface and the water table. This process was repeated for comparison of the water table and the Corydon Chert Member.

Once the maps were contoured by SURFACE II, the 3-dimensional surfaces of the Lost River Chert and the water table were numerically described as x, y, and z values at 88 evenly spaced grid nodes. Each node on the chert map has a corresponding "partner" node on the water table map that occupies the same x, y position when the maps are superimposed. The two surfaces then were tested for correspondence by comparing the z values, or elevations, at each pair of partner nodes. A pair of values was said to correspond if the elevation of the water table at a point either was within 6 m above or below the elevation at the top of the chert bed. This range was chosen because 1) some cave streams have cut down into the chert beds, 2) since a contoured surface is largely inferred, some tolerance must be allowed for the differences between the elevations on the map and reality, 3) collapses of cave passages can cause damming of the cave stream and thus a higher water table upstream from the breakdown dam, and 4) the water table in interstream areas can be somewhat higher, but have a vertical position still controlled by chert beds as the water flows downgradient toward the cave streams that are perched on the chert. Using a 6 m acceptance level will show where groundwater is flowing within the entire "Lost River/Corydon Chert Zone," and where the water table is found to be within this zone, will show which chert bed is most closely correlated to it.

RESULTS AND DISCUSSION

The results of this research are presented in a series of maps produced by the DISSPLA computer cartography software package. These include maps of geologic structure, the water table, and a map produced by comparison of the water table within the cherts. The percentage of the total area within the basin over which the water table is found to correlate with the cherts is used to accept the appropriate hypothesis.

Geologic Structure

Figure 6 shows the geologic structure of the Lost River Groundwater Basin, based on the top of the Lost River Chert. The chert generally strikes northeast-southwest, dipping to the northwest towards the axis of the Illinois Basin. Total structural relief of the chert is just over 60 m, which compares with 100 m for the Chattanooga Shale over the same area (McGrain and Sutton, 1973). The southern half of the basin is quite flat, with dips on the order of 2 m/km, increasing to 3 to 7 m/km as one moves toward the northwest. A relatively steep monocline in the northwest part of the basin dips to the northwest. As mapped, the eastern flank of this structure has dips that exceed 18 m/km. A low dome is mapped in the southwest.

The Water Table Surface

The contour map of the water table for the Lost River Groundwater Basin (Fig. 9) is similar to previous, hand-contoured water table maps of the basin. The water table reaches a high of about 190 m at the southern edge of the basin, and has a base level elevation of 128 m at the Lost River Rise. The general gradient of groundwater flow is 2 to 4 m/km, northward towards the Barren River, with the main trunk of the Lost River having a measured average gradient of 2.9 m/km along the mapped and inferred route.

Comparison and Correlation of Cherts and Water Table

Of the 88 grid nodes within the drainage basin at which the Lost River Chert and water table elevations can be compared, 42.6% show a correlation between the two surfaces as defined for this research. In addition, 40.7% of the nodes show correlation between the Corydon Chert Member and the water table. Summing these two quantities shows that the water table correlates with bedded cherts over 83.3% of the study area, and therefore hypotheses #1 is accepted: the Lost River Chert Bed and the Corydon Chert Member are concordant with shallow karst groundwater flow within the Lost River Groundwater Basin.

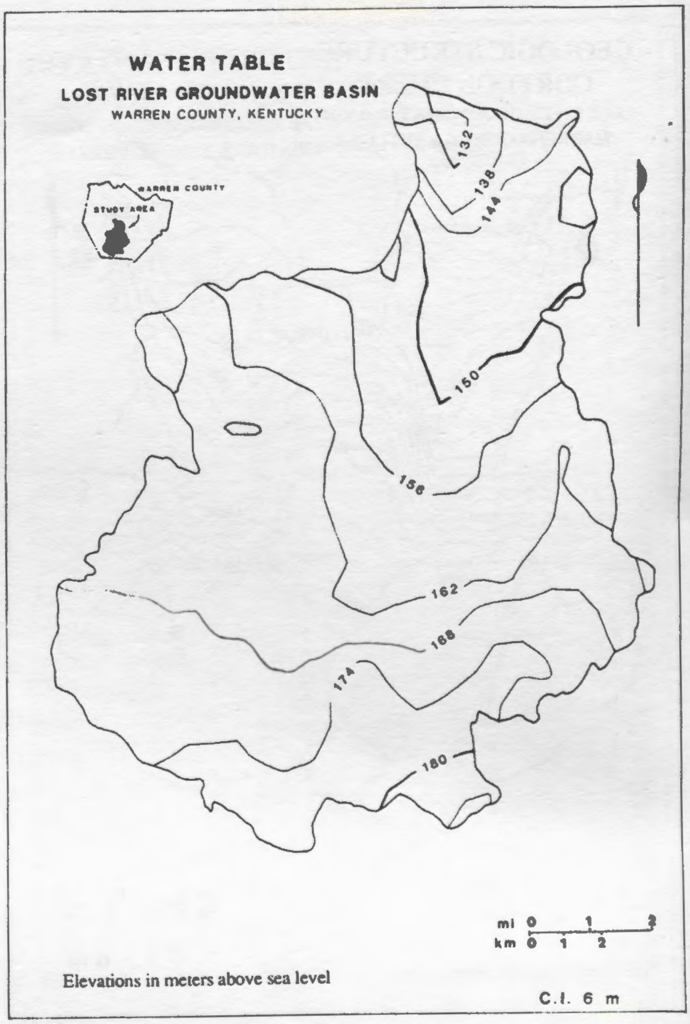

Figure 9. **Water table, Lost River Groundwater Basin.**

Figure 10 is a contour map of the elevation differences between the water table and the Corydon Chert throughout the study area. A value of zero indicates a perfect correlation between the water table and chert—positive values show areas where the water table is higher than the chert, and conversely, negative values occur where the top of the chert is higher than the water table. The areas that show correlation between the water table and the cherts are shaded.

Conclusion

With the results of this analysis, one can see the relationship between the chert beds and shallow groundwater flow

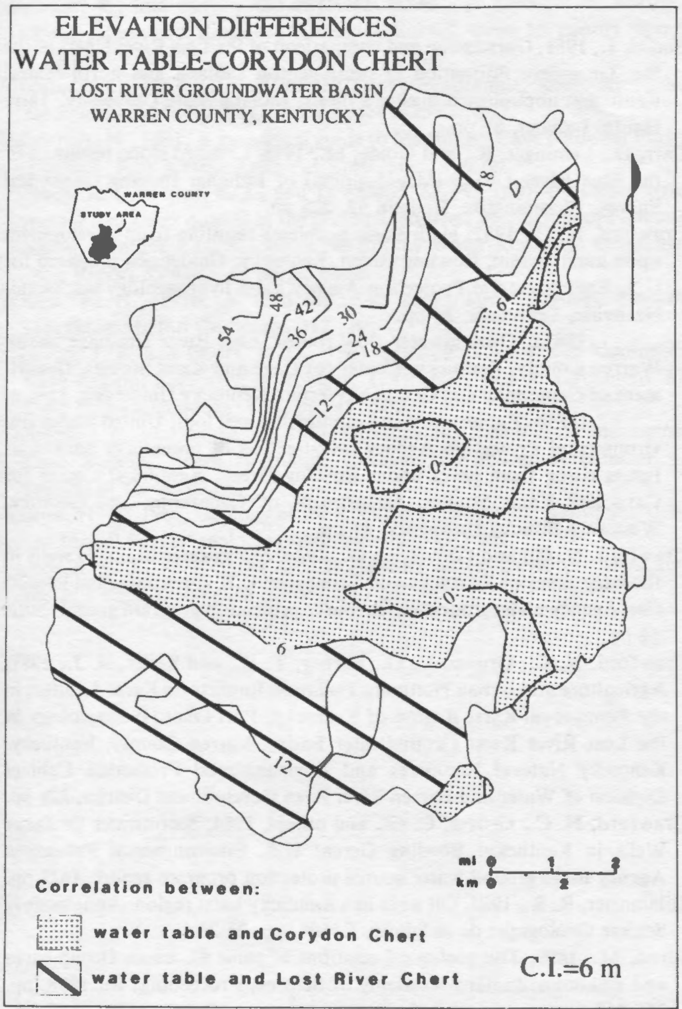

Figure 10. **Elevation differences, water table—Corydon Chert.**

for various parts of the basin. At the headwaters of the basin, groundwater flow (near or at the surface in the few perennial streams) is upon the Lost River Chert. Flow is generally to the north. As it flows downstream, groundwater soon breaches this chert unit, moving downward through the section until it reaches and is impeded by the Corydon, as may be observed at the Church Karst Window and Glacier Cave.

In the vicinity of the Lost River Uvala, the water table moves upsection to emerge on top of the Lost River Chert. This is apparently due to the increase in dip in the northern part of the basin, and the fact that the water table gradient here is less steep than the geologic structure owing to the proximity to the Barren River (although the water table is, in fact, steeper than in other parts of the basin). After this "jump" in section, the Lost River flows upon the Lost River Chert for some distance. Several kilometers of stream passage within the Lost River Cave System can be followed where the Lost River is flowing directly upon the Lost River Chert. The river again breaches the chert at an unknown location somewhere in the downstream end of the basin, as observations in Robinson Cave and Sullivan Cave have shown (tributary streams in these caves breach the chert, and since they join and flow out of the basin at the same level as the Lost River, it, too, must breach the chert). In the very downstream part of the basin the water table is "artificially" high and does not represent the true path of groundwater flow. This is due to 1) approximately 9 m of alluvium in the bed of the Barren River, and 2) an additional 2 to 3 m of recent fill behind a man-made dam on Jennings Creek downstream from the Rise. Although the water table has been raised in the downstream section, groundwater still flows through its original cave passage and flows upward from a depth of about 10 m at the Lost River Rise (Maegerlein and Dillon, 1980).

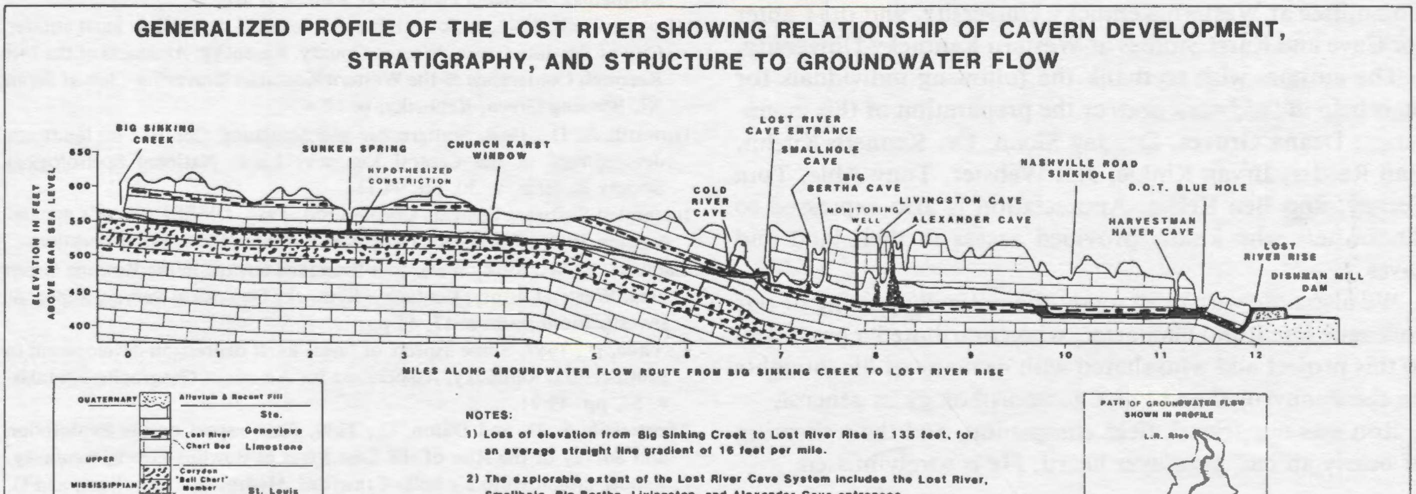

Figure 11. **Generalized profile of the Lost River.**

The correlation between shallow karst groundwater flow and the cherts over much of the basin, in conjunction with discreet observations of groundwater flow at or near the tops of the cherts at many locations, suggests that the beds are, in fact, influencing the vertical position of flow. In addition, the basin is somewhat cuestaform in cross section (Fig. 11) with a gentle dipslope toward the north. Precipitation falling in the very southern edge of the basin flows nearly 19 km to the Lost River Rise, rather than just over one kilometer south to Drake's Creek. This also suggests that the cherts are acting as perching layers for groundwater as the water flows downdip over this long route.

How can the results of this research for the Lost River Drainage Basin be extended outward to other areas of the Pennyroyal Plateau? The two chert beds are only present near or at the surface over part of this area. As the ground surface is relatively flat compared to the regional dip of the strata, in areas to the northwest the cherts are too deeply buried to have an effect on shallow groundwater flow. To the southeast (updip) the two cherts have been removed by erosion. Along the strike, however, the cherts are present over a large area, as they may be on the other side of the Cincinnati Arch, as McGrain (1969) noted for the Lost River Chert. It is suggested that in the areas of the Pennyroyal Plateau where shallow groundwater flow occurs within the same part of the geologic section as the Lost River Drainage Basin that the relationships between the cherts and groundwater flow may be similar. Other bedded chert units appear in various parts of the upper Mississippian System (Badiei, 1981) and may also influence some karst drainage systems.

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We also note with deep regret the untimely passing of our colleague Dr. Ron Dilamarter, who contributed a great deal to this project and who shared with us many of his thoughts on the Pennyroyal and karst geomorphology in general.

Ron was our friend, field companion, and the wellspring of nearly all bad puns ever heard. He is sorely missed.

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EXTRAORDINARY SUBAQUEOUS SPELEOTHEMS IN LECHUGUILLA CAVE, NEW MEXICO

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Many helictites in Lechuguilla Cave, New Mexico, show conclusive evidence for a subaqueous origin. Some are apparently still growing, or remain in the environment in which they grew. Helictites have previously been interpreted as exclusively subaerial in origin. Growth of the subaqueous helictites is triggered by the common-ion effect, where calcite-saturated water comes in contact with gypsum blocks. Dissolution of gypsum rapidly drives the water to supersaturation with calcite, forming helictites if the water enters a pool as discrete strands. Another previously undescribed speleothem type, tentatively named "pool fingers," is also subaqueous in origin. They are elongate growths of calcite that appear to coat organic filaments. Many are connected by curved bridges ("U-loops"). A third variety of speleothem, rarely described in the literature, consists of iron-oxide stalactites and columns lined with calcite. At least part of their origin was subaqueous. The iron oxide coats organic (bacterial?) filaments associated with oxidation reactions.

INTRODUCTION

Lechuguilla Cave, located in the Guadalupe Mountains of New Mexico in Carlsbad Caverns National Park, is an extensive cave in Permian limestones and dolomites. The area is part of the semi-arid Chihuahuan Desert. At the time of writing (1990) the cave has been mapped to a depth of about 477 m and a total length of more than 77 km. Its entrance is 1412 m above sea level. The cave consists of numerous large ramifying passages with complex spongework mazes superimposed in many areas (Fig. 1). Like other caves in the Guadalupe Mountains, it appears to have developed mainly in a phreatic and water-table environment by sulfuric acid derived from hydrogen sulfide from nearby petroleum reservoirs. The geologic setting of the cave has been described by Jagnow (1989).

Parts of the cave contain large deposits of gypsum, which was a byproduct of the sulfuric-acid reaction during cave enlargement. The cave also has numerous pool basins, both active and dry, which contain an exceptional variety of speleothems. The most remarkable speleothems resemble ordinary helictites and stalactites but are either actively growing underwater or are associated with old pool deposits that suggest subaqueous growth.

Nearly all published interpretations of helictites and stalactites (e.g., Hill and Forti, 1986) have assumed that they grow only in air. One exception (Peck, 1979) reported small tubular features similar to helictites and stalactites, consisting of sulfide minerals that apparently grew in phreatic cavities. Peck believed they were formed by injection of ascending fluids into geode-like openings. In recent years oceanographers have documented the origin of tubular mineral growths where sulfide-rich thermal water rises into sea water (Rona, 1986). Peck's speleothems are probably miniature analogs. In contrast, with one possible exception, the subaqueous speleothems of Lechuguilla Cave were formed by circulating fresh water from nearby sources of infiltration, rather than by hydrothermal processes.

The subaqueous helictites of Lechuguilla Cave, as well as several other types of speleothems of probable subaqueous origin, were first noted by Davis (1988a, 1988b). In this paper he describes these features and their associations. M. V. Palmer examines their mineralogy and crystallography, and A. N. Palmer analyzes their chemistry (see also Palmer, 1988).

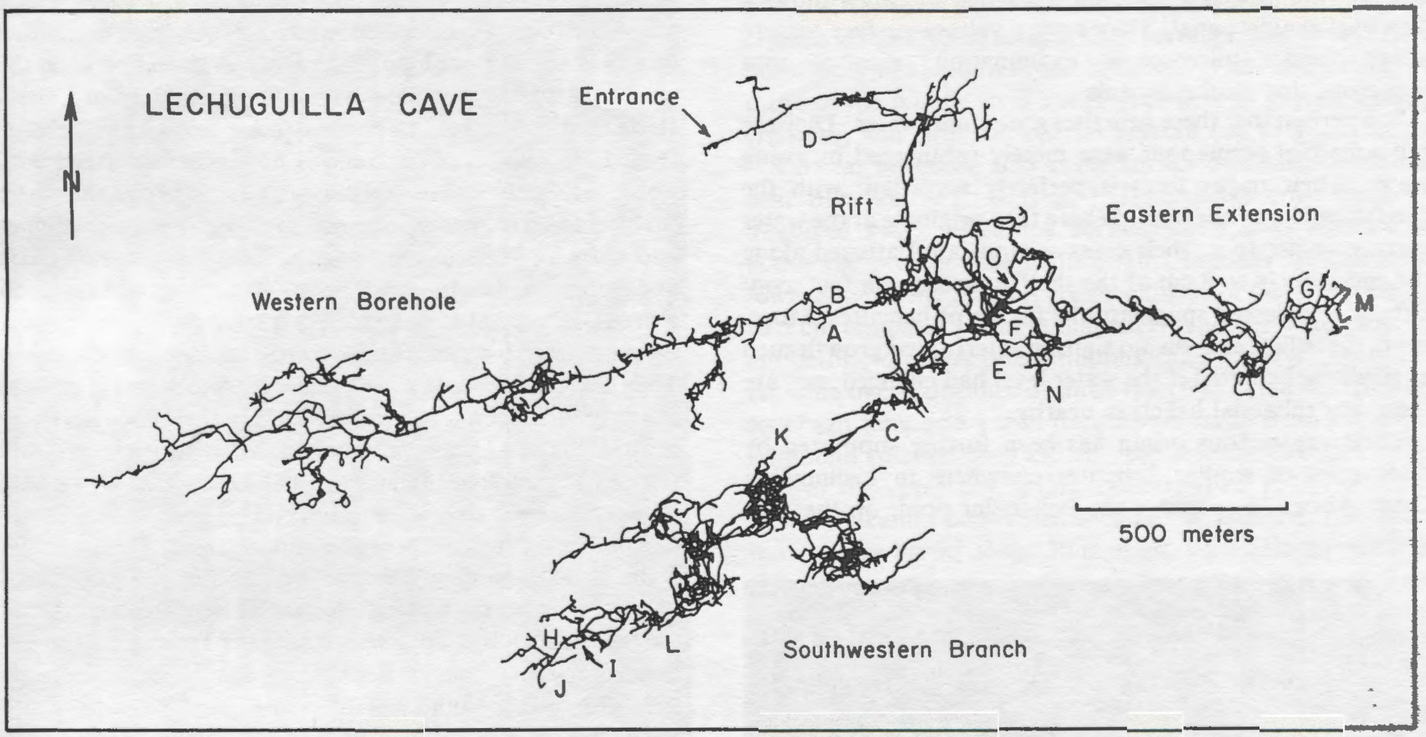

Figure 1. **Map of Lechuguilla Cave showing location of areas described in this paper. Map copyrighted by the Lechuguilla Cave Project, Inc.**

SUBAQUEOUS HELICTITES

Location and Description

The subaqueous helictites were first discovered in Pellucidar, a large room in the Southwestern Branch of Lechuguilla Cave, about 230 m below the entrance (Fig. 1, A). The room is decorated with active flowstone, dripstone, and cave pearls. Gypsum blocks up to several meters thick cover parts of the limestone floor. The helictites are located in a pool about 0.6 x 1.0 m in surface area and 2.4 m deep, within a gentle slope of wet white calcite flowstone. The pool is filled to overflowing by a thin film of water flowing over nearby stalactites and flowstone. It is lined with a subaqueous mammillary calcite crust and rimmed by thin shelfstone up to 12 cm wide.

Beneath the water surface is a spectacular display of twisting, worm-like helictites of white calcite up to about 35 cm long (Fig. 2). They are circular in cross section and between 1.5 and 5 mm in diameter, each having a rather uniform diameter. They emerge from the wall crust and extend from just below the water surface to depths of about 30 cm. They bend in both sinuous curves and erratic kinks but do not branch. On the average, those originating at greater depth are longer and inclined more sharply downward. One of the longest descends vertically for more than 10 cm parallel to the pool wall and only one centimeter away from it. The

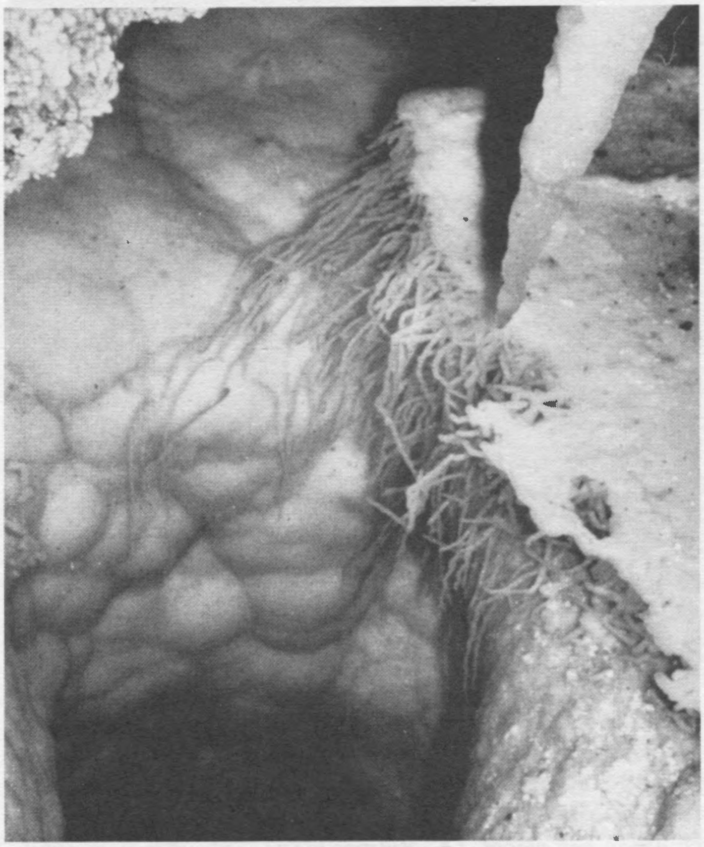

Figure 2. **Subaqueous helictites in Pellucidar (Fig. 1, A). Width of pool is about 60 cm. (Photo by Norman Thompson.)**

tips are rounded, and a naturally broken specimen shows a tiny oval central canal. They have a velvety surface texture which, under microscopic examination, resolves into numerous tiny calcite crystals.

It is certain that these helictites grew underwater. They are not subaerial forms that were merely submerged by rising water. Their upper limit is perfectly accordant with the shelfstone rim of the pool. Where they originate at the water surface or rise to it, their cross sections are flattened along the surface. Up to 8 cm of the shelfstone edge, in fact, consists of a coalesced spaghetti-like fusion of helictites. Moreover, the helictites show no signs of interrupted growth such as might be expected if the water level had changed, nor are there any subaerial helictites nearby.

Their subaqueous origin has been further supported by discoveries of similar helictites elsewhere in Lechuguilla Cave. About 30 m above the Pellucidar pool, on the Barsoom balcony, is a smaller pool in a crevice at the base of a wet flowstone slope, which contains shelfstone-accordant helictites up to 8 cm long (Fig. 1, B). A third site is on the upstream side of a shallow basin only 5 cm deep in a flowstone floor in the Southwestern Branch below Lake Lebarge (Fig. 1, C; Fig. 3). This basin is now only half filled with water, so the helictites are partly exposed above the water surface and are presumably not growing. They are thinner and more variable in cross section than those in Pellucidar and Barsoom. The largest is about 10 cm long and tapers to a needle-like point at the end. Their irregular widths suggest varying growth rates caused by the unstable pool depth. They are located near a well-traveled route, and a careless passerby damaged some, revealing that they, too, have tiny central canals. Other subaqueous helictites were recently (1990) reported from an arm of Lake Lebarge itself; we have no details about this occurrence.

Figure 3. **Pool fringed by subaqueous helictites below Lake Lebarge (Fig. 1, C). Gypsum blocks and wall crust in background are in contact with water seeping over the flowstone floor into the pool, forcing precipitation of calcite by the common-ion effect. (A. Palmer.)**

Another important locality is a pool about 5 m long, 1 m wide, and 0.3 m deep near the west end of the Sugarlands passage (Fig. 1, D). Several clusters of helictites are located on the side of the pool where seepage from a flowstone-floored crawlway passes beneath gypsum blocks. The helictites grow both from shelfstone and from pool crust well below the water line. They average 1 to 2 mm wide and up to 20 cm long and undulate in all directions. Several come up to the surface, where they terminate in slightly widened "pads." Helictites in one small cluster are much wider than the others, about 7 mm in diameter.

In the eastern branch of the cave, on a balcony west of Ghost Town, a fringe of incipient helictites emanates from shelfstone along the upslope side of a small pool (Fig. 1, E). They resemble those in Pellucidar but are no longer than 2.5 cm. Others up to about 10 cm long are found in a sulfur-bearing passage east of Ghost Town and are reported in a pool above Hoot 'n' Holler Hall (Fig. 1, F). Small examples are also located in the Boundary Waters Section (Fig. 1, G). These locations in the eastern branch also lie downflow from gypsum deposits.

A particularly interesting site is in Old Country, a large room in the far Southwestern Branch of the cave (Fig. 1, H; Fig. 4). In a desiccated flowstone floor is a dry pool slot about a meter wide and deep, and several meters long. Along one wall of the basin are three distinct levels of shelfstone, each bearing pale-yellow helictites clearly of subaqueous origin. The sequence is as follows, from top to bottom: (1) a vertical limestone wall covered with crystalline gypsum crust about 5 cm thick; (2) an old water line below which the gypsum crust has been dissolved away; (3) discontinuous shelfstone 56 cm below the base of the remaining gypsum crust, up to 5 cm wide and 2 m long, composed of, and bearing from its underside, a fringe of helictites; (4) a second shelfstone/helictite level about 5 cm below the first; (5) a third shelfstone level 57 cm below the second and 25 cm

Figure 4. **Helictites of subaqueous origin in Old Country (Fig. 1, H). Vertical extent of photo is about 30 cm. (N. Thompson.)**

above the bottom of the basin. In the upper shelf there are three lesser tiers within a vertical distance of 2 cm.

The helictites emanating from the upper shelf are highly contorted, nearly vertical, and slender (1–3 mm diameter). They extend as much as 28 cm downward. Several make abrupt bends at their lowest extent and grow back upward as much as 15 cm. Unlike the still-submerged helictites elsewhere, a few of these have branching, brush-like termini. The middle shelf is also about 5 cm wide, but its helictites are much shorter (no more than 10 cm) and thicker (6–8 mm). They are more coarsely crystalline, with visible crystals up to 1.5 mm long in their surfaces. The lowest shelf bears only 6 well-developed helictites. They are similar to those of the middle level but slightly shorter (9 cm), thicker (12 mm), and with still coarser crystals (up to 5 mm long).

Vertical helictites also occur on an overhanging ceiling near Blue Velvet Lake (Fig. 1, I). Gypsum crust has been dissolved away beneath a former water level about 6 m above the present lake surface. Just beneath the truncated lower edge of the gypsum are six clusters of tiny helictites up to 5 cm long and 1 mm in diameter. Another water line, 45 cm below, displays discontinuous calcite shelves up to 2.5 cm wide over a distance of more than 6 m. Thin helictites up to 17 cm long and 1 mm in diameter hang from the shelf. Although smaller, they closely resemble the elongate helictites of the upper tier in Old Country (Fig. 1, H).

More recently discovered subaqueous helictites (1990) are in two large basins about 30 m apart, on a balcony north of Ghost Town (Fig. 1, E). One of the pools is dry, and from its shelfstone margin hang helictites that are more varied in size and shape than those in any other locality. The thickest helictite yet seen is about 2 cm in diameter and 8 cm long. A vertical cluster includes helictites 1 cm in diameter and up to 40 cm long, with both upward and downward-growing elements. Others are tangles of tiny "wires," some almost hair-thin and less than 5 cm long. Helictites were fed by water that passed over flowstone covered with gypsum deposits. Calcite rafts also occur in this pool basin, a rare association.

Other recently discovered subaqueous helictites (Jan. 1991) are reported from an extension of the cave off a balcony south of Ghost Town. There are said to be good examples in one active basin and in two others partly or completely dry. Finally, in April 1991, small helictites up to 7 mm long were found in a dry basin, about 60 by 30 mm by 10 mm deep in the Spar City complex north of the Western Borehole.

Subaqueous helictites, either active or inactive, have been observed to date in about 18 locations in Lechuguilla Cave: one in the entrance complex, three in the Western Branch, five in the Southwestern Branch, and nine in the Eastern Branch. Speleothems believed to be subaqueous helictites have also been observed in Spider Cave, 3 km east of Lechuguilla Cave (personal communication, 1990, Kevin Downey, Northampton, Mass., and Roland Vinyard, Johnstown, N.Y.) These are tiny examples, about 1 cm long, and not reported to be associated with gypsum.

Speleothems superficially resembling subaqueous helictites have also been reported from Ceuva de la Puente, San Luis Potosi, Mexico. These have been found to be calcified mucus tubes formed by fly larvae, and have been tentatively termed "Larvites" (personal communication, 1991, George Veni, San Antonio, Texas).

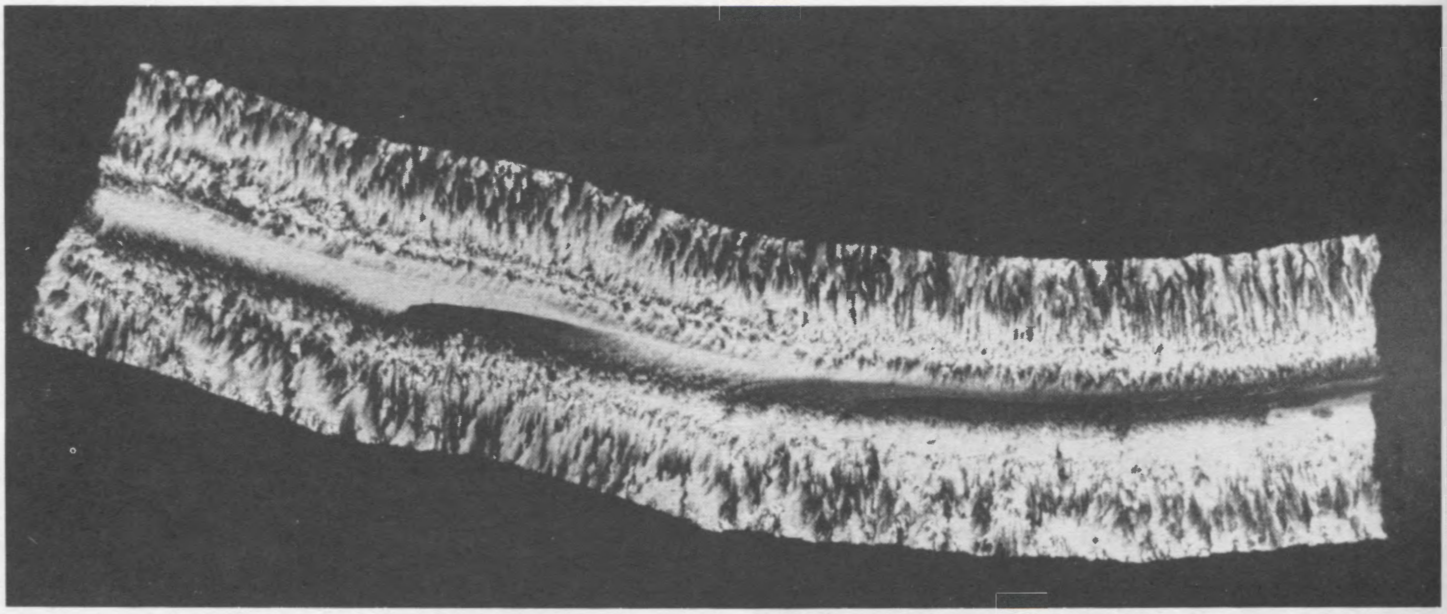

Figure 5. **Thin section of helictite fragment from the Pellucidar pool, photographed in cross-polarized light to delineate crystal structure. Diameter of helictite is 3 mm. (A. Palmer.)**

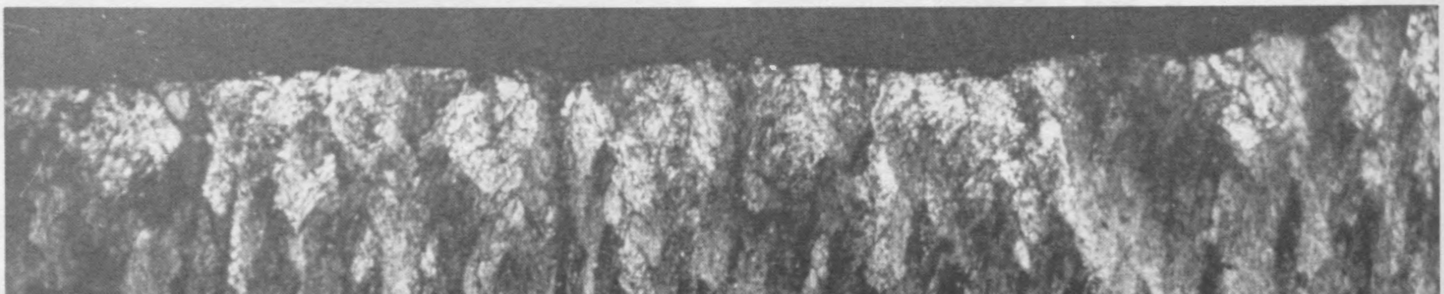

Figure 6. **Feathery crystal structure in the walls of the Pellucidar helictite, in cross-polarized light. The canal is located at the bottom of the photo, and the edge of the helictite at the top. Thickness of wall = 1.5 mm. (M. Palmer.)**

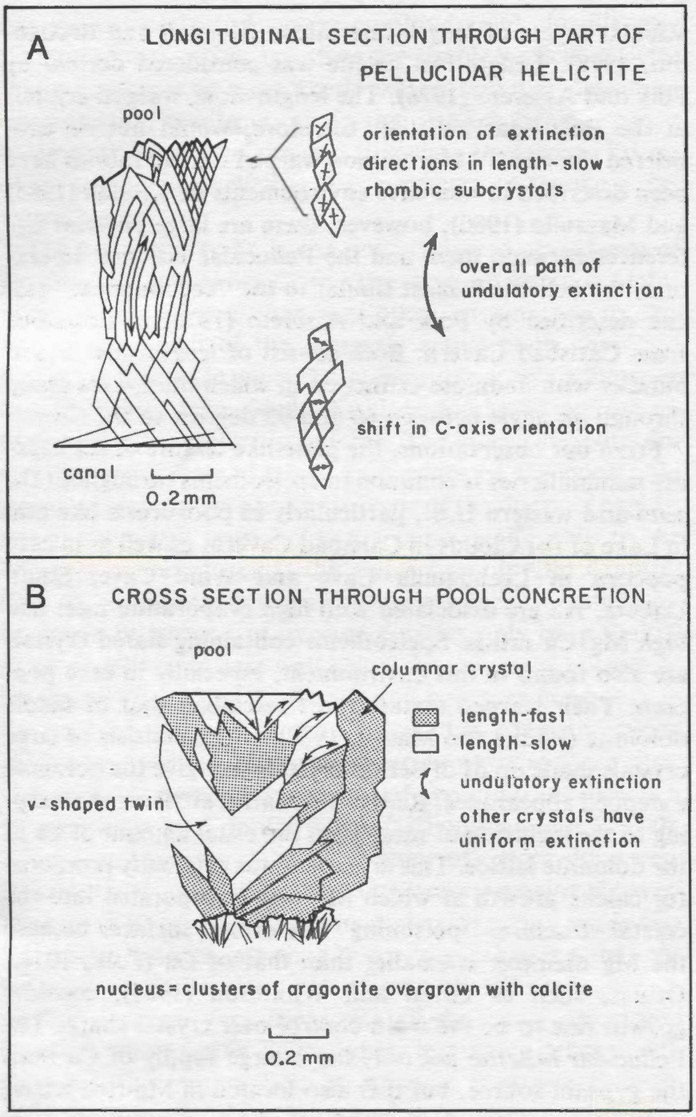

***Figure 7.* Schematic of calcite structure in the Pellucidar helictite (A), compared to that of the concretion from the same pool (B). Arrows show directions of undulatory extinction.**

Mineralogy and Crystallography of the Pellucidar Helictites

A naturally broken helictite fragment 1.5 cm long was obtained from the bottom of the pool at the original discovery site in Pellucidar (Fig. 1, A; Fig. 5). A small calcite concretion was removed from the same pool to provide information about the pool wall deposits. Half of each sample was thin-sectioned for microscopic analysis.

The helictite consists of unlayered translucent calcite surrounding a central canal 0.35 mm in diameter. The canal wall consists of blocky calcite spar. Crystal boundaries in the outer two-thirds of the helictite wall are highly irregular and feathery, but long, slender crystals can still be distinguished (Fig. 6). At high magnification each crystal is seen to consist of elongate rhombohedra bounded by cleavage planes (Fig. 7A). Cleavage angles are approximately 70 to 110 degrees, with the elongation of the rhombohedra roughly perpendicular to the canal. Groups of crystals flare in divergent fans toward the outer surface. At the surface the calcite appears to consist of thin, subparallel plates stacked into crude mammillaries.

Crystals in spelean calcite crusts normally have radial extinction in polarized light; i.e., when the long axis of a crystal is oriented parallel to the plane of polarization the entire crystal darkens simultaneously. As the microscope stage is rotated, crystals that have grown radially from their substrate show extinction that moves from crystal to crystal (Fig. 8A). In contrast, in the Pellucidar helictite the extinction of most crystals migrates in a direction nearly perpendicular to the canal, curving away from the perpendicular toward the outer edge of the helictite (Fig. 7A). This extinction and the flared crystal shape are caused by the C axis of each unit rhomb being slightly rotated with respect to that of the rhomb upon which it grew. Extinction is undulatory, moving in a wave across each crystal as the stage is rotated and appears to move outward toward the growing edge of the crystal.

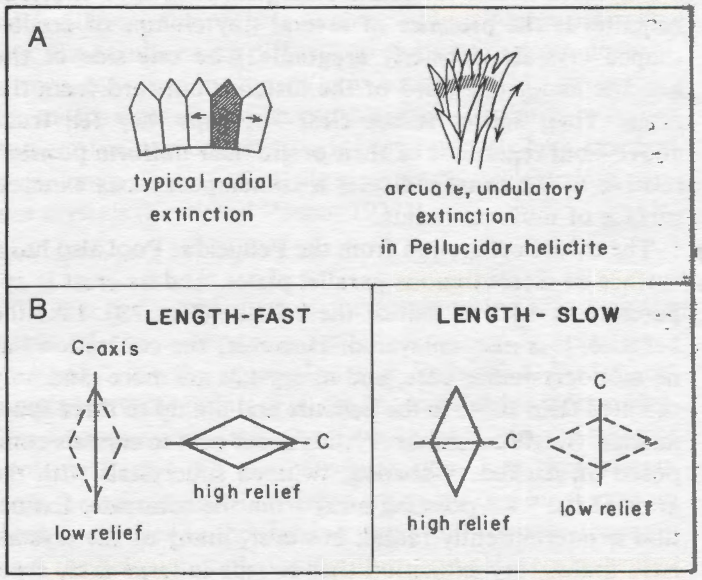

***Figure 8.* (A) Extinction of typical calcite cement is radial; i.e., individual crystals darken as the microscope stage is rotated and elongate crystals are oriented N-S with respect to the stage. Note contrast with Pellucidar helictite. (B) Criteria used to distinguish length-fast vs. length-slow calcite crystals under polarized light.**

The optical relief of cleavage elements in the crystals shows that the entire helictite consists of length-slow calcite (see Folk and Assereto, 1976). In other words, the C axes are perpendicular to the direction of growth (Fig. 8B). Length-slow calcite has been associated with calcite that has recrystallized from preexisting calcite or aragonite. However, the Pellucidar helictite supports the theory of Dickson (1978) that length-slow calcite can be primary.

The nearly uniform helictite diameters show that initial growth must have been fairly constant, with most of the thickening taking place after the helictites ceased to elongate. Apparently the helictites grew as thin crystalline rinds around the exit points for supersaturated water entering the pool. The thinnest helictites in Sugarlands (Fig. 1, D), with rhombic cross sections about 0.7 mm in diameter, probably represent the initial helictite growth form. The original structure was monocrystalline, elongated in the direction of the canal, now overgrown with thin scattered calcite patches. Their fairly uniform dimensions strengthen the inference that growth occurred mainly at the tip.

The tips of the large helictites we have seen are occluded by crystalline calcite, possibly because of diminishing flow through the capillary tubes. It is probable that much of the outward thickening took place by seepage of supersaturated water through the helictite walls after the canal was blocked. This may help to account for the surprisingly uniform helictite diameters.

Further evidence for radial enlargement of the Pellucidar helictites is the presence of several tiny clumps of needle-shaped crystals (formerly aragonite?) on one side of the helictite about one-third of the distance outward from the canal. Their source is not clear—perhaps they fell from above—but regardless of their origin their uniform position relative to the canal indicates a contemporaneous exposed surface of uniform radius.

The calcite concretion from the Pellucidar Pool also has a surface of discontinuous parallel plates, and its crust is apparently as thick as that of the helictite (Fig. 7B). Like the helictite, it is also unlayered. However, the concretion has no monocrystalline core, and its crystals are more randomly oriented than those in the helictite and are up to three times as wide. Single columnar crystals occur next to crystals composed of stacked, V-shaped, twinned subcrystals with the arms of the "V" pointing away from the substrate. Extinction is intermittently radial; however, many of the crystals have undulatory extinction that sweeps in large arcs, especially in the V-shaped wedges at the exposed surface. Length-slow crystals appear to be randomly distributed. Because the thickness of the helictite wall and concretion crust are about the same, the growth rates may have been rather similar, despite the probability that the helictites grew by water seeping from their canals.

Most common speleothems consist of columnar and acicular layers of length-fast calcite (Kendall and Broughton, 1978). Length-fast calcite was considered normal by Folk and Assereto (1976). The length-slow, warped crystals in the Pellucidar helictite, therefore, would not be considered "normal." Uncommon warped cement fabrics have been described in non-cave environments by Kendall (1985) and Mazzullo (1980); however, there are large physical differences between them and the Pellucidar helictite. In texture, the helictite is most similar to the "coconutmeat" calcite described by Folk and Assereto (1976) in flowstone from Carlsbad Cavern. Both consist of length-slow crystal bundles with undulose extinction in which the C-axes sweep through an angle between 60 and 90 degrees to the fibers.

From our observations, the plate-like texture of the helictite mammillaries is common in speleothems throughout the semi-arid western U.S., particularly in pool crusts like that in Lake of the Clouds in Carlsbad Cavern, as well as in cave popcorn in Lechuguilla Cave and Wind Cave, South Dakota. All are associated with high evaporation rates and high Mg/Ca ratios. Speleothems containing flared crystals are also found in this environment, especially in cave popcorn. Their warped texture most resembles that of saddle dolomite (Radke and Mathis, 1980), which consists of large crystals made up of offset subcrystals that give the dolomite a stepped appearance. Radke and Mathis attribute the warping to the inclusion of more than the usual amount of Ca in the dolomite lattice. This argument was originally promoted for calcite growth in which Mg was incorporated into the crystal structure, "poisoning" the crystal surfaces because the Mg diameter is smaller than that of Ca (Folk, 1974). Others, such as Given and Wilkinson (1985), consider growth rate to be the main control over crystal shape. The Pellucidar helictite not only has a large supply of Ca from the gypsum source, but it is also located in Mg-rich water, and it is not yet clear which of the two mechanisms is responsible for warping of the crystals.

THE COMMON-ION EFFECT AS A MECHANISM FOR SPELEOTHEM DEPOSITION

The helictites described in this paper represent a new variety of speleothem that had not been recognized until identified in Lechuguilla Cave. Although they vary in size and minor details, even in different parts of the same site, they all share certain basic traits: (1) an obvious subaqueous origin, (2) association with a present or former water surface, (3) narrow central canals, and (4) growth downslope from a gypsum deposit.

What is the origin of such bizarre speleothems, and why are they so characteristic of Lechuguilla Cave? Ordinary subaqueous speleothems grow by precipitation of calcium carbonate from the pool water and typically line the pool surfaces with a rather uniform, usually mammillary crust. A few exceptions include tower coralloids and chenille spar (Hill and Forti, 1986), but these are simple vertical growths that may require no more exotic explanation than density or concentration gradients, or, in the case of chenille spar, a coating of dependent organic filaments by calcite. To explain subaqueous helictites it is necessary to invoke a different mechanism.

The key seems to lie in the unprecedented abundance of gypsum in Lechuguilla Cave. The cave probably contains more of this mineral than any other known cave in carbonate rocks. All of the presently active pools described earlier are located in a wet flowstone surface only a meter or two downslope from partly dissolved blocks of gypsum up to several cubic meters in volume. The largest helictites invariably grow from the upstream edges of the pools. No gypsum blocks are present near some of the dry pool basins, but gypsum wall crust holds the same relationship to the helictites as the blocks at the active sites. In Old Country (Fig. 4), seepage from the wall crust has been non-uniform, as shown by growth of gypsum rosettes only on a limited patch of otherwise bare wall below the truncated bottom of the crust. The calcite shelfstone and helictites grew *only* in the part of the pool directly below this zone, indicating preferential seepage of gypsum-bearing water at those sites.

A particularly striking example of gypsum truncation at a former water level occurs in the far Southwestern Branch, where fused gypsum slabs on an angle-of-repose slope were dissolved and undercut by standing water, creating an overhang up to 60 cm wide. From its underside grew a single tiny branching helictite less than 5 cm long, showing that the helictites can issue even from a non-carbonate substrate. A similar situation prevails at one of the newly-discovered basins south of Ghost Town, where numerous delicate, filiform helictites, up to about 10 cm long, grew directly from the underside of a water-truncated gypsum-plane overhang along a zone about 5 m long.

Dissolution of gypsum by water rich in dissolved calcite appears to be critical in the development of the subaqueous helictites. The sudden uptake of gypsum boosts the concentration of dissolved Ca^{++}, driving the less-soluble calcite to supersaturation. This is an example of the common-ion effect, in which the dissolution of two minerals that have an ion in common causes the solubility of both minerals to decrease. Gypsum is so soluble that it rarely reaches saturation in the present cave environment without evaporation; but calcite is much less soluble and usually is forced to precipitate when gypsum is dissolved by the cave water. If this supersaturated water enters a cave pool by seeping through pores or as narrow independent streamlets, subaqueous helictites are formed.

If we are correct that the common-ion effect is instrumental in creating subaqueous helictites and associated deposits, it is likely that it plays a role in the origin of certain

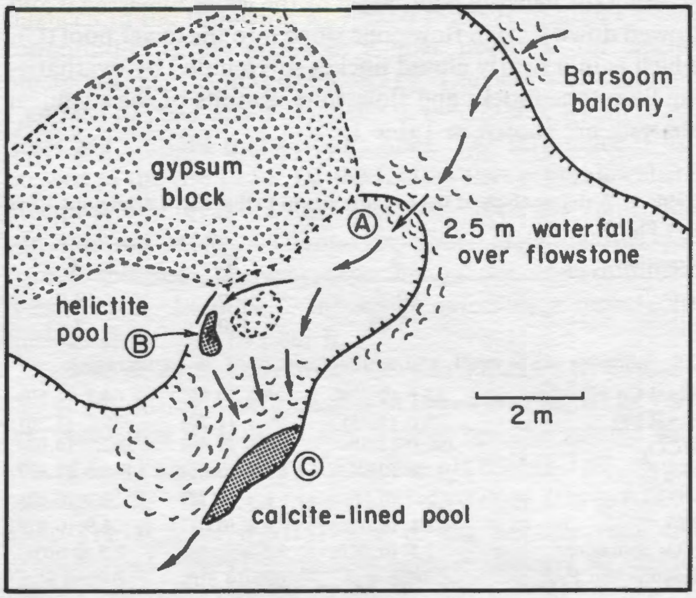

***Figure 9.* Sketch of sample locations for geochemical study of the Pellucidar helictites.**

deposits in other caves that have previously been attributed entirely to carbon-dioxide loss or evaporation. Such deposits are apt to be far more numerous than subaqueous helictites, which are restricted to a narrow range of conditions. Where waters are not sharply differentiated, or mixing is diffuse, the process might simply enhance precipitation, the results being difficult to distinguish from deposits formed by other means. Where conditions are favorable, this mechanism should be considered a possibility in future studies of cave geochemistry. It has previously been invoked to explain a deposit of calcite cave pearls nucleated on gypsum crystals (Forti and Pasini, 1977).

Geochemical Analysis

To test the significance of the common-ion effect on the origin of the subaqueous helictites, water samples were analyzed from the original site in Pellucidar. Water was sampled at three points (Fig. 9): (A) rapid drips from stalactites 4 m upstream from the helictite pool, (B) the helictite pool itself, and (C) a calcite-lined pool 2 m downstream from the helictite pool. A fallen fragment of a large gypsum block is exposed to the flowing water immediately upstream from the helictite pool. Temperature and pH were measured at the sampling sites.

The samples were collected on August 12, 1988, during a rather dry period. Total flow rate was roughly one liter per minute, but this estimate is crude, and is the relative distribution between the two pools, because of scattering of droplets at site A and the thinness of the sheet flow elsewhere. At this time only a small percentage of the water was entering the helictite pool. Most of the water bypassed it and flowed down a steep flowstone slope into the lower pool (C), which is in a nearly closed pocket surrounded by overhanging limestone blocks and flowstone. Results of the chemical analyses are shown in Table 1.

Table 1. Water analysis at the helictite site in Pellucidar, Lechuguilla Cave (see Fig. 9).

Location:	A	B	C
T (°C)	20.2	19.0	19.0
pH	7.87	8.09	7.87
The following are in mg/L and millimoles/L (mM, in parentheses):			
Total Ca	25.6 (0.639)	26.8 (0.669)	64.1 (1.60)
Total Mg	28.0 (1.15)	29.1 (1.20)	29.2 (1.20)
HCO_3^-	206.0 (3.38)	211.0 (3.46)	282.0 (4.62)
$SO_4^=$	13.6 (0.142)	23.8 (0.248)	160.0 (1.67)
Cl^-	5.7 (0.16)	6.4 (0.18)	4.2 (0.12)
NO_3^-	3.1 (0.050)	4.8 (0.077)	4.9 (0.079)
SiO_2 equivalent	2.5 (0.042)	2.5 approx.	2.5 approx.
Equilibrium P_{CO2}	0.0026 atm	0.0016 atm	0.0034 atm
Saturation Indices {SI = log (ion activity product/equilibrium constant)} − = undersaturated, 0 = saturated, + = supersaturated:			
Calcite	+0.10	+0.32	+0.50
Aragonite	−0.05	+0.17	+0.36
Dolomite	+0.21	+0.43	+0.42
Gypsum	−2.72	−2.47	−1.39
Quartz	−0.30	−0.28	−0.28
Opal	−1.28	−1.26	−1.26
Mg/Ca (molar)	1.80	1.79	0.75

Probable error = approx. 1%, except approx. 0.5% for pH and T and 15% for silica.

Evolution of the Pellucidar Water

There is clearly a strong increase in saturation from the incoming water (A) to the helictite pool (B) and the crystal-lined pool (C). This change is remarkable over such a short distance, suggesting that there is indeed a strong control on helictite precipitation by dissolution of gypsum. However, the importance of this process must be compared with three other influences on calcite saturation: carbon dioxide content, temperature, and evaporation.

To sort out these various effects, all the relevant equilibria must be calculated (see any reference on carbonate chemistry, such as Garrels and Christ, 1965), and the major sources of dissolved material must be identified. The bedrock in this area includes mainly calcitic limestone, dolomite, and quartzose siltstone. The bedrock includes some gypsum, but most of it is probably encountered in the cave. The dissolution reactions are:

Calcite, possibly some aragonite:

$$CaCO_3 + H^+ \rightleftarrows Ca^{++} + HCO_3^-$$

Dolomite, high-Mg calcite:

$$CaMg(CO_3)_2 + 2H^+ \rightleftarrows Ca^{++} + Mg^{++} + 2HCO_3^-$$

Gypsum: $CaSO_4 \bullet 2H_20 \rightleftarrows Ca^{++} + SO_4^=$

Quartz and amorphous silica: $SiO_2 + 2H_2O \rightleftarrows H_4SiO_4$

Nitrate is derived mainly from organic sources. Cl^- comes from residual amounts in the bedrock and from organic sources. These probably have negligible effect on the helictite geochemistry.

The incoming water (A) is slightly supersaturated with calcite, as would be expected in the presence of dripstone. The high Mg/Ca ratio (1.8) shows that much of the dissolved load is from dolomite or high-magnesium limestone. However, even pure dolomite would only yield a Mg/Ca ratio of roughly 1.0, so Mg must have been enriched by precipitation of Ca. Evaporation at and near the surface is the simplest explanation. Rain and snowmelt infiltrate into the soil and shallow bedrock fissures, but most is evaporated back into the atmosphere during the ensuing dry period. Calcite and possibly aragonite are precipitated as the solute concentration rises, but precipitation of dolomite is negligible because of its sluggish kinetics. The Mg/Ca ratio of the remaining water is inflated in this way. Some water infiltrates deeper into the bedrock, especially when displaced by the next influx of infiltration. The Mg/Ca ratio is below the threshold of 2 suggested by Folk and Assereto (1976) as a requirement for precipitation of aragonite, which crystallizes instead of calcite when Mg poisons the crystal structure. No evidence of subaqueous aragonite was found at the sampling site.

The slight increase in Mg from A to B, despite the absence of a likely source, suggests enrichment by evaporation. Assuming that no Mg minerals have precipitated, evaporative enrichment was 4.26% from A to B at the time of sampling. This applies to other dissolved species as well and helps boost the SI for all minerals. It is likely that some Mg is incorporated into the calcite pool lining and helictites, which would make the estimate of evaporative enrichment low. Larger but less precisely measured rises in nitrate and chloride from A to B also suggest evaporation, as there is little chance for them to have been acquired between the two sites.

Meanwhile, the P_{CO2} decreased 0.001 atm by degassing, further enhancing supersaturation. However, the slight decrease in temperature boosts the solubility of the carbonate minerals, partly offsetting the other effects. The temperature drop may be caused partly by loss of heat through evaporation.

The stream flows under the gypsum block into the helictite pool (B), acquiring SO_4^{--} from the gypsum and presumably an equivalent amount of Ca (which cannot be distinguished from the Ca supplied by the carbonates). Since all samples are supersaturated with calcite and dolomite, neither can dissolve to increase Ca. The amount of gypsum dissolved between A and B is 0.095 mM (0.0071 cc/L), equal to the measured gain in sulfate minus 4.26% evaporative enrichment. If there were no calcite precipitation, the increase in Ca between A and B should equal the net increase in dissolved sulfate (0.106 mM). However, Ca increases only

0.030 mM, indicating that 0.076 mM (0.0028 cc/L) of calcite has precipitated in and just upstream from the helictite pool. Growth of helictites is the most striking result, although thickening of the calcite pool lining may be volumetrically greater. Depending on the flow rate, varying amounts of water bypass pool B and flow directly into pool C, so it is difficult to estimate the rate of helictite growth. However, it is clear that the helictites grew quite rapidly compared to most other speleothems.

Among all these influences, how important is the common-ion effect? Regardless of the amount of calcite actually precipitated, the *potential* amount of precipitation must be determined. At point A, calcite equilibrium could be achieved by precipitation of enough calcite to reduce the total dissolved Ca to 0.584 mM. Using this figure as a basis of comparison, from A to B the potential amount of precipitation caused independently by each of the controlling factors is summarized below:

1. Evaporation, by reducing water volume, would potentially cause 0.025 mM (0.00093 cc/L) of calcite to precipitate.
2. CO_2 loss would cause the dissolved Ca at calcite saturation to drop to 0.435 mM, for a potential of 0.149 mM (0.0055 cc/L) calcite precipitation.
3. The drop in temperature would increase the total Ca at calcite saturation to 0.603 mM, which would require dissolution of 0.019 mM (0.00070 cc/L) of calcite.
4. Dissolution of gypsum by calcite-saturated water would potentially allow one equivalent of calcite to precipitate for each equivalent of gypsum dissolved. Between A and B the dissolution of gypsum (adjusting for enrichment by evaporation) causes a potential for 0.095 mM (0.0035 cc/L) of calcite to precipitate.

Therefore, in the helictite pool, and slightly upstream from it, the three positive factors cause a total potential calcite precipitation of 0.297 mM (0.011 cc/L), in the following proportion: evaporation = 9%; CO_2 loss = 55%; and common-ion effect = 36%. During higher flow, when a greater percentage of the water from point A enters the helictite pool, the common-ion effect would account for a greater proportion of the potential calcite precipitation.

The importance of the common-ion effect is greater than suggested by these figures. Evaporation and CO_2 loss are both slow, whereas the uptake of gypsum causes a rapid burst of supersaturation. Calcite immediately begins to precipitate, resulting in shelfstone and helictites that grow directly downstream from the gypsum block. The thin film of supersaturated water is concentrated in isolated tendrils as it enters the pool and begins to precipitate calcite before it has a chance to mix with the pool water. CO_2 loss and evaporation probably contribute to coating the walls of the pool and helictites, but not to the origin of the helictites.

By similar analysis, precipitation of calcite in pool C is influenced in the following way from point A to pool C: The common-ion effect produces a huge potential for calcite precipitation, 1.53 mM (0.057 cc/L). The net potential is for 1.44 mM (0./053 cc/L) of calcite to precipitate, which is 2.5 times the amount that actually does. The difference is expressed in the rise in SI. Of the factors that promote calcite precipitation, evaporation accounts for 2% and the common-ion effect accounts for 98%. Helictites do not form in this pool, however, showing that the mode of inflow is important. In pool C the water enters as a more continuous sheet than in pool B.

Sample A was rapidly losing CO_2, as indicated by a steady rise in pH during measurements in the cave. The pH of 7.87 was measured directly in the drip. When the sample was stirred in a breaker for a minute, the pH rose to 7.97, representing a loss of CO_2 in the water from 0.0026 atm to 0.0021 atm. The P_{CO2} of 0.00157 atm in the helictite pool (B) probably represents the ambient value in the cave air, as the water would have had ample opportunity to equilibrate with the air. This is verified by an almost identical P_{CO2} (0.00159) for a pool in the passage below Pellucidar (Palmer, 1988). The mean P_{CO2} in all pools measured in Lechuguilla Cave to date is 0.0025 atm.

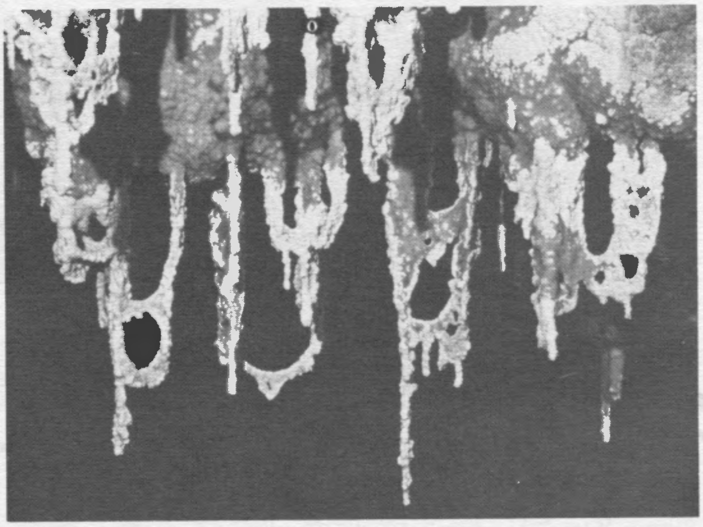

***Figure 10.* U-loops above the Bitter Water Pool (Fig. 1, K). Width of photo = 30 cm. (A. Palmer.)**

OTHER SPELEOTHEMS OF PROBABLE SUBAQUEOUS ORIGIN

Pool Fingers and U-loops

The Bitter Water Pool in the Southwestern Branch is located in a chamber 12 m in diameter (Fig. 1, K). From parts of the yellowish mammillary crust lining the walls of this room extend beard-like groupings of hundreds of slightly wiggly, slender calcite fingers up to 30 cm long and about

Figure 11. **Pool fingers with web-like interconnections at site L in Figure 1. Height of photo = approximately 30 cm. (N. Thompson.)**

1.5–6 mm in diameter (Fig. 10). Most are essentially vertical, but others incline as much as 20 degrees toward the center of the chamber. Many are connected by U-shaped inverted arches, spanning average distances of about 5 cm, which look as if shoestrings had been draped between the fingers and calcified. These decorations are now inactive, but it appears that they grew underwater and are not subaerial stalactites. They are all located below the former water level marked by the calcite wall crust and share the color and surface texture of the crust. Many taper to dimensions far less than the 5 mm normal minimum diameter of a soda-straw stalactite (Curl, 1972). Also, there are no ordinary stalactites associated with them. The U-loop connections do not have straws hanging from the bottoms, as might be expected if water had dripped from them.

In a preliminary description, Davis (1988b) called them "stalactoids" because they resembled stalactites but had not grown by a stalactitic mechanism. However, this resulted in some confusion with stalactites, and in this paper we use the more descriptive but tentative term "pool fingers."

A second pool-finger locality has been found in a dry shelfstone pool, about 1.3 m wide and 2 m deep, in a basin downslope from old flowstone in the Southwestern Branch (Fig. 1, L). A dense display up to 25 cm long extends primarily from the lower edges of chenille spar about 60 cm below the main shelfstone level. Like the original fingers, they share the color of adjacent subaqueous crust, often are narrower than soda straws, and do not associate or intergrade with ordinary stalactites. In the new locality they are not only linked by U-loops, but some entire groups are interconnected by perforated, web-like sheets, which are paper thin and lighter colored than the fingers, and apparently only lightly attached to them (Fig. 11). These webs are probably lightly-calcified bacterial mats.

Figure 12. **Rounded and pointed pool fingers in the Eastern Extension (Fig. 1, M). Length up to 50 cm. (P. and A. Bosted.)**

The finest display of pool fingers is in the Eastern Extension (Fig. 1, M; Fig. 12) in a large irregular basin up to 15 m wide and deep. In several alcoves of this chamber, dense growths of brown fingers, many up to 50 cm long, hang from nearly flat ceilings coated with rough mammillary crust. In the most profuse grouping they are spaced only about 2 to 6 cm apart over an area some 5 m across. Where attached, most are up to 1.5 cm wide, with a tendency for the top few centimeters to curve slightly toward the center of the chamber, becoming more vertical below. Unlike those in the Southwestern Branch, these fingers show two distinct morphologies: some taper to needle-like points, but others interspersed apparently at random are more cylindrical, with rounded tips 1 cm or more across. Some of the rounded ones seem to have engulfed pointed ones, often leaving several centimeters of the pointed tips protruding from the overgrowth. Another puzzling feature is a slightly darker-brown stripe down the side of each finger that faces the center of the basin, suggesting some physiochemical gradient in the formative water. Lateral connections between fingers are varied and abundant, including U-loops and stacked and interwebbed loops.

About 100 m from here, in a dry basin below Lost Pecos River, are the smallest pool fingers observed in the cave, consisting of vertical fingers only about 1 mm wide and up to 7 cm long, with miniature U-loop connections 1 cm long.

Since pool fingers usually occur well below the margins of their basins, they are not as easily observed in full lakes as are subaqueous helictites. David Bunnell (Goleta, CA, personal communication, 1990), during a scuba dive in 15-m-deep Stud Lake in the Eastern Branch, observed an area of fingers more than 20 cm long at a depth of 3 to 11 m (Fig. 1, N; Fig. 13). They are located along a single wall and angle toward the larger part of the lake. Photos show a similarity to pool fingers, but U-loop connections seem to be absent. They are also broader-tipped than the fingers at other sites. They may have formed on bacterial filaments in a water current strong enough to sweep them laterally and to prevent cross-attachments.

Cave-pool fingers, especially those at site L in Figure 1, have superficial similarities to subaqueous helictites, notably those at sites H and J. Both include "beards" of slender, wiggly, downward-growing projections and overlap in general size and shape. However, the two species differ in four ways: (1) pool fingers have no central canals; (2) pool fingers are not integral with shelfstone levels; (3) U-loop and weblike connections are associated only with pool fingers; and (4) pool fingers are not located beneath obvious sources of gypsum-rich seepage.

The cross sections of naturally broken fingers show a fine-grained radial structure. They contain a central crystalline zone with opaque, very finely crystalline patches. Opaque patches of this kind may be of bacterial origin (Kuznetsov and others, 1962).

Microscopic examination of a naturally broken pool finger from the Bitter Water Pool area verifies that the pool fingers and U-loops originated by subaqueous accretion of calcite around strands of organic filaments. Slender organic

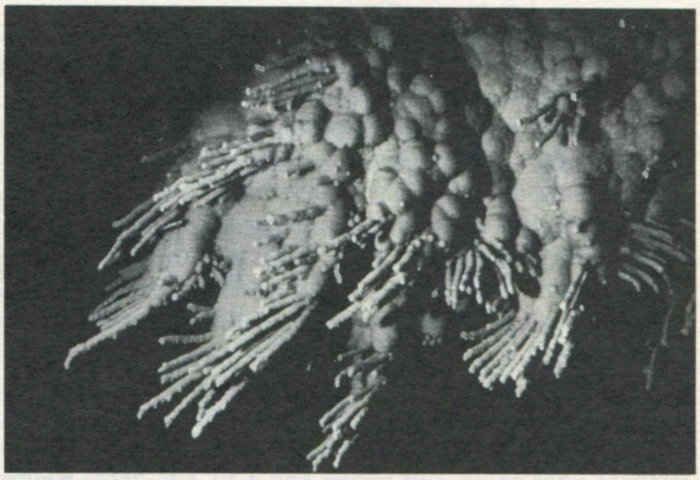

Figure 13. **Submerged pool fingers in Stud Lake, at a depth of 4.5 m (Fig. 1, N), photographed during a dive by David Bunnell. Length of fingers is about 20 cm.**

filaments hang from the tips of some of the fingers. If the strand is attached at one end, a finger forms; if at both ends, a U-loop. The source of the organic strands is still uncertain, but they are probably bacterial.

The center of the pool finger is lined with an irregular black band 5 um thick that consists of very fine crystals (0.25 um) of nonreflective material. It probably consists of either organic carbon or a manganese oxide. The black band is covered with several thin layers of translucent calcite.

On the floor crust beneath the greatest concentration of pool fingers at site M in Figure 1 are scattered white twisted strands, no more than 1 mm wide and up to 2 or 3 cm long. They seem to be segments of organic filaments that fell before becoming coated and were calcified on the floor. Inactive flowstone where water once drained into the basin shows varicolored parallel lines on its surface. Some of the lines terminate in wavy points that appear to be calcified bacterial streamers like those in certain mineral springs or mine-drainage waters.

The Bitter Water Pool is a remnant of a formerly much larger pool in a calcite-encrusted basin. The pool is now about 2 m in diameter and 15 cm deep. When tasted, this water was found to be a bitter brine. An analysis by Gregg Oelker (Pasadena, CA, personal comm., 1988) gave the results shown in Table 2.

Table 2. Chemical analysis of Bitter Water Pool, in mg/L (G. Oelker, 4/29/88).

pH	8.4	HCO_3^-	1330.0
CA^{++}	57.3	CO_3^{--}	60.1
Mg^{++}	5690.0	Cl^-	4250.0
Na^+	4160.0	SO_4^{--}	22500.0
K^+	313.0	NO_4^-	800.0

The Bitter Water Pool lies downslope from a series of rimstone pools in the Lebarge Borehole in the Southeastern Branch, which have the composition of normal dripwater. It is likely that the strange composition of the Bitter Water Pool—in which the more soluble constituents are concentrated about a thousandfold compared to pools upslope—is the result of lengthy evaporation of the terminal pool in the series. When the basin was full, the water may not have been exceptionally mineralized, and it is possible that evaporative concentration contributed to the growth of the pool fingers and U-loops.

Like the Bitter Water Pool, the basin at site M in Figure 1 was a deep sump at the downflow end of a series of rimstone/shelfstone pools more than 100 m long. It contains a remnant pool about 5 m across and 1 to 2 m deep, whose water may also be concentrated by evaporation but, unlike the Bitter Water Pool, is still potable. Its surface pH measured 8.0 when sampled on-site.

Related growths have been seen in other caves. The "Crown Jewels," in a vug in Wind Cave, South Dakota, are fingers connected by U-loops, but are composed of crystalline quartz overgrowths on black filament cores. Rudimentary fingers up to 5 cm long cover a lake-basin ceiling in Holey Sheep Cave, Wyoming, but lack U-loops. Miniature fingers up to 2 cm long, with 1-cm U-loops, line a 30-cm-wide pool basin in Huccacove Cave, Colorado.

In 1990, a fine display of white fingers, to about 30 cm long, was found in Falling Rock Cave, a small, shallow cave in a travertine deposit in northern New Mexico (Bill Heath, Taos, N.M., personal communication). U-loops are not prominent here, but the fingers are associated with calcified roots—consistent with our proposed mode of origin.

Cave pool fingers also seem to intergrade with chenille spar, which are thin, vertical, adjoining columns of calcite growing just below shelfstone. Thin organic strands up to 1 cm long are suspended from the ends of chenille spar columns in a pool in the New Mexico Room of Carlsbad Cavern, implying an organic origin for this speleothem.

Iron-Oxide Stalactites

The third species of speleothem considered here is an aberrant form of stalactite of possible subaqueous origin. These deposits are also the most complex and cryptic. Their subaqueous growth has not been so decisively established as that of the helictites and pool fingers. They are located in a bizarre terrain—itself poorly understood—which includes about a kilometer of phreatic maze passage in low-lying areas about 350 m below the entrance in the Eastern Branch (Fig. 1, O). Wall limestones and breakdown have been spectacularly eaten and etched into chains of pothole-like cavities connected by shallow, partly sinuous channels, in places overlain by smaller-scale dendritic rillenkarren. This profound corrosion is not ordinary vadose stream channeling. There are no stream-laid sediments, graded floors, or other normal vadose features. Like rilled terrain of different morphology elsewhere in the cave, the pitting appears to have been caused by intense corrosion where acidic vapors condensed above remnant basins of water late in the drainage phase of the cave. Such intense condensation corrosion may imply that the water was hot.

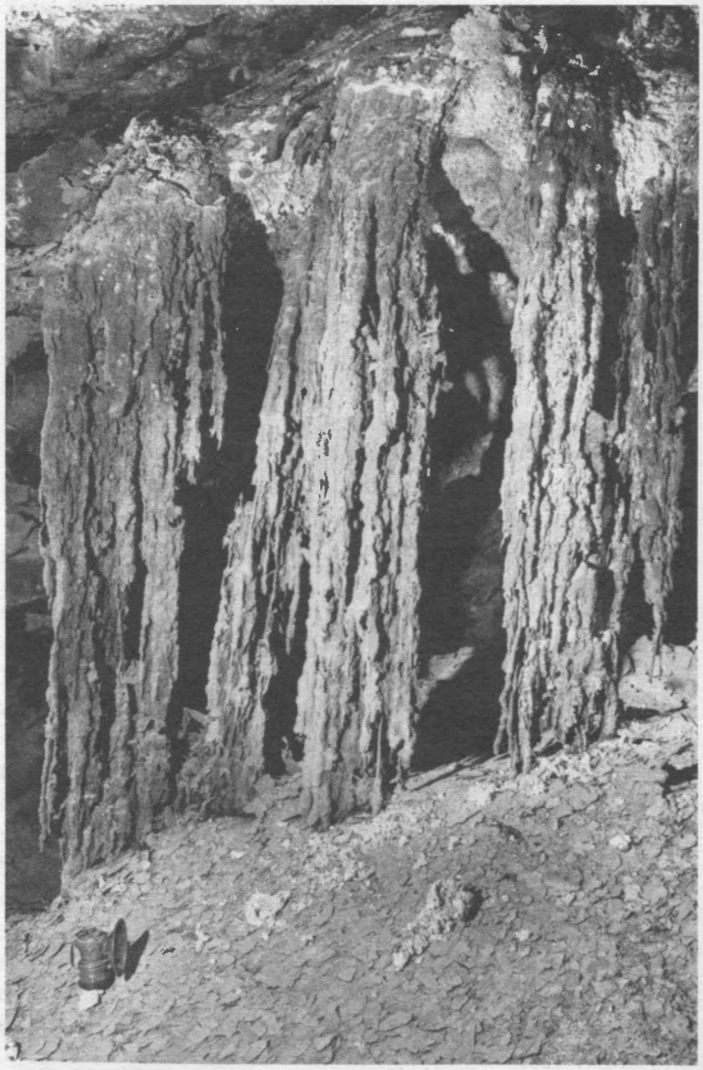

Figure 14. **Iron-oxide stalactites and columns coated with calcite (see Fig. 1, O). Height of columns is 1.3 m. (A. Palmer.)**

The lowest 6 m of this zone is stained blackish below a distinct water line, showing that mineralized water was once ponded there. Below this line are calcite rafts up to 30 cm deep on the floor. Part of this old pond basin is occupied by groups of ropy, skeletal-looking stalactites and columns up to 1.5 m long, issuing from ceiling joints or from ledges mantled by calcite rafts (Fig. 14). They are not only much

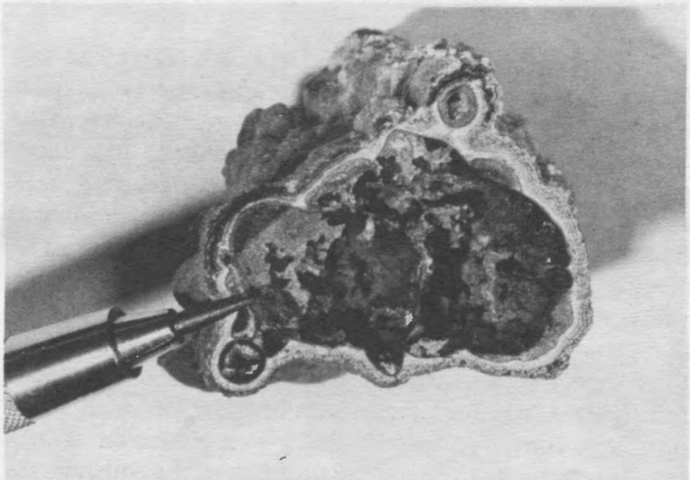

Figure 15. **Cross section of a naturally broken fragment of iron-oxide stalactite, showing friable center and hard calcite coating. Width of stalactite = 3 cm. (A. Palmer.)**

larger than subaqueous helictites or pool fingers, but have true stalactitic form. They resemble the iron/manganese speleothems described by Peck (1986). Some segments are deflected from the vertical by 10 or 20 degrees but never grow upward. Unlike pool fingers, they have internal canals.

Their dimensions are strikingly odd. Some slender ones maintain almost uniform diameters as small as 2 mm over lengths up to 45 cm. This is far smaller than the diameter of a water drop, which normally controls the minimum diameter of a soda straw, and would seem to preclude growth from dripping water, at least at the temperature and composition of the present cave water. In larger examples the internal canals are generally large and irregular, up to 2.5 cm in diameter, and in places are open to the outside through windows in the columns. Large irregular voids in the centers of the columns probably formed by loss of friable iron oxide (Fig. 15). In places the centers consist of a loose aggregate of iron oxide interspersed with 1-mm-wide

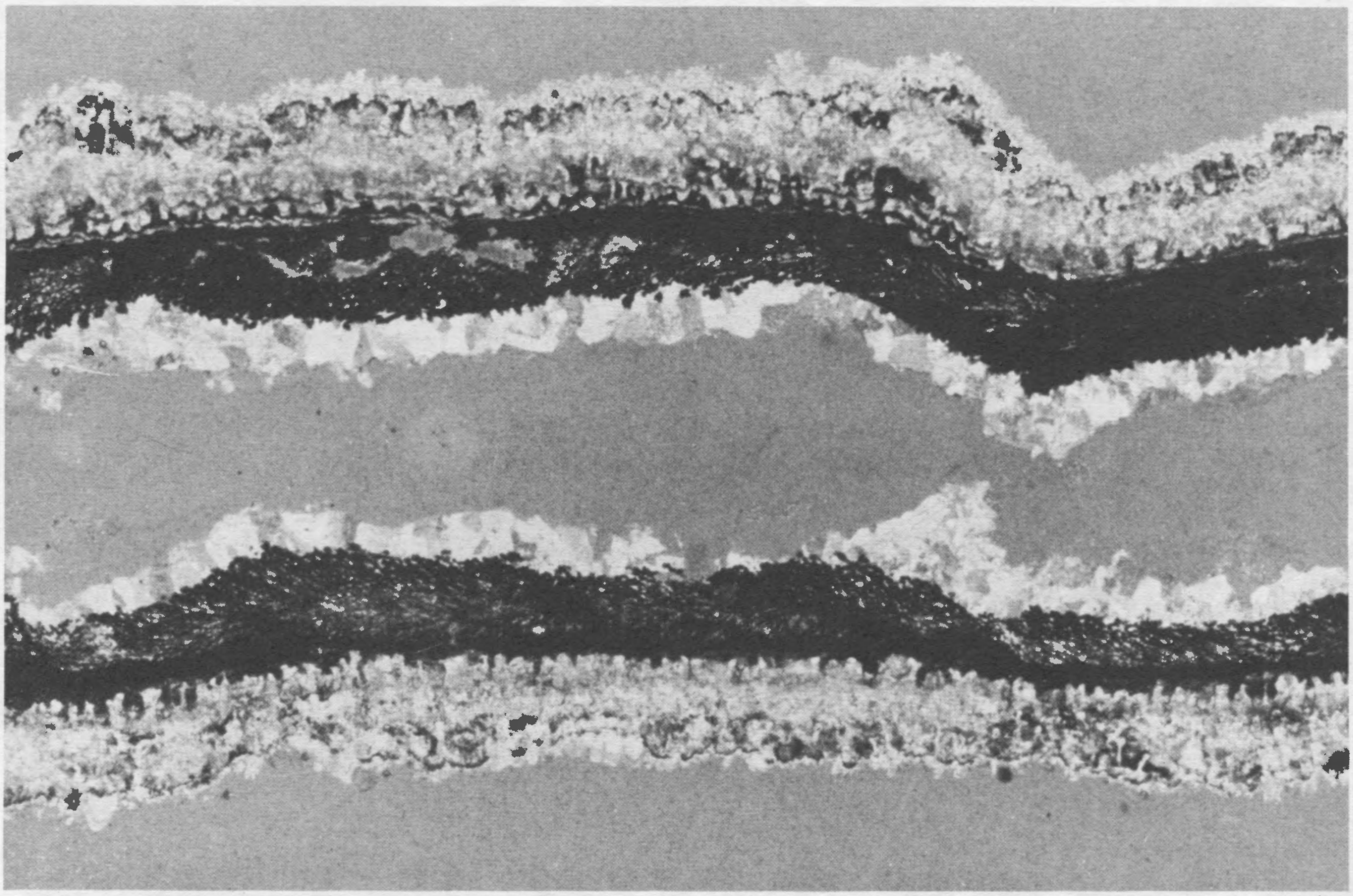

Figure 16. **Thin section of an iron-oxide-rich soda straw stalactite at site O in Figure 1. Diameter = 5 mm. The iron oxide, which forms the dark layer, coats a tangle of organic filaments. That layer is coated both inside and out by calcite. (M. Palmer.)**

calcite crystals. Some (but not all) of the larger columns are aggregates of fused and somewhat braided "straws" that have adhered together without fully coalescing, as ordinary stalactites would.

Some columns have at their bases stalagmite-like mounds up to 10 cm high, but these show a peculiar lobate structure reminiscent of the overflow of a candle, not the smooth or terraced surface of typical drip-formed stalagmites. These mounds generally occur only where a column has touched the floor—not as independent stalagmites such as grow beneath a dripping stalactite. Of three possible exceptions, two are crowned by crooked straws up to 10 cm long, suggesting that they, too, were once columns that have since broken. No broken fragments were seen, but these may have been buried in raft deposits. The stalactites themselves show no drapery-like forms, even beneath ledges, where such shapes would be expected if they were true dripstone. Despite their obvious similarity to subaerial dripstone, such as internal canals and stalactitic form, their braided morphology, lobate bases, filament-lined walls and often very wide diameter canals suggest an absence of surface tension during their formation.

These speleothems are coated with layered calcite and lined internally with red iron oxide and organic filaments. (Fig. 16). The filaments are sub-parallel and extend into the central tubes at various angles (Fig. 17). Each filament is only 1–6 um in diameter at the base, but enlarges to as much as 6–9 um in the direction of growth (Fig. 18). According to Buchanan and Gibbons (1974), the filaments appear to be those of sheathed iron-fixing bacteria, probably *Clonothrix,* whose filaments can have false branches with swollen tips that grow up to 1 cm long on firm substrates. During growth the filament sheaths become lined with ferric hydroxide or ferrihydrate, which enlarges them considerably, and which in time is converted to hematite. For a discussion of bacterial deposition of iron and manganese in speleothems, see Peck (1986).

Clonothrix and similar bacteria thrive at rather neutral pH in oxygenated water, but oxygen concentrations can be as low as 0.1–1.0 mg/1. The filaments do not belong to the

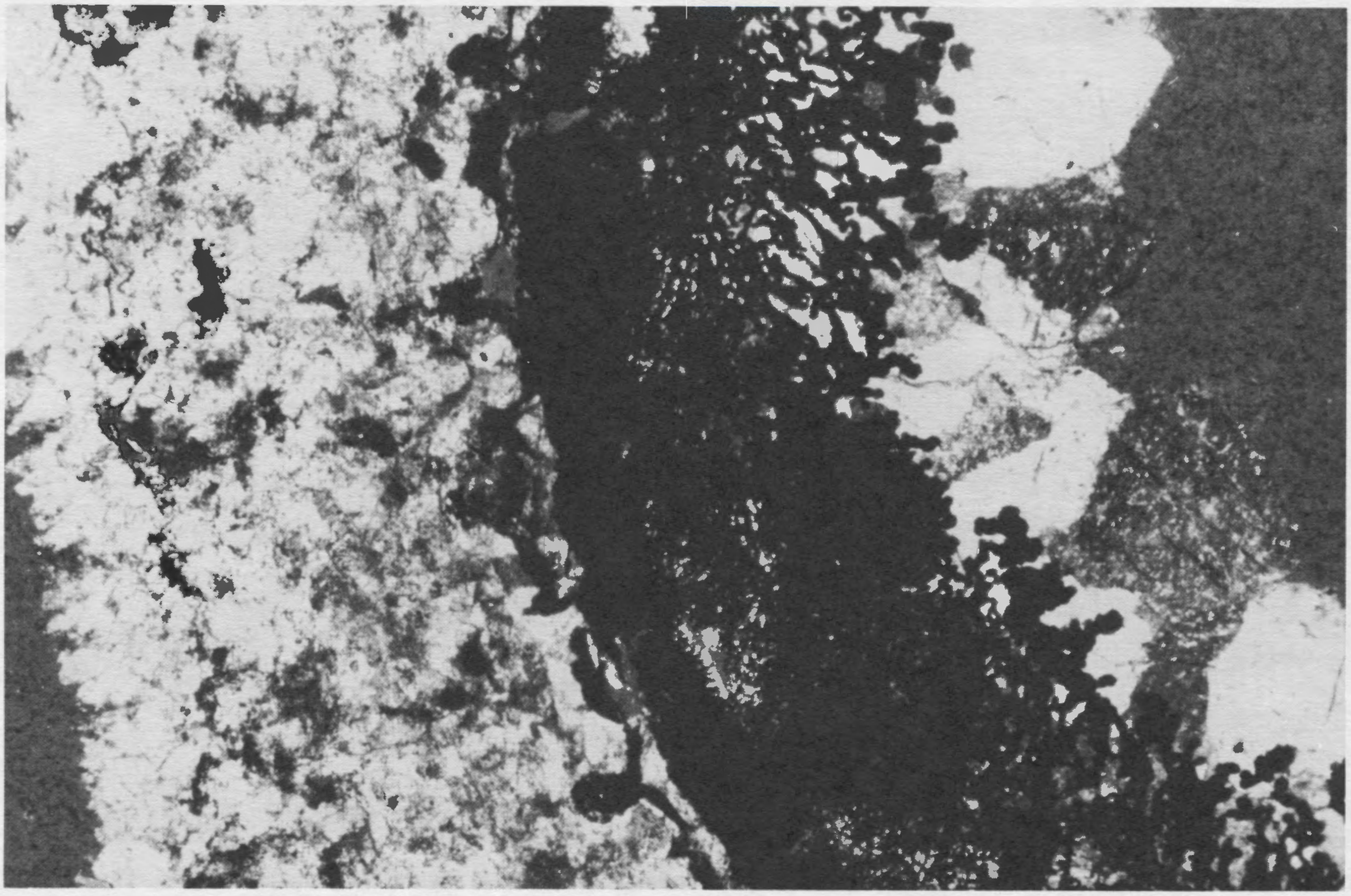

Figure 17. **Cross section of the iron-oxide-rich soda straw in Figure 15, showing growth pattern of bacterial filaments. (M. Palmer.)**

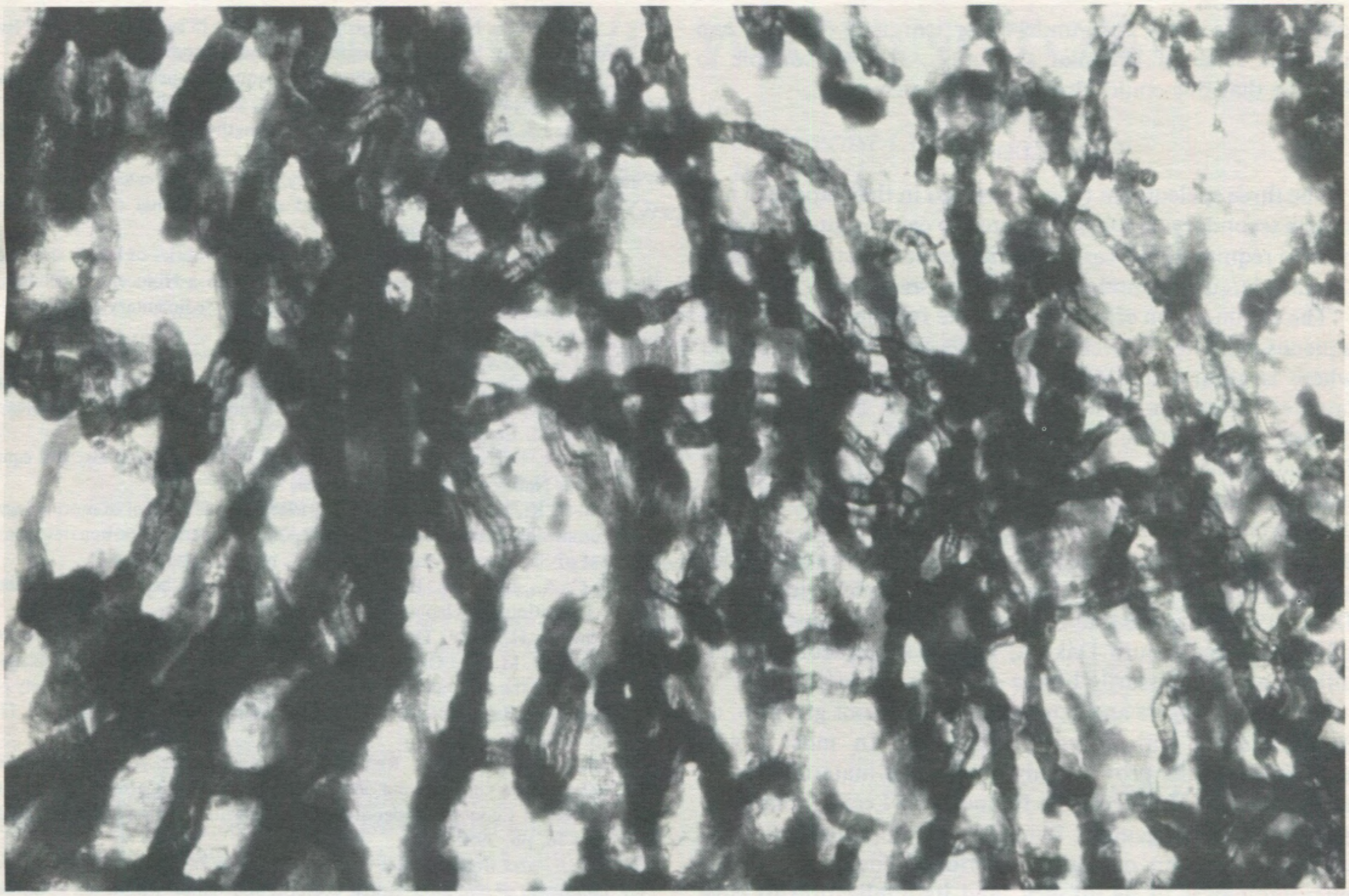

Figure 18. **Photomicrograph of bacterial filaments in the sample shown in Figure 15. Average filament diameter = 3–20 um. (M. Palmer.)**

sulfur oxidizer *Thiobacillus ferrooxidans,* which consists of tiny rods and grows in very acidic water.

The iron-oxide stalactites occur only below the old water line, in the calcite-raft zone, and share the dark staining of the walls. None occur above the water line. Some have either engulfed calcite rafts or have rafts adhering, showing that they were immersed in water either before or after their growth was complete. This, as well as their morphology, suggests a subaqueous origin. However, it is not certain that all the evidence of submersion dates from the period of actual growth. The pothole-like terrain extends below the raft-and-stalactite levels, showing that at least one drainage episode preceded the final flooding. Some of the rafts could have accumulated during a reflooding episode and may not represent the conditions in which the iron-oxide speleothems formed.

The iron-oxide speleothems directly underlie a passage a few meters above, in which leached bedrock is so heavily coated with a residue of iron oxide as to resemble weathered iron ore. The iron oxide deposits were probably derived from oxidation of iron sulfide (e.g., pyrite) in the bedrock. Exposed to oxygen in an aqueous environment, iron sulfide alters to iron hydroxide, which reverts in time to iron oxide. In the oxidizing environment necessary to form these speleothems, iron is soluable only in highly acidic water. Sulfuric acid produced by oxidation of iron sulfide or hydrogen sulfide during cave origin may have been abundant enough to mobilize the iron, which would quickly have precipitated as iron hydroxide when contact with carbonate rock or bicarbonate-rich water raised the pH.

It is suggestive that some iron-mineral "stalactites" actually observed growing have indeed been subaqueous—the marine "rusticles" found on the hull of the sunken ship *Titanic* (Pellegrino, 1988). These oceanic growths, arising in salt water from metallic iron, cannot be identical to the cave variety, but some of the pictures show a remarkably similar morphology. The cave forms are closer in appearance to the "rusticles" than to most cave stalactites. At least the marine observations show that underwater iron-rich deposits can take the form of irregular stalactites. It is reasonable to assume that this could also happen in a cave.

CONCLUSIONS

The three speleothem types described in this paper are not simple secondary deposits unrelated to cave origin. They appear to require, or at least thrive in, environments unique to the sulfide/sulfate mode of speleogenesis.

The common-ion effect, responsible for the origin of subaqueous helictites in Lechuguilla Cave, can only be expected where large amounts of gypsum are encountered by carbonate waters. In most caves, gypsum is an evaporative deposit incompatible with flowing water. Only in caves formed by the hydrogen sulfide/sulfuric acid mechanism is it likely for massive amounts of gypsum to be encountered by flowing calcite-saturated water. The discovery of these unusual speleothems has stimulated a reinterpretation of similar features elsewhere. For example, field evidence by Davis (1990) suggests that helictite bushes in the lower levels of Wind Cave, South Dakota, have a subaqueous origin, although not necessarily by the common-ion effect.

Cave-pool fingers and iron-oxide speleothems grow on, or in association with, organic filaments. In most caves, organic filaments grow mainly on organic material carried in by streams. However, the filaments in Lechuguilla Cave appear to be indigenous. Some, or all, apparently derived their energy from oxidation reactions. The field of geomicrobiology is in its infancy, and its application to cave processes shows great promise for future studies of cave (and other geologic) paleo-environments.

ACKNOWLEDGEMENTS

We wish to thank Paul Rubin for assisting with the collection of the Pellucidar water samples and James Whitney of the Illinois State Water Survey for analyzing them for $SO_4^{=}$, Cl^-, and $NO_3^{=}$. Rick Olson verified the identification of the iron oxide speleothems with SEM-EDX analysis and helped in editing the text. Gregg Oelker provided the chemical analysis of the Bitter Water Pool. Norman Thompson, David Bunnell, and Peter and Ann Bosted supplied several of the photographs.

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PALEOMAGNETISM OF SPELEOTHEMS IN GARDNER CAVE, WASHINGTON

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Cores were extracted from a Late Quaternary stalagmite and a flowstone from Gardner Cave, Washington. Dating by $^{234}U/^{230}Th$ methods give an age range from 20 to 4 Ka.

The stalagmite paleomagnetic record tentatively agrees with paleomagnetic directions previously reported for Fish Lake in southeast Oregon from 12 to 4 Ka. Flowstone paleoinclinations tentatively correlate with the Fish Lake record from 8 to 4 Ka. The flowstone paleodeclinations are shifted 40–80° eastward relative to the Fish Lake record, although the flowstone and cave sediment records show "parallel" declination changes to the west from 5 to 2 Ka. Agreement between the speleothem data and the geocentric dipole field suggests dip-slope bias is not present. A chemical remanent magnetization is inferred. Intensities of magnetization of both speleothems are about 10^{-3} A/m and 100 times weaker compared with the glacially derived cave sediment.

Agreement exists between the Gardner Cave sediment paleomagnetic record (N = 17, n = 8) and the Fish Lake record but the Gardner Cave record maybe shifted earlier in time relative to Fish Lake. The Gardner Cave sediment record is bracketed by two corrected ^{14}C dates at 6.7 to 2 Ka, implying mid-Holocene flushing of glacial sediment through the cave.

INTRODUCTION

Paleomagnetic field directions can be recorded in speleothems if ample magnetic minerals are present (Latham et al., 1982). Such minerals are usually incorporated with calcite during speleothem genesis and become aligned parallel to the geomagnetic field. Recent papers (Latham et al., 1987; 1986) show speleothem paleomagnetic records may supplement other existing records. This study examines the remanent magnetization (RM) of two calcite speleothems and also detrital sediments in Gardner Cave, Crawford State Park, northeast Washington state (Fig. 1).

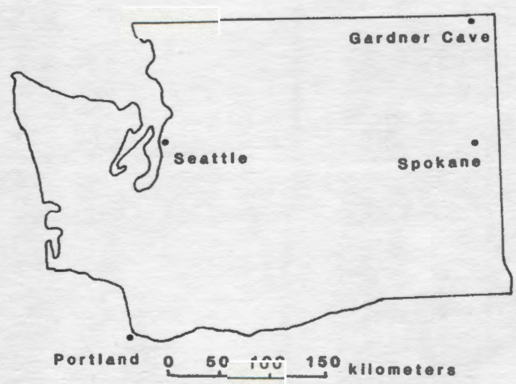

Figure 1. **Location map of Gardner Cave.**

DESCRIPTION OF SAMPLES

A Pomeroy core drill extracted pairs of cores from the Dry Stream Stalagmite (40 cm tall, 90 cm diameter), or DSS, located 60 m downslope of the entrance, and a flowstone floor located 170 m downslope of the entrance. One core from each pair was analyzed for RM and the companion core $^{234}U/^{230}$ Th dated by Dr. Derek C. Ford.

The DSS core revealed creamy microcrystalline calcite, ten major growth laminae (> 0.5 mm) or depositional hiatuses (distinctive detrital layers), and numerous minor growth laminae (Fig. 2). A thin section revealed that the dark layers do not appear to be airborn dust particles, but as water-deposited detrital material—limonite or iron clay for example (R. Boggs, personal comm., 1987). Further petrographic study is needed to determine the precise composition. The flowstone core exhibited macrocrystalline (> 1mm) vuggy calcite and seven major growth laminae (Fig. 3), although the seventh growth laminae is not shown.

The Grotto, located 100 m downslope from the entrance, is a small open chamber adjacent to the main passage. Grotto sediments cover the floor as a 2 m sequence of alternating clay and fine sand laminae.

METHODS

Harmon et al. (1975) list criteria which determines the suitability of speleothems for U/Th dating. The processed samples met these criteria except the contaminant ratio

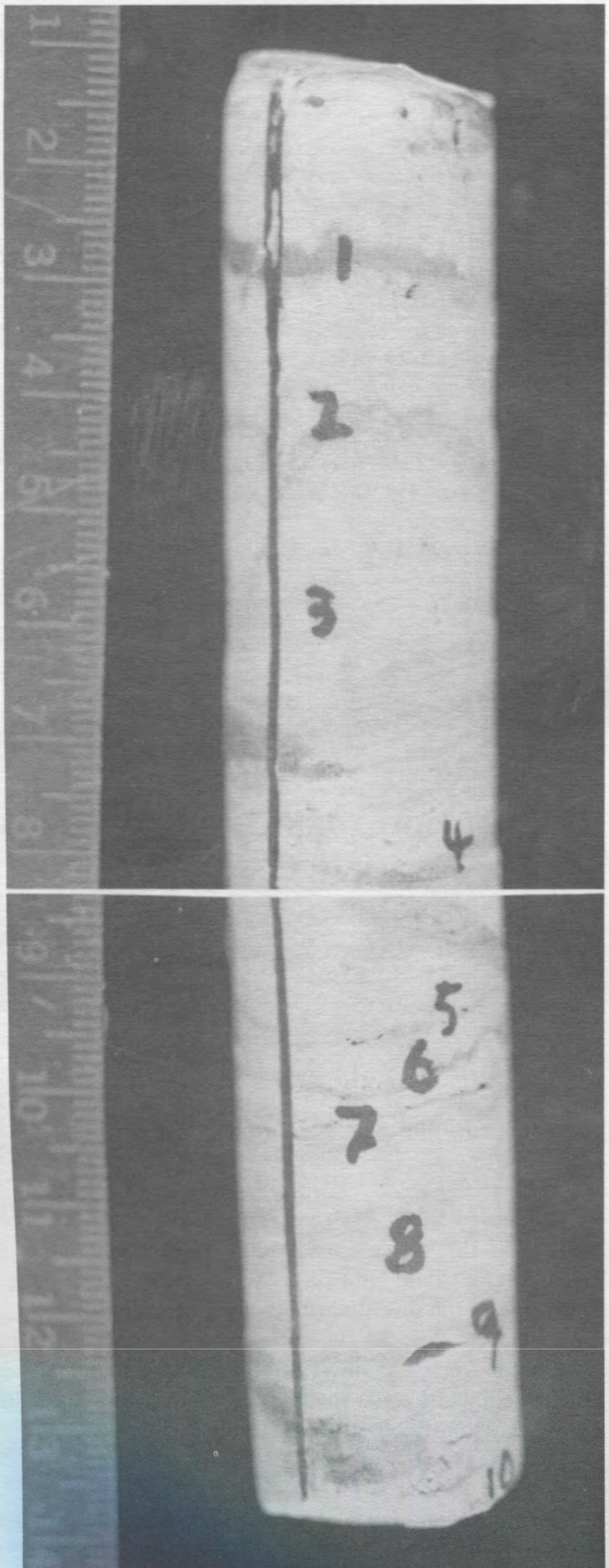

Figure 2. **Sample core from the Dry Stream Stalagmite.**

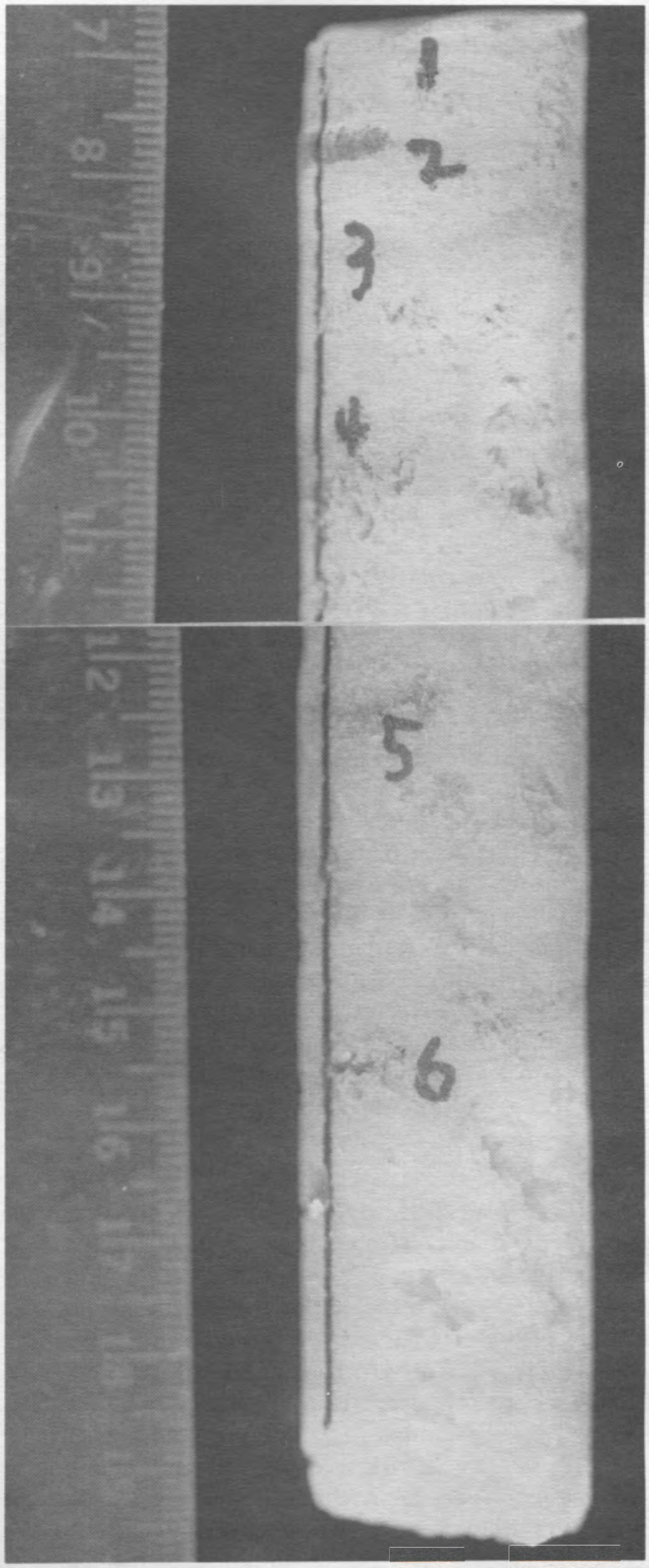

Figure 3. **Sample core from the flowstone floor.**

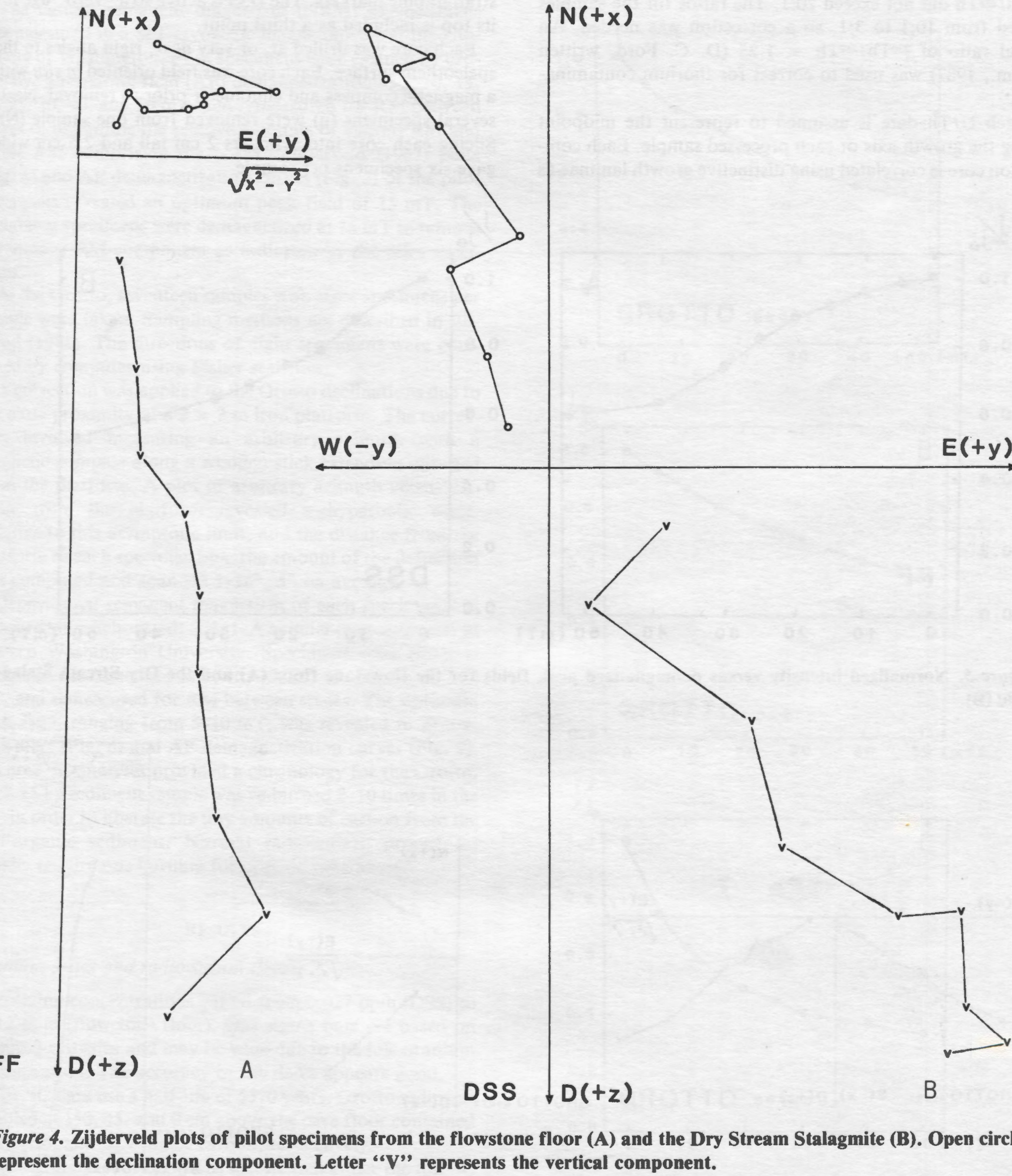

Figure 4. **Zijderveld plots of pilot specimens from the flowstone floor (A) and the Dry Stream Stalagmite (B). Open circles represent the declination component. Letter "V" represents the vertical component.**

$^{230}Th/^{232}Th$ did not exceed 20:1. The ratios for the samples varied from 10:1 to 3:1, so a correction was needed. An initial ratio of $^{230}Th/^{232}Th$ = 1.25 (D. C. Ford, written comm., 1987) was used to correct for thorium contamination.

Each U/Th date is assumed to represent the midpoint along the growth axis of each processed sample. Each companion core is correlated using distinctive growth laminae as stratigraphic markers. The DSS is active so a "zero" age for its top is included as a third point.

Each core was drilled at, or very near, right angles to the speleothem surface. Each core was field oriented *in situ* with a magnetic compass and clinometer prior to removal. Next, several specimens (n) were removed from one sample (N). Slicing each core into cylinders 2 cm tall and 2.5 cm wide gave six specimens per sample.

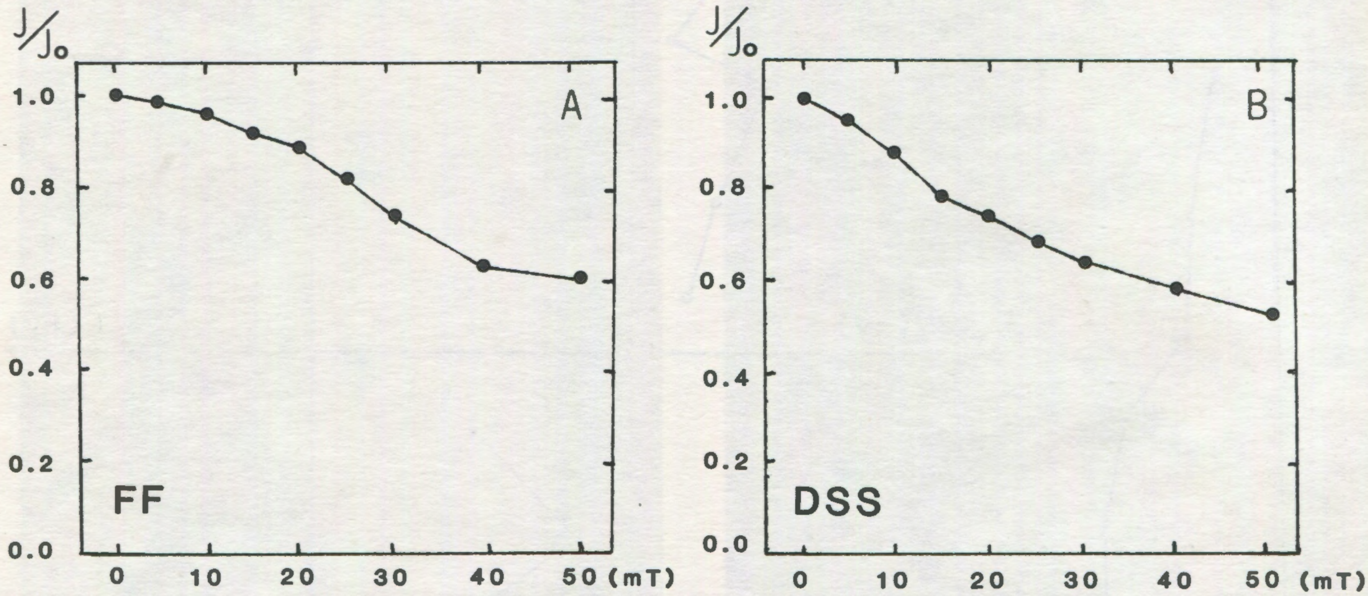

Figure 5. **Normalized intensity versus demagnetized peak fields for the flowstone floor (A) and the Dry Stream Stalagmite (B).**

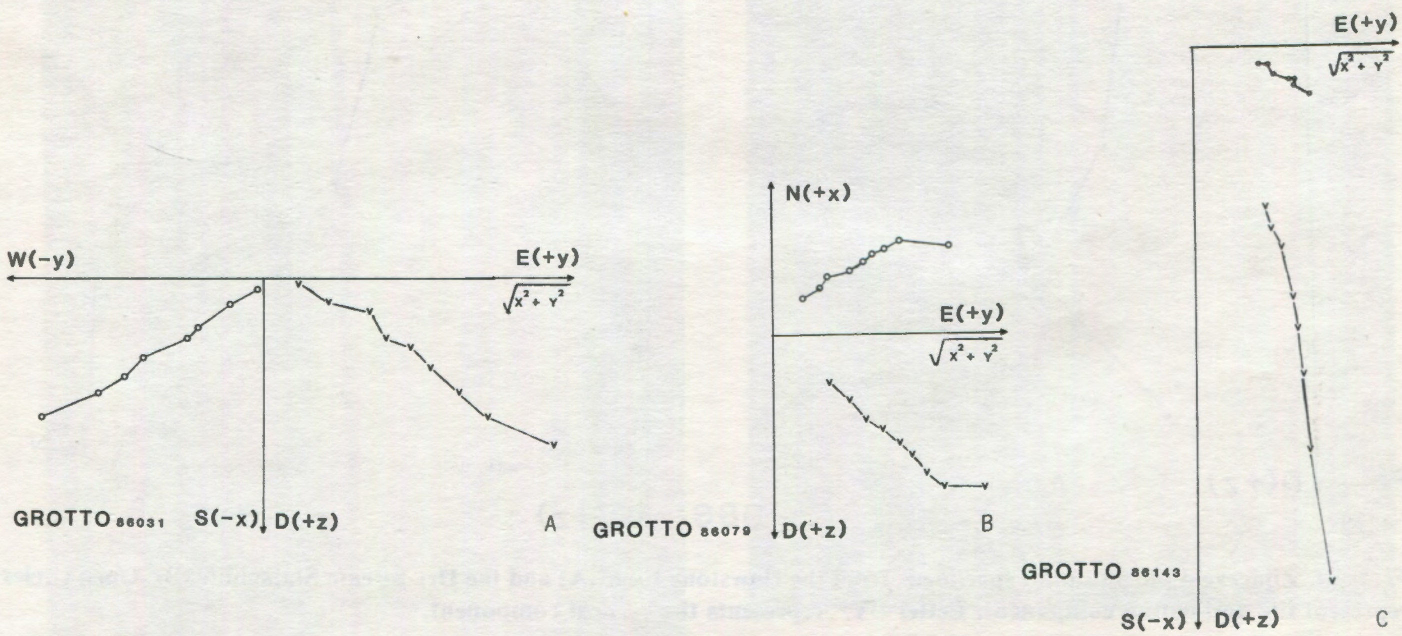

Figure 6. **Zijderveld plots of pilot specimens from the Grotto. Specimens range from youngest (A) to oldest (C). Open circles represent the declination component. Letter "V" represents the vertical component.**

The natural remanent magnetism (NRM) of specimens was measured in a ScT cryogenic magnetometer at the University of Washington. Pilot specimens were cleaned by progressive alternating field (AF) demagnetization at 5, 10, 15, 20, 25, 30, 40, and 50 mT, and remeasured for RM between stages to determine the relative amounts of viscous RM and stable RM comprising the NRM. Zijderveld (1967) plots (Fig. 4) and AF demagnetization curves (Fig. 5) of the pilot specimens revealed an optimum peak field of 15 mT. The remaining specimens were demagnetized at 15 mT to remove the viscous RM component as indicated by the pilot specimens.

At the Grotto, seventeen samples with eight specimens per sample were taken. Sampling methods are described in Ellwood (1971). The directions of eight specimens were combined by computer using Fisher statistics.

A correction was applied to the Grotto declinations due to the close proximity of a 2 × 2 m iron platform. The correction involved measuring an arbitrary azimuth with a magnetic compass along a wooden stick extending outward from the platform. A plot of arbitrary azimuth versus distance from the platform revealed a hyperbolic trace. Relative to this asymptotic limit, and the distance from the platform to each specimen box, the amount of the deflection was computed and spanned 1–18°, 5° on average.

Depositional remanent magnetism of each specimen was measured in a Schonstedt SSM1-A spinner magnetometer at Eastern Washington University. Specimens were progressively AF demagnetized at 5, 10, 15, 20, 25, 30, 40, and 50 mT, and remeasured for RM between stages. The optimum peak field, ranging from 5–10 mT, was revealed in Zijderveld plots (Fig. 6) and AF demagnetization curves (Fig. 7).

Three ^{14}C analyses provided a chronology for the Grotto. Each 15 kg sediment sample was re-burned 8–10 times in the lab, in order to liberate the tiny amounts of carbon from the non-organic sediment. Normal radiocarbon procedures usually require one burning for organic substances.

RESULTS

Uranium-series and radiocarbon dating

Uranium concentrations varied from 0.027 ppm (DSS) to 0.112 ppm (flowstone floor). One-sigma bars are based on counting statistics and may be wide due to the low uranium concentration. The accuracy of the dates appears good.

The ^{14}C data use a half-life of 5570 years. Grotto sediment sampled at 150, 45, and 0 cm above the cave floor contained 25.2, 29.0, and 24.7 mg of carbon, respectively.

Stratigraphic considerations dictate that either the top ^{14}C date is too old or the middle date is too young. The middle date was discarded because samples for the middle data were taken near small holes and were probably bioturbated. A

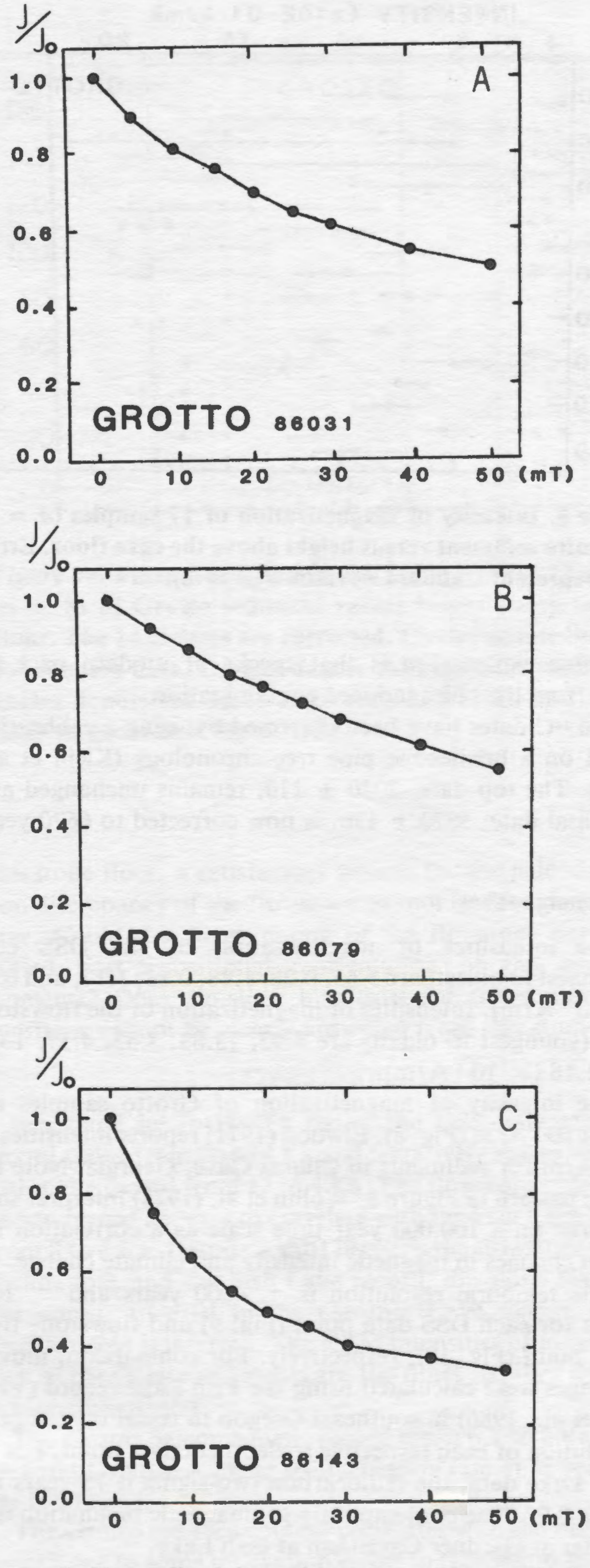

Figure 7. **Normalized intensity versus demagnetized peak fields of pilot specimens from the Grotto. Specimens range from youngest (A) to oldest (C).**

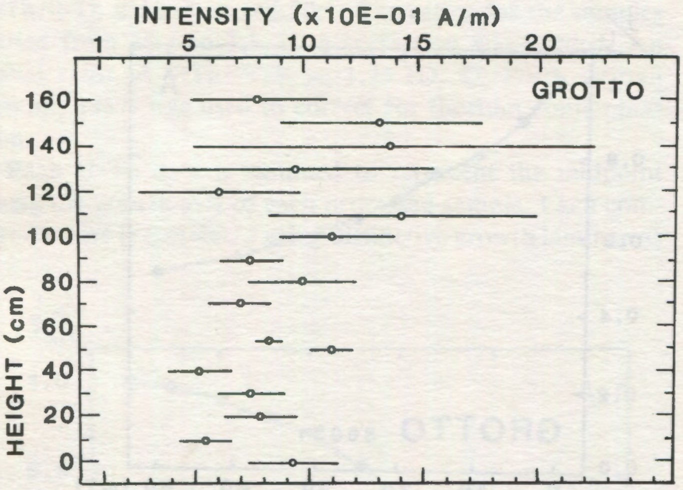

Figure 8. **Intensity of magnetization of 17 samples ($n = 8$) of Grotto sediment versus height above the cave floor. Error bars represent standard deviation ($n = 8$).**

plausible explanation is that specks of modern pack-rat feces from the holes induced contamination.

The [14]C dates have been corrected by using a calibration based on a bristlecone pine tree chronology (Klein et al., 1982). The top date, 2020 ± 110, remains unchanged and the basal date, 5870 ± 130, is now corrected to 6670 years B.P.

Paleomagnetism

The intensities of magnetization of the DSS core (youngest to oldest) are 5.65, 0.28, 1.48, 0.83, 1.17, and 0.61 ($\times 10^{-3}$ A/m). Intensities of magnetization of the flowstone core (youngest to oldest) are 4.93, 13.83, 3.33, 4.57, 1.05, and 1.18 ($\times 10^{-3}$ A/m).

The intensity of magnetization of Grotto samples are about 10^{-1} A/m (Fig. 8). Elwood (1971) reports intensities of 10^{-3} A/m for sediments in Climax Cave, Georgia. Note the cyclic pattern in Figure 8. Wollin et al. (1978) interpret such patterns on a 100,000 year time scale as a correlation between changes in magnetic intensity and climate change.

The temporal resolution is ~ 4000 years and ~ 1000 years for each DSS data point (Fig. 9) and flowstone floor data point (Fig. 10), respectively. For comparison, moving averages were calculated using the Fish Lake record (Verosub et al., 1986) in southeast Oregon to equal the temporal resolution of each respective speleothem data point. For the Fish Lake data, the radiocarbon two-sigma is 75 years and A_{95} is 0.9°. The contemporary geomagnetic inclination is 5° greater at Gardner Cave than at Fish Lake.

Fish Lake and Mara Lake (Turner, 1987) data are plotted with the Grotto data for comparison (Fig. 11). Moving averages calculated for both records equal the temporal

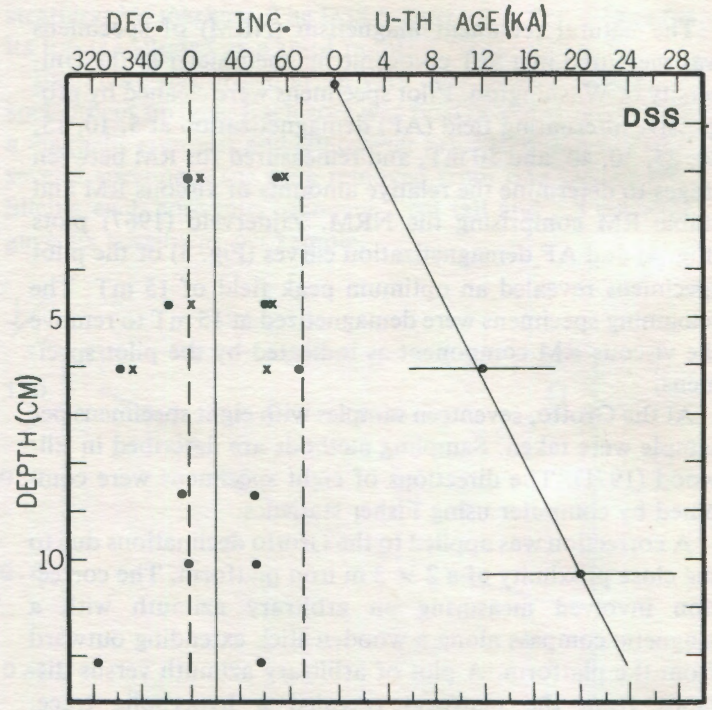

Figure 9. **Paleo—declination and inclination of the core from the Dry Stream stalagmite versus depth of the core. Circles denote the Dry Stream stalagmite data. Crosses denote the data from Fish Lake, Oregon. Dashed line denotes modern geocentric dipole field.**

resolution of each Grotto sample. Directional error bars indicate A_{95}. Grotto specimens at heights of 0, 50, 55, 70, and 80 cm contained high percentages of clay which may induce inclination errors. The poor correlation may be caused by discordant [14]C dates. A good correlation results if the Grotto record is shifted back in time relative to the Fish Lake data. Note the flowstone and Grotto records show "parallel" declination changes to the west from 5 to 2 Ka.

DISCUSSION

Paleomagnetic directions from the DSS with an U/Th time scale agree well with the Fish Lake paleomagnetic record which has a high resolution [14]C time scale (Fig. 9). This fact helps support the validity of the stalagmite record for the past 12,000 years. Furthermore, the DSS record is similar to the Vancouver Island stalagmite paleomagnetic record (Latham et al., 1987), which spans 18.5 to 15 Ka.

The flowstone record shows a consistent eastward shift in declination relative to the Fish Lake record (Fig. 10). Is dip-slope bias possible? Latham et al. (1982) investigated dip-slope bias by demonstrating a topographic test for the internal consistency of stalagmite paleomagnetic directions. For

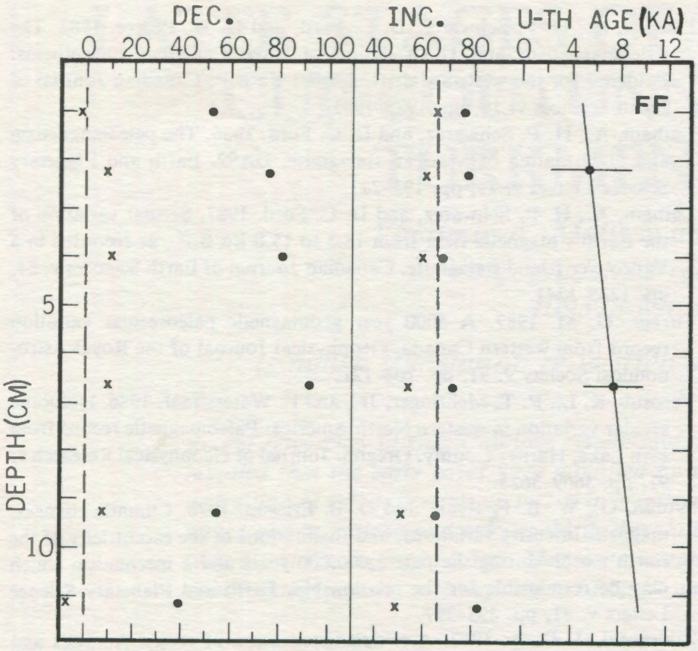

Figure 10. **Paleo—declination and inclination of the core from the flowstone floor versus depth of the core. Circles denote the flowstone floor data. Crosses denote the data from Fish Lake, Oregon. Dashed line denotes modern geocentric dipole field.**

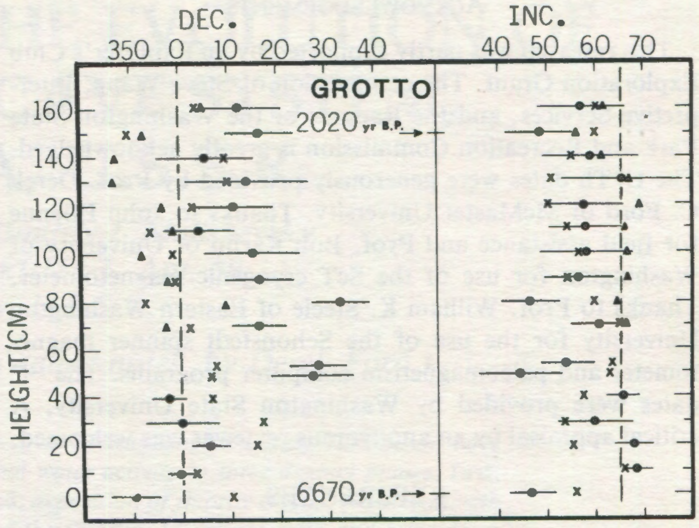

Figure 11. **Paleo—declination and inclination of 17 samples (n = 8) of Grotto sediment versus height above the cave floor. The 14 C dates are corrected. Circles denote the Grotto sediment data. Crosses denote the Fish Lake record. Triangles denote the Mara Lake record. Dashed line denotes modern geocentric dipole field.**

a Mexican stalagmite, Latham et al. (1986) showed excellent concordancy in paleomagnetic directions between central drip cap and lateral specimens. They compared their data with other independent records which strengthened the validity of this internal topographic test for stalagmites.

Two flowstone samples from Crowsnest Pass that Latham et al. (1982) examined exceeded the range of U/Th dating. They lacked a time scale and could not compare their flowstone paleomagnetic data with other dated paleomagnetic records of similar time periods to resolve whether or not paleomagnetic directions are biased by slope on flowstone. Latham et al. (1982) concluded their flowstone samples were probably not biased by dip-slope, as the AF-cleaned paleomagnetic directions differed from the orientation of the flowstone and were near the geocentric dipole field.

The drilling site on the Dry Stream Stalagmite dipped 33° at an azimuth of 159°. The sample area of the flowstone dipped 49° in a direction 213°. These directions do not appear to have biased the paleomagnetic record. Figures 9 and 10 show the Gardner Cave data to be near the geocentric dipole field, except for the paleodeclination record of the flowstone. Hence, the Gardner Cave record does not appear to be dip-slope biased and a chemical RM is suggested.

Until more samples are taken at various slopes along the flowstone floor, a satisfactory answer for the paleodeclination discrepancy of the flowstone cannot be provided at this time. Could the vuggy nature of the flowstone have permitted post-depositional reorientation of the magnetic moments? Does dip-slope bias exist for flowstone? Such questions cannot be clearly answered from this study.

CONCLUSIONS

Bracketed by $^{234}U/^{230}Th$ dates, the DSS paleomagnetic record spans from 20 to 4 Ka and tentatively agrees with the Fish Lake record from 12 to 4 Ka. The flowstone paleoinclination record appears to correlate with the Fish Lake record from 8 to 4 Ka. Despite a shift between the flowstone paleodeclinations and the Fish Lake record, dip-slope bias does not appear to exist in the Gardner Cave record, and a chemical RM is inferred. The NRM intensities of magnetization of the Grotto sediments are 100 times stronger than the calcite speleothems.

The Grotto paleomagnetic record spans 6.7 to 2 Ka based on two ^{14}C dates. A linear sedimentation rate of 31 years per cm is inferred.

The Grotto sediment poorly correlates with the Fish Lake record due to some inclination errors in the Grotto sediment and a possible disconcordance with the ^{14}C dates between sites. A better fit results if the Grotto record is shifted earlier in time relative to the Fish Lake record.

ACKNOWLEDGEMENTS

This research was partly supported by an Explorer's Club Exploration Grant. The cooperation of Steve Wang, Interpretive Services, and the Rangers of the Washington State Park and Recreation Commission is greatly acknowledged. The U/Th dates were generously provided by Prof. Derek C. Ford of McMaster University. Thanks to John Etienne for field assistance and Prof. Bob Karlin of University of Washington for use of the ScT cryogenic magnetometer. Thanks to Prof. William K. Steele of Eastern Washington University for the use of the Schonstedt spinner magnetometer and paleomagnetism computer programs. The ^{14}C dates were provided by Washington State University. A critical appraisal by an anonymous reviewer was welcomed.

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ON HYDROTHERMAL PHASES DURING LATER STAGES OF THE EVOLUTION OF CUP COUTUNN CAVE SYSTEM, TURKMENIA, U.S.S.R.

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(English translation re-written with added notes, by Derek Ford.)

Recent researches in Cup Coutunn system and other caves of Kugitangtau Ridge, Turkmenia, have revealed that the early caves were modified by thermal water activity in three distinct phases: first, *deposition of calcite crystals up to 2 m in length;* second, *deposition of fluorite crystals (≲ 10 cm) with minor quartz, galena and calcite;* third, *dissolution of the fluorite. Limestone walls and breakdown were altered to depths of 0.5 m. Thermal minerals have since been attacked and reworked by H_2S-rich condensation waters, with deposition of barite and celestite.*

Recent mineralogical researches in the Cup Coutunn System and other caves of the Kugitangtau Ridge, Turkmenia (Fig. 1), have yielded an unexpected result—the caves have been affected by thermal waters. This was not known before because (1) thermal water erosional effects upon the bedrock morphology are very minor and any thermal minerals have been destroyed where later meteoric waters flowed; thus, evidences of thermal activity are confined to a few places in upper levels, (2) the caves are beautifully decorated with conventional speleothems so that the speleologists readily overlook the less prominent thermal deposits. For example, in 1981 the first author took the leading Soviet cave mineralogist, Victor Stepanov, to the Promeszutochnaya Cave section of the system; both walked past a display of large fluorite crystals in the floor without even noticing them. (However, in the same expedition Stepanov did discover evidences of thermal activity in a different system in these mountains, Fata Morgana Cave, near Gaurdak; see Fig. 1.)

The Promeszutochnaya fluorite crystals (CaF_2) were the first thermal deposits to be reported, being noted by the Gorky caving group in 1984 (Moroshkin, 1984; Maltsev, 1987). The crystals were displaced from their growth sites and their genetic relationships could not be established. There were several similar finds by different groups during 1985-88 but no special studies were undertaken.

This note presents preliminary results from our research in 1988 and 1989 in Promeszutochnaya. Many of the observations are supported by findings of other groups in other parts of the system, chiefly Cup Coutunn Cave and Tush-Jyruck Cave. The results may be of great significance in the understanding of speleogenesis in this particular region, and perhaps in many other karst areas. The scientific resources of our caving groups are not sufficient to solve many of the problems; other cave scientists are invited to collaborate if they would like to.

Following creation of the caves themselves, thermal water activity in them appears to have occurred in three successive phases, followed by a distinctive post-thermal phase of reworking:—

Phase 1; crusts of gigantic calcite crystals were deposited. Crystals may be up to 2 m in length. They contain inclusions with sulfides (metacinnabar, HgS, has been recognized) and manganese oxides (Fig. 2). The calculated temperatures of deposition are 100–150° C.

Phase 2; deposition of fluorite crystals up to 10 cm in length, with galena (≲ 0.5 mm), quartz (≲ 1.0 mm) and tiny calcite crystals (Fig. 3). Fluorite depositional temperatures are estimated to be 70–100° C. Interesting features of the fluorite include high concentrations of strontium (up to 4%), high internal stress in the crystals, the absence of any luminescence, or of any gas phase in the fluid inclusions.

Phase 3; was marked by strong dissolution of the fluorite; it penetrates to depths of 4 cm along crystal boundaries.

During all three phases apparently, there was alteration of the limestone walls. Alteration zones are up to 0.5 m in depth. We have studied several rock falls that occurred before the thermal water deposition and found the blocks to be altered to the same depth on all of their faces. Yet the

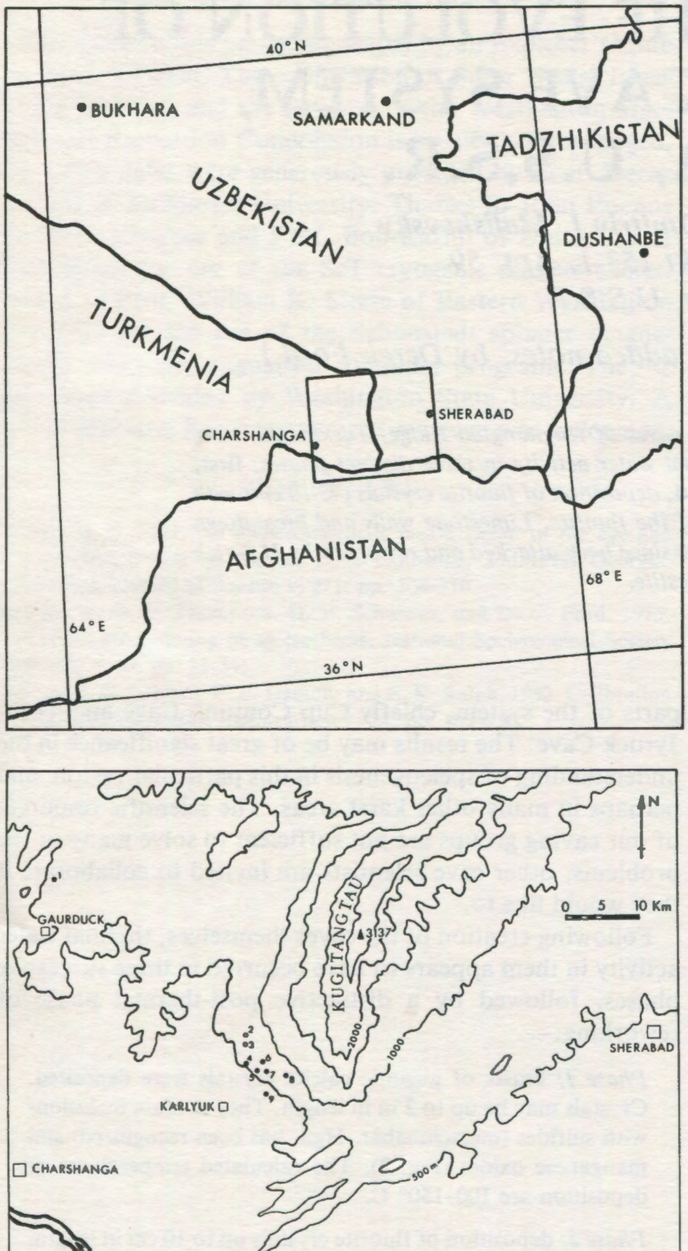

Figure 1. **A) Location of the Kugitangtau cave region, Turkmenia, U.S.S.R. B) 1—Fata Morgana Cave. 2—Verticalnaya Cave. 3—Geophyzicheskaya Cave. 4—Hushm-Oyeek Cave. 5—Promeszutochnaya Cave. 6—Tush-Jyruck Cave. 7—Cup Coutunn Main Cave. 8—Bezvimyannaya Cave. 9—Provull Cave.**

Figure 2. **Remnants of large calcite crystals covered by coatings of sulfides. The crystal exposed at right center is 45 cm in length, now reduced to 1 cm in thickness.**

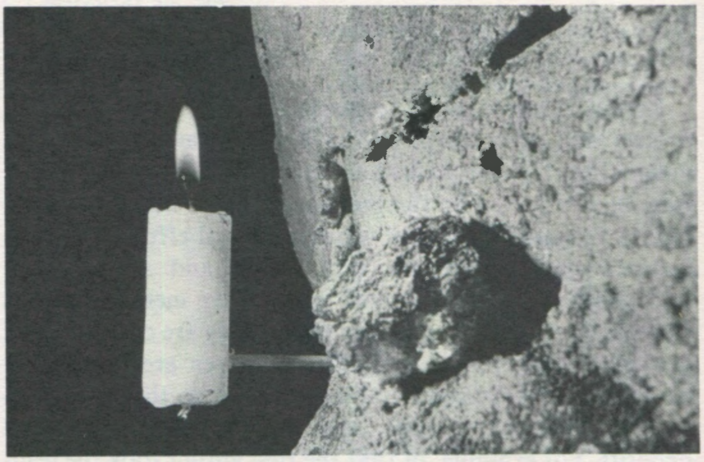

Figure 3. **A large crystal of fluorite (2nd thermal phase), with a surface that has suffered strong dissolution (3rd thermal phase).**

morphology of these rock falls cannot be distinguished from that of falls occurring after the thermal phases. This is the basis for our contention that the thermal waters cannot have contributed significantly to the erosional sculpturing of the caves (above).

The thermal mineral phases have not been dated but it may be supposed that they are younger than the middle Quaternary because the principal phases of enlargement of the caves themselves are believed to be of that age (Kucheriavyh and Abduszabarov, 1982). We have not been able to find any source tectonic structures for the thermal activity, such as faults displaying the same sequence of mineralization. Such young thermal activity has not been recognized by students of the geology of the Kugitangtau Mountains. It appears to us that the entire period of thermal activity in the caves was quite short, but it was also vigorous.

Effects that followed the thermal phase are most interesting. As noted, thermal deposits are preserved only where there could be no significant flow of meteoric water subsequently. In such areas there is corrosion by condensation waters from the atmosphere only. Such corrosion has weathered the limestone walls to depths up to 10 cm. The calcite is partially gone and fluffy residual clays remain that preserve the limestone texture. The clays range from yellow to red or black in color, and contains such ore-associated minerals as galena, metacinnabar and manganese oxide. Their concentration in the weathered zones is 15–20% by weight/volume, compared to only 2–6% in the parent limestone. Hydrohematite (fine grained hematite, Fe_2O_3) and hydromuscovite (or illite, $KAl_3Si_3O_{10}[OH_2]$) have been reported in similar residual clays in Fata Morgana Cave (Lazarev and Philenko, 1976).

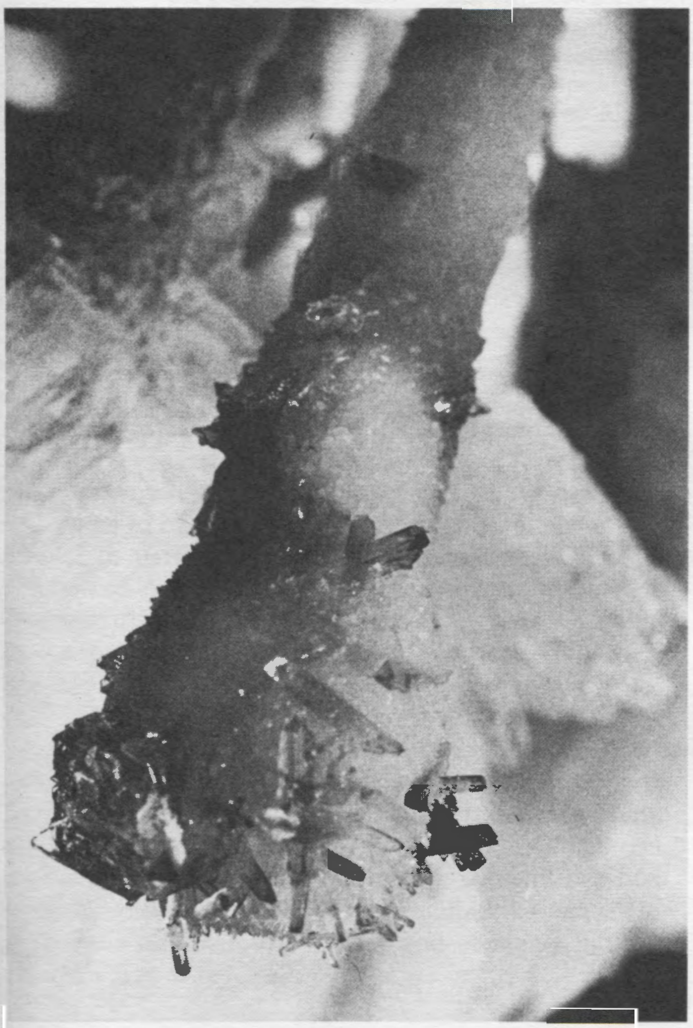

Figure 4. **Reworking and redeposition of thermal minerals in post-thermal phases—celestite crystals grow from a normal calcite helictite. Larger crystals are 1.5 cm in length.**

The gigantic calcite crystals of the first thermal water phase have also been attacked by condensation waters and almost dissolved away, many being reduced to relicts 0.5 to 2.0 cm thick. A distinctive feature is that very thin (1–5 μm) overgrowths of sulfides that formed during the second thermal phase are not dissolved; they are now very fragile sheaths enclosing the surviving crystals. From spectral analysis, the sheaths consist chiefly of metacinnabar. There may also be some metallic mercury but this cannot be proven because the crystals are too small for X-ray analysis. There are similar, but thicker, sheaths in Fata Morgana Cave that have not been linked to any thermal water activity as yet. Lazarev and Philenko (1976) suggested that these may have been deposited at the same time as gypsum crusts formed upon other calcite surfaces, but we believe that such gypsum developed when H_2S-rich condensation waters oxidized and reacted with the calcite.

In some areas of Promeszutochnaya seeping groundwaters (i.e., the type that normally deposit calcite dripstones, etc.) flow over the thermal precipitates and rework them. Individual crystal rosettes of celestite ($SrSO_4$) and barite ($BaSO_4$) are found; they are up to 3 cm in diameter (Fig. 4). There is also sokonite, a zinc-rich clay mineral of the montmorillonite group that is colored green by the presence of ~ 2% Ni, and other minerals not yet identified. Their scale and location indicates that all are derived from alteration of the earlier thermal minerals.

There are evidences of other possible thermal water effects in areas of the Cup Coutunn System that are intersected by a series of upthrown faults termed the Chilghaz Zone. This zone is considered to be a source structure for Pb-Zn mineralization elsewhere in the mountains. In Cup Coutunn, young flowstones in it contain up to 1% of Pb and Zn. There are also large deposits of aragonite (which is rare in all other regions of the caves) and some deposits of cerussite ($PbCO_3$). However, there are no morphological or other features of direct thermal water activity.

In conclusion we suggest that:

(1) There was hydrothermal activity in the Cup Coutunn System. However, it cannot be classified as a hydrothermal system in its origin because the thermal phases appear to have been quite brief and to have done little work to enlarge already existing caves.

(2) Thermal contributions to, and effects upon, the secondary mineral suites in the caves, however, are quite considerable. Their study is only just beginning.

(3) The residual clays in the altered wall rocks may be particularly significant because they should concentrate the products of the thermal activity. We have yet to study them.

(4) These thermal phases were not known in the Cup Coutunn area before. They are probably similar to those recognized previously in the Fata Morgana area.

[*Editor's Note.* The sequence of events proposed above is basically similar to that suggested by Dr. D. E. Deal for Jewel Cave, South Dakota, i.e., a phreatic cave system excavated by earlier, presumably cool meteoric, waters is invaded by hot waters that do little erosional work but alter the wall rock and precipitate characteristic mineral deposits. Following the thermal phase the caves drain and are not, thereafter, re-occupied by cool waters. M. Bakalowicz and I have questioned such sequences on the grounds that evidence of early, *erosional* thermal water phases will not be preserved, only the late, depositional or alteration phases. See Deal, 1961, M.S. thesis, University of Wyoming; Bakalowicz et al., 1987, Bulletin, Geol. Soc. Am., pp. 729-738]

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THE INFLUENCE OF SEASONAL CHANGES OF CAVE MICROCLIMATE UPON THE GENESIS OF GYPSUM FORMATIONS IN CAVES

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(English translation re-written with extensions, by Derek Ford.)

Many anemolites (wind-controlled speleothems) appear to be examples of short period deposition ("ephemerae"), particularly associated with seasonal reversals of airflow. Gypsum frostwork in Jurinskaya gypsum cave, Podolia, Ukraine, grew when external air flowed in during the summer and relative humidity was 70–90%: it was partly dissolved when cave air flowed out in the winter, with relative humidity rising very close to 100%. Maximum accumulation in one season was 7 mm. Frostwork only occurs close to present or past cave entrances and thus is a guide to exploration.

Gypsum frostwork occurs in the limestone maze caves of Kugitangtau Ridge, Turkmenia. Twenty-six studied sites all lay within 400 m of an entrance, where relative humidities could be seasonally depressed to 60–70% or below. Anemolites of calcite in the same caves also form in the deeper interiors where relative humidity does not fall below 90–95%.

Most authors treat the crystallization of speleothems as a continuous process or, at the least, as one that operates over lengthy periods of time. There are only a few discussions of short cycle deposition, and they are concerned primarily with the age of the deposits (e.g., Shopov and Rusanov, 1985). Many anemolites (wind-controlled speleothems) must be examples of short-period deposition. They are well known and quite abundant in many regions, yet there does not appear to be a well formulated physical model of their genesis. In particular, there are seasonal changes in the direction of airflow in most draughty caves so that we cannot make the simple physical assertion that "anemolites grow into the wind." The purpose of this paper is to investigate this and some other problems of short-period deposition.

Seasonal change (reversal) in the direction of airflow is perhaps the most common short-period, cyclic process encountered in caves. Reversal itself will not initiate or terminate speleothem crystallization but, rather, the change of relative humidity that may occur as a consequence of it. Normally, there are seasons of lower humidity when airflow is from the surface into the cave, and of higher humidity when this flow is reversed.

The Gypsum Caves of Podolia

To investigate these effects I have studied the celebrated gypsum maze caves of Podolia, Ukraine (e.g., Lomaev, 1978; Klimchouk, 1986). They are suitable because they possess large but compact volumes that "breathe" through few entrances and because the deposition of gypsum anemolites can be very rapid and will be controlled only by evaporative processes, whereas deposition of calcite is also strongly regulated by CO_2 partial pressure in the cave atmosphere.

To begin with I analyzed observations of "gypsum snow" (Hill and Forti, 1986) reported by members of the Lwow Speleological Society, and examined some deposits in Jurinskaya Cave. The snow has been described as an aerosol product. There were also some strange deposits of "gypsum frostwork" or "rims" that the Lwow cavers had discovered. At particular sites frostwork was present and, apparently, growing in September 1987 (Figure 1) and then was strongly dissolved in May 1988. At other sites it was absent in May 1988 but abundant in October that year. The Lwow cavers attempted to relate its appearance and disappearance to airborne mass transport of gypsum resulting from their manual enlargement of some constricted passages, but could not produce a convincing physical model.

My analyses suggest that the "frostwork" and "snow" in Jurinskaya Cave represent one type of speleothem at different seasonal stages of its evolution: *In the summer,* airflow is into the cave and the relative humidity is 70–90%. Thin films of slightly undersaturated condensation water are evaporated and deposit delicate frostwork on the walls at rates up to 7 mm per season. *In the winter,* the airflow is reversed and the humidity rises to ~ 100%. New films of condensation water attack the frostwork at its base (its junction with walls, etc.), causing it to fall. The accumulations of partly dissolved frostwork constitute the "gypsum snow."

Some specific findings are:

(i) that the maximum growth of frostwork during a single season is 7 mm, as noted.
(ii) that the maximum amount of frostwork dissolution that has occurred in a single season without its falling away as "snow" is 0.5 mm.
(iii) that dissolution continues in the deposits of snow but there is no further precipitation occurring in them. Rates of dissolution were not established.
(iv) that the maximum thickness of the deposits is ~ 100 mm. The increment from "fresh snow" in 1988 was as great as 2 mm.
(v) some deposits are "dead"; i.e., there is no further addition of fallen frostwork. The degree of dissolution apparent on their surfaces is greater than normal.
(vi) the maximum age of deposits is probably around 500 years.

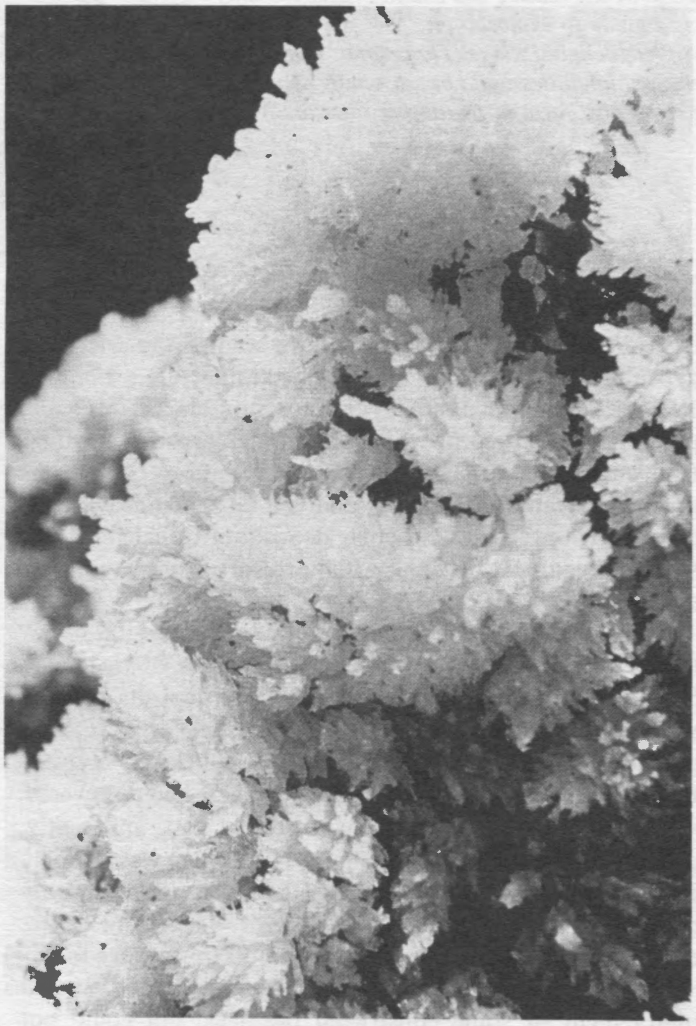

Figure 1. **Gypsum frostwork ("ephemerae") developed during a single season, Jurinskaya Cave, Podolia, U.S.S.R. Greatest crystals are 7 mm in length.**

Thus, there are good examples of crystalline speleothems whose growth and decay are controlled by seasonal cycles of humidity. To distinguish them from other categories of frostwork, etc. in caves the general term "ephemerae" has been suggested for them (Maltsev and Turchinoff, 1989).

[*Editor's note:* Very similar seasonal frostwork and snow may form where hoarfrost is precipitated on to cold cave walls during spring or summer or autumn periods of moist airflow in arctic and alpine caves subject to such seasonal reversals of airflow. Carol Hill notes that there may be seasonal growth and decay of deliquescent salts such as epsomite ($MgSO_4$ $7H_2O$), mirabilite (Na_4SO_4 $10H_2O$) and nitrocalcite ($Ca(NO_3)_3$ $4H_2O$).]

The Limestone Caves of the Kugitangtau Mountains, Turkmenia

For a long time the gypsum ephemerae of the Podolian caves have been used by cavers as practical indicators of the presence of large new volumes of cave. The genetic model proposed here indicates why their presence in, for example, a tight crawl, has such practical significance. The model has been adopted in the search for new extensions in the maze caves of the Kugitangtau, which are the chief exploration focus of my own caving club in Moscow (for a location map, see Maltsev and Malishevsky, in press, Fig. 1).

The Kugitangtau caves are developed in limestone. They have abundant gypsum deposits (from interbeds or other sources—Editor) but their rate of formation appears to be much slower than it is in the wholly gypsum caves of Podolia. As a consequence, the "snowfalls" are not annual events but, rather, suggest some longer climatic cycle; thus, there are no true "ephemerae" present. Instead, some other distinct seasonal effects of the same processes are noted. Chief among them is that recrystallization of the speleothems takes place as they are growing. First generation gypsum forms on the outer surface during the low humidity growth season, to be succeeded by dissolution and recrystallization in the higher humidity season.

Almost all of the large gypsum speleothems in the Kugitangtau display this feature, although the very small ones do not (Maltsev, 1987). Thick wall crusts are disconnected from

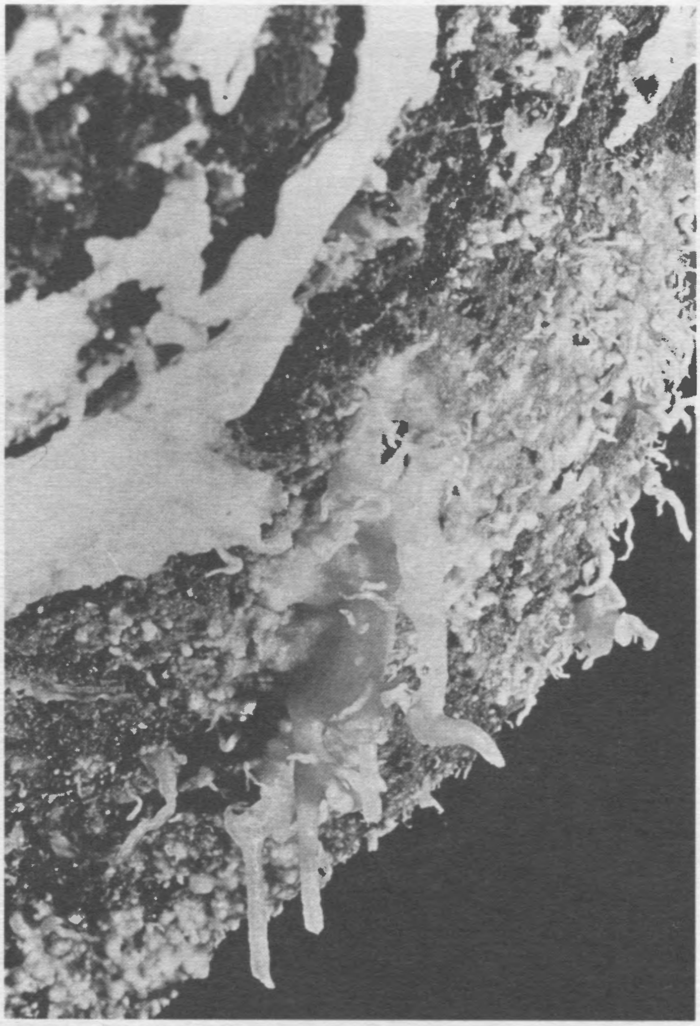

Figure 2. **Gypsum wall crusts spalling from the substrate. They are partially recrystallized, and covered with seasonal frostwork. Cup Coutunn System.**

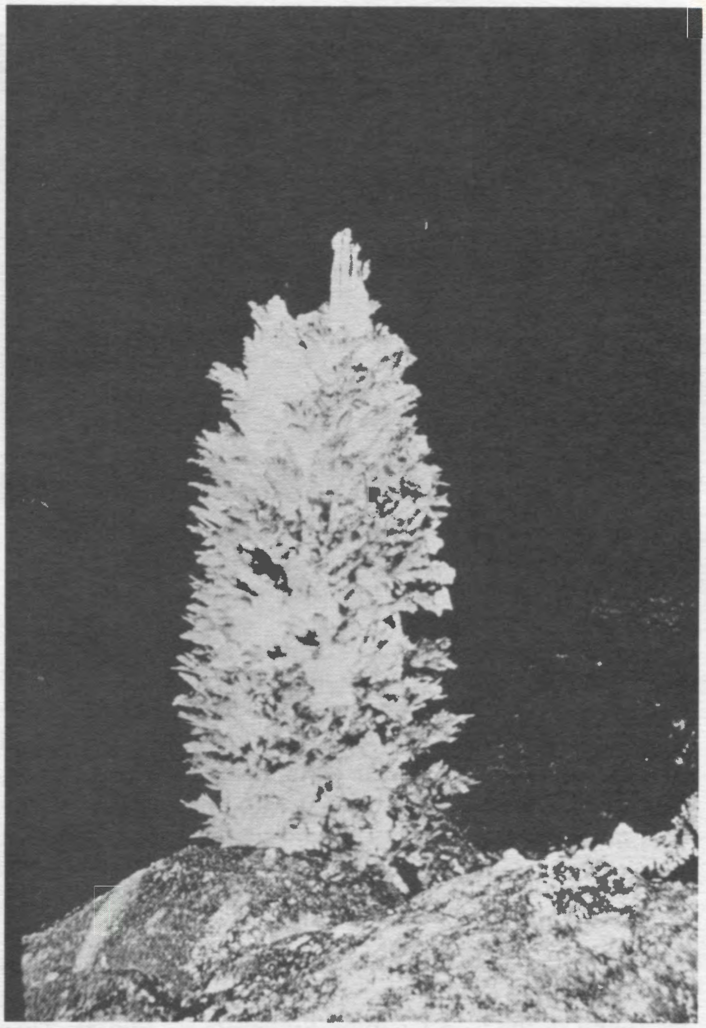

Figure 3. **Hollow stalagmite of gypsum recrystallized into a dendritic form. This example is 2 m in height. Cup Coutunn System.**

the substrate (Fig. 2), are strongly recrystallized and covered by frostwork. Gypsum flowers growing outward from these crusts are found to extend through them, but have no "roots." Stalagmites present (Fig. 3) are hollow and may be up to 3 m in diameter. Their walls are 1–30 cm in thickness and consist of recrystallized frostwork.

It appears, therefore, that the larger gypsum forms are strongly affected by seasonal changes of relative humidity. To test this hypothesis, series of microclimate measurements were undertaken in four different cave entrance areas, with ten or more measuring stations in each area. It was determined that when the airflow was outward from the deep interior of the caves, the relative humdity was ~ 100% to each entrance. When airflow was in from the outside, relative humidity fell to seasonal minima of 60–70% at stations 500 m inside the caves, and to 90–95% at 800 m. Areas of the caves with displays of the large gypsum deposits are always quite localized and, in the 26 cases studied, lay within 400 m of a past or present entrance.

In the Kugitangtau, therefore, I suggest that gypsum speleothems are controlled by seasonal processes. Gypsum is too soluble to permit long-lived speleothems to survive unless they are supported by on-going recrystallization. The recrystallization is forced by seasonal changes of the relative humidity at a site. Without such changes large speleothems will "die," i.e., dissolve and fall.

These observations help our search for new extensions in the Kugitangtau caves: (i) presence of large gypsum formations in remote galleries indicates that there is a past or present entrance within a few hundred metres, and (ii) sites of strong past airflows will be marked by deposits of thin, fallen crusts of gypsum displaying crystals up to about 3 mm

Figure 4. **"Normal" anemolites (wind flags or wind-oriented helictites), Cup Coutunn System.**

in length. These criteria led to important new discoveries in the Promeszutochnaya Cave section of the Cup Coutunn System (50 km) in 1988 and 1989. [*Editor's note:* Carol Hill reports that there are similar features in Lechuguilla Cave, New Mexico. They are associated with areas where the airflow rates are high, but not necessarily near entrances.]

Wind-Controlled Calcium Carbonate Speleothems in the Kugitangtau Caves

Both of the classical forms of anemolites (helictites and ellipsoidal stalactites or stalagmites—Hill and Forti, 1986) are found in the Cup Coutunn System (e.g., Fig. 4). They are also confined to areas where there are significant seasonal reversals of the direction of airflow, but in this case are deeper inside the caves so that any related changes of relative humidity are minor, being beyond the resolution of our measuring instruments. The carbon dioxide balance in the air is probably controlled by seasonal changes of the wind also, but that is a subject for future studies.

One particular feature is that many calcite speleothems (stalactites, stalagmites or helictites) display a strong color and textural asymmetry that is also wind-controlled. One side is white and opaque, the other yellow and translucent. In a given area all asymmetric speleothems display white, opaque or yellow translucent surfaces with the same orientation (Fig. 5). Study shows that the opaque surfaces are suffering net re-solution while the translucent surfaces are growing, i.e., the dissolved material from the opaque sides is being reprecipitated on to them. [*Editor's note:* Similar phenomena are noted in Carlsbad Cavern.]

Our micro-climate studies revealed that this kind of speleothem is confined to regions where the relative humidity falls to seasonal minima of 90–95% but not lower, i.e., relatively far inside the caves. The opaque (net re-solution) surfaces are oriented towards the further interior (i.e., are impacted by seasonally outward-flowing air) and the translucent surfaces are oriented towards the nearest entrance. These characteristics are also valuable aids in our search for new sections in these great caves.

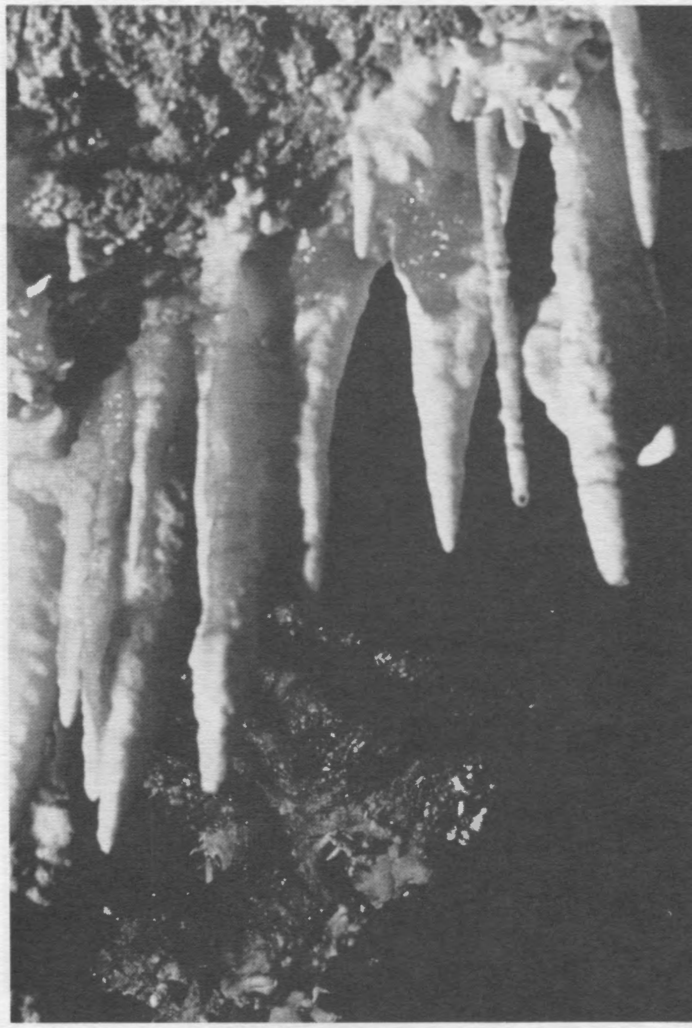

Figure 5. **Bi-color and textural assymetry of calcite stalactites, Cup Coutunn. Dull opaque surfaces suffer net re-solution; the dissolved calcite is redeposited as yellow translucent crystal on protected, downwind sides.**

A general conclusion is that the role of short-period (seasonal) effects of relative humidity in cave mineralogy is greater than has been supposed hitherto. This is especially true for the more soluble minerals such as gypsum. Knowledge of these effects assist in our exploration for new sections in extensive caves.

ACKNOWLEDGEMENT

The author thanks the members of the Lvov Caving Club, especially Igor Turchinoff, for their observations and for their guide to Jurinskaya Cave; the members of his own caving group—Vladimir Detinich, Dmitryi Malishevsky and others for their help in the research; Nina Skorobogatova, Oleg Bartenev and many others for their professional advice.

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FACTORS THAT MAY AFFECT RADON DAUGHTER CONCENTRATIONS IN WHIPPLE CAVE AND GOSHUTE CAVE, NEVADA

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Radon daughter measurements taken in Whipple Cave and Goshute Cave in Nevada in the summer of 1984 showed low levels of radioactivity. Working levels ranged from .048 to .073 in Whipple Cave and from .056 to .128 in Goshute Cave. Factors other than meteorological conditions that may contribute to radon concentrations are suggested. The higher average radon concentrations in Goshute Cave may relate to its orientation away from prevailing winds, its small entrance, and its complexity of passageways, all of which may tend to hinder free airflow and reduce ventilation. Airflow within Whipple Cave may be aided by a large entrance oriented toward prevailing winds and a large, straight passageway; both may serve to dilute radon in that cave.

INTRODUCTION

Presence of the naturally occurring radioactive gas radon and its daughters (or decay products) has been considered to be a potential health hazard to cavers since Breisch first publicized its existence in caves in 1968. Radon daughters, which form when radon decays, are a health problem because when inhaled into the lungs, they attach to the airways and subject the tissue to increased levels of radioactive elements (Jacobi, 1972). Investigations in mines, directed toward the health and safety of miners, suggests that such exposures to radon daughters contribute to an increased incidence of lung cancer and deaths. According to OSHA (Occupational Safety and Health Administration) and other federal agencies, radon in small doses is not known to be harmful. However, extended exposure to low concentrations may be as or more hazardous than short-term exposure to high concentrations (Yarborough, 1982). Although much early radon research was conducted in mines, since 1968, various cave scientists have also investigated the factors contributing to radon concentrations in cave systems.

The National Park Service (NPS) has sponsored a majority of the research in order to ensure the safety of rangers who spend considerable time underground conducting tours for visitors. Radon daughter concentrations may be measured in working levels (W.L.). "One W.L. is defined as any combination of short-lived radon and/or thoron daughters in one liter of air which will result in the ultimate emission of 1.2×10^5 MeV [million electron volts] of potential alpha energy" (Yarborough, 1977).

Because radon may also pose a potential threat to the recreational caver, wild caves as well as NPS caves should be monitored if possible for their radon concentrations in order to assess that hazard. The vast number of wild caves and the lack of equipment and/or trained personnel, however, preclude the ideal scenario of extensive data collection and research. An alternative, perhaps, is to discover other parameters that may be used as indicators of radon concentration.

Problem

This study has two objectives, one very simple, the other quite complex. First, what are the radon daughter levels in two wild caves, Whipple Cave and Goshute Cave, where radioactivity has never before been measured? Even a one-time measurement will provide an approximation of the radon concentrations of these two wild caves, both on Bureau of Land Management (BLM) property in Nevada. Such estimates may be useful to BLM personnel and to recreational cavers.

The second purpose of the study is to propose hypotheses concerning the importance of factors other than the meteorological parameters of temperature and barometric differences that may control and alter radon concentrations in these two caves. It is hoped that a tentative link can be made between the radon levels and the physical configurations of Whipple Cave and Goshute Cave, which could provide the basis for further research.

Several studies have indicated a direct relationship between radon concentrations and exterior weather conditions, primarily temperature and/or barometric pressure. In right-side-up caves (RSU), i.e., caves where most of the passageways are below the level of the entrance, there often is a distinct flow of air in or out depending on the season and the ambient surface temperature compared with cave temperature (Yarborough, 1977; 1978). Such an air exchange is a consequence of barometric pressure differences, primarily induced by temperature differences. These natural ventilation patterns are the means by which radon can be dissipated or concentrated (Gascoyne, 1978; Yarborough, 1978). In winter when the surface air is colder than that in the cave, air flows into the cave, displacing the cave air and diluting the radon. Conversely, in summer when the ambient air temperature is warmer than the cave, the colder cave air stagnates and radon accumulates.

Radon concentrations in some caves have shown seasonal variations that positively correlate with the annual march of outdoor temperature. Ahlstrand (1977, 1980) and Ahlstrand and Fry (1979) successfully correlated daily minimum temperatures with seasonal fluctuations in radon levels in Carlsbad Caverns. They found that cave radon was lower in winter months when outside air was colder than cave air and there was, consequently, more air exchange. During the hotter summer months when there was less air exchange, cave radon concentrations were higher. In extensive studies reported in 1979, Yarborough and others consistently concluded that summer increases and winter decreases in radon in several caves throughout the United States were due to differential air flow controlled by temperature-induced seasonal ventilation patterns. Variations from this seasonal pattern of radon concentrations were noted primarily at Mammoth Cave, Crystal Cave of Floyd Collins, and Great Onyx Cave in Kentucky. The National Park Service policy of erecting a steel door at the entrance to Mammoth Cave during the winter interrupts the natural airflow pattern of these interconnected caves and alters the expected radon levels. Quinn (1988) reported a significant relationship between minimum surface temperatures and cave radon levels at Lehman Caves, Nevada. If outside temperatures fell below the temperature of cave air, radon daughter concentrations consistently measured lower than .300 W.L. Conversely, when the minimum surface temperature was above the cave temperature, the radon measured higher than .300 W.L.

Factors which influence radon release and concentration in caves are many and complex. However, easily recognizable factors which may influence the concentration of radon daughters may include the following:

(1) Orientation of the cave opening toward prevailing winds may have the effect of increasing the amount of air entering the cave, which then dilutes the cave radon regardless of the ambient temperature.

(2) Configuration of the cave, in addition to RSU or USD, may play a role. A cave with a complex network of

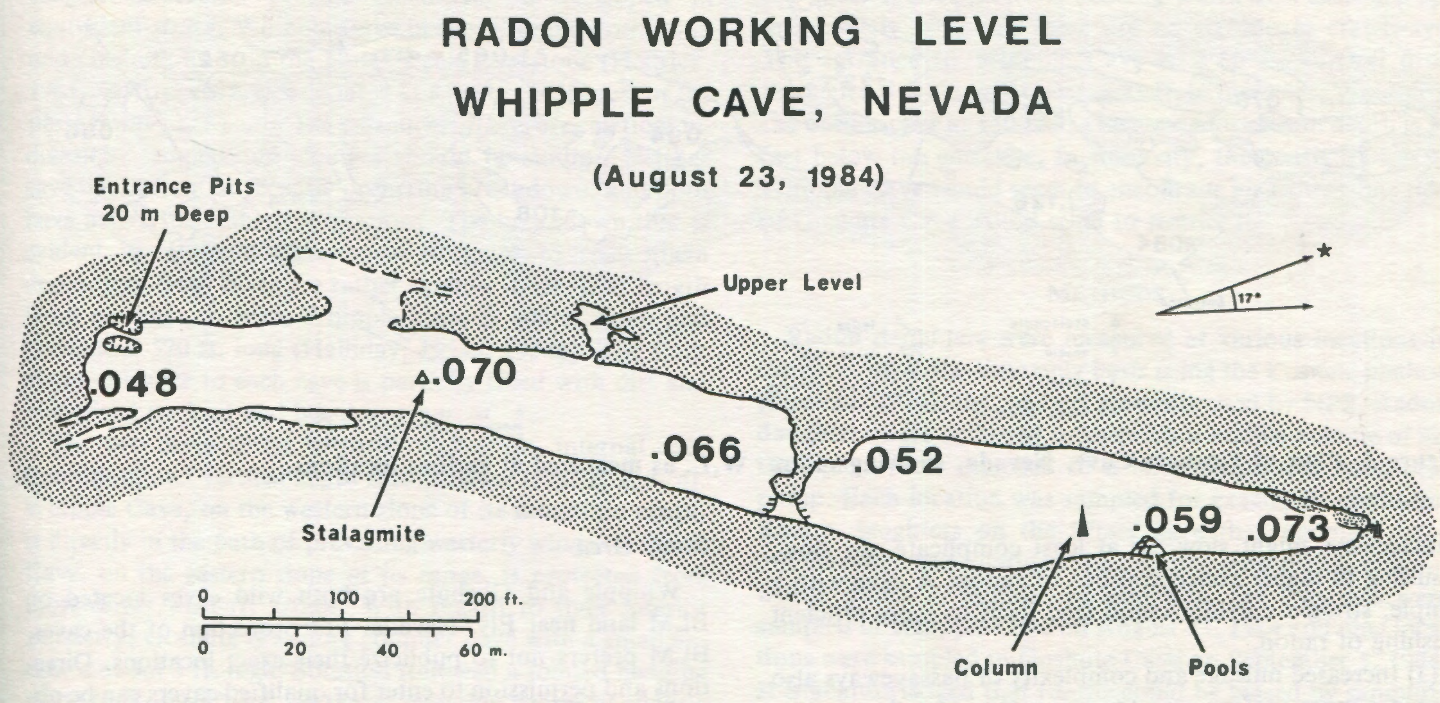

Figure 1. **Plan of Whipple Cave, Nevada, showing radon W.L. as measured at various locations.**

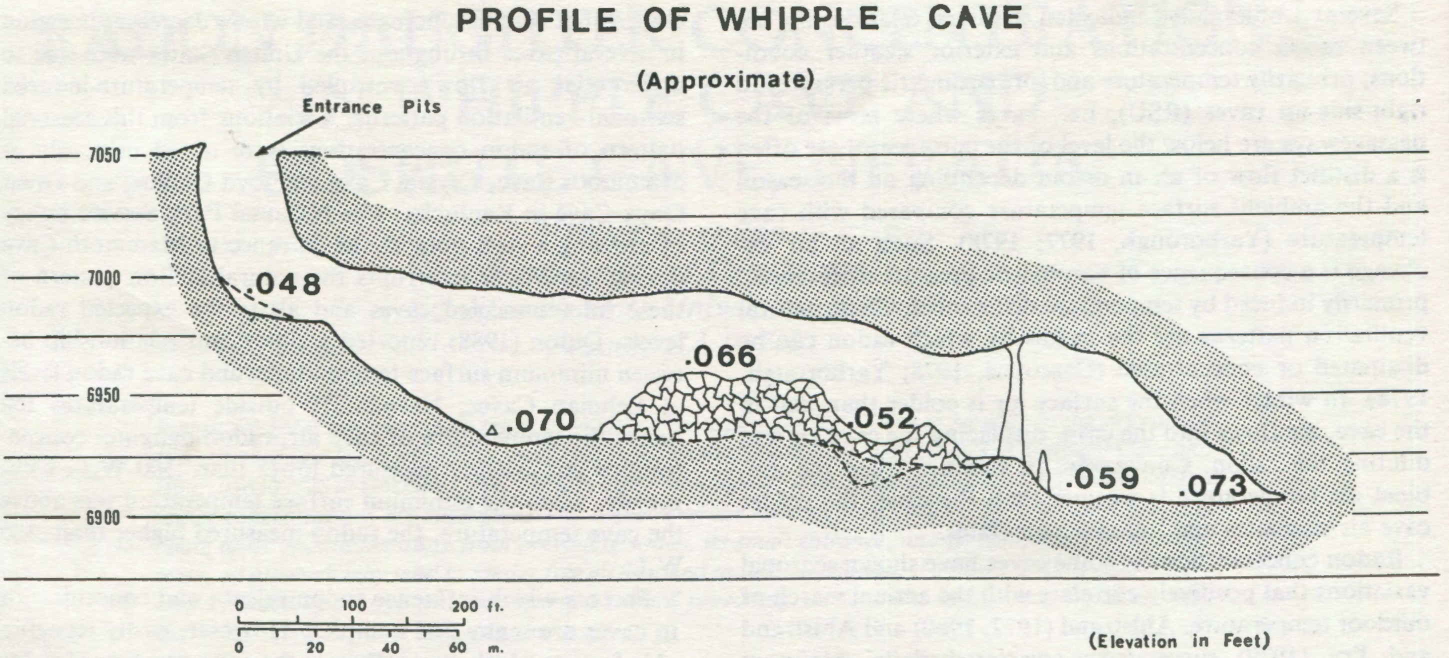

Figure 2. **Approximate cross-section of Whipple Cave, Nevada, showing radon W.L. as measured at various locations and depths.**

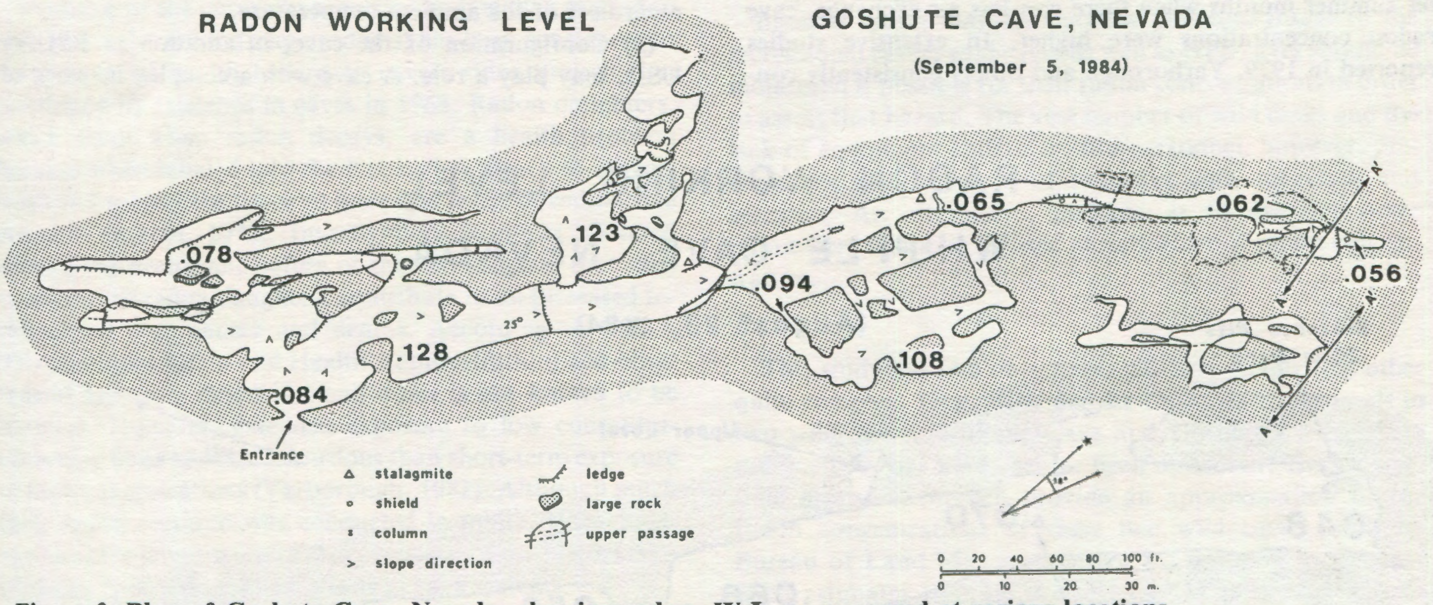

Figure 3. **Plan of Goshute Cave, Nevada, showing radon W.L. as measured at various locations.**

passageways might slow, or at least complicate, air flow, resulting in higher concentrations of radon. A cave with a simple, straight corridor could lend itself to more efficient flushing of radon.

(3) Increased mileage and complexity of passageways also provides more surface area for emanation of radon.

Study Area

Whipple and Goshute are both wild caves located on BLM land near Ely, Nevada. For protection of the caves, BLM prefers not to publicize their exact locations. Directions and permission to enter for qualified cavers can be obtained from the Ely BLM office: Star Route 5, Box 1, Ely, NV 89301.

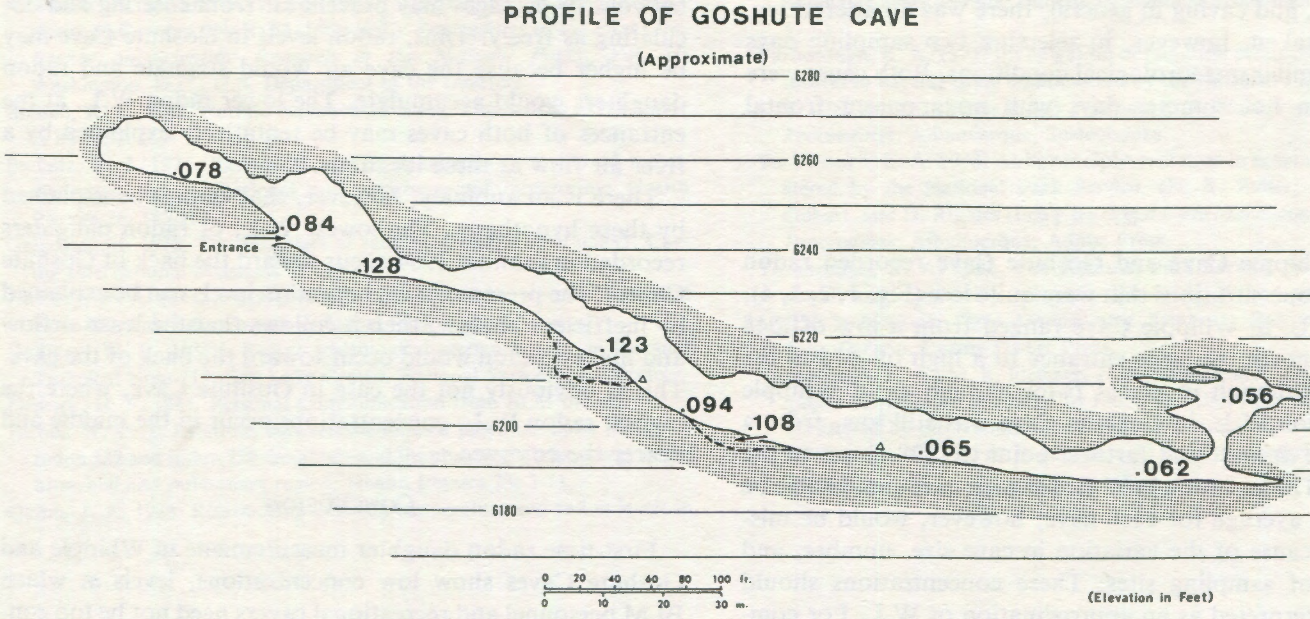

Figure 4. **Approximate cross-section of Goshute Cave, Nevada, showing radon W.L. as measured at various locations and depths.**

These two caves have several similarities. Both Whipple and Goshute are on north-south trending fault block mountains in the Great Basin (Fig. 1, 2, 3, 4). Their entrances are at approximately the same elevation (Whipple—7060 ft.; Goshute—6240 ft.) (Hedrick, 1985), and both are located near the top of the bajada at the base of their respective ranges. Geologists consider them to be developed in equivalent strata, Whipple Cave in the Whipple Formation and Goshute Cave in the Notch Peak Limestone (Hedrick, 1985). Both caves are designated at RSU because most of the passageways are below the entrances; therefore, airflow induced by temperature changes should be similar. Neither cave has current or recently occurring breakdown, and both have active speleothem formations. The breakdown that is evident in Whipple Cave occurred prior to speleothem deposition. Both caves are rather shallow in terms of length from entrance to rear. Whipple Cave is 800 ft. long and Goshute is 520 ft. long (Halliday, 1954a; 1954b). The floor at the entrance to each cave is partially filled with dirt and pulverized packrat and bat droppings.

The two caves differ in their orientation, internal configuration, and vertical depth from surface (Fig. 1, 2, 3, 4). Whipple Cave, on the western slope of its mountain range, is directly in the path of prevailing westerly winds. Goshute Cave, on the eastern slope of its range, is protected from most direct westerly winds. With regard to their interior geometry, Goshute is a more complex cave with a small entrance about 3 ft. high. Solution followed a jointed bedding plane that dips 30-40° northeast, creating rooms at slightly different levels joined at right-angles by inclined narrower passageways (Halliday, 1954b; Hedrick, 1985). The last room also has an upper unit. Crawlways are numerous and contribute to a total length of 2,000 feet (Hedrick, 1985). Whipple is a large, straight tube except for one constriction near the center. This main passageway follows a northeasterly joint in the limestone bedding which dips eastward approximately 30°, and there are no significant crawlways. The entrance to Whipple Cave is a 60 ft. vertical drop through a double sink opening 25 ft. in diameter. Whipple is the deeper cave at 150 feet; Goshute's maximum depth is 50 feet below the entrance. In summary, the configuration of Whipple Cave would seem to encourage air flow, while that of Goshute Cave would seem to restrict it.

METHODS

Radon daughters were measured at various locations in the caves on a one-time only basis using the Kusnetz method (Budnitz, 1974), the method generally used by NPS. Radon daughters were collected by drawing a known volume of air through a filter attached to a battery-operated hand-held pump. Each location was sampled for exactly five minutes. Radon daughters on the filters were then counted for 4 minutes on a scintillation-type detector with digital readout within 40 to 90 minutes of sampling. Six locations were sampled in Whipple Cave on August 23, 1984 and nine locations were sampled in Goshute Cave on September 5, 1984. It is acknowledged that results could be biased by sampling on different days, but due to the logistics of personnel, equipment, and caving in general, there was no alternative. Care was taken, however, in selecting two sampling days that had similar meteorological conditions. Both caves were sampled on hot summer days with no apparent frontal activity.

RESULTS

Both Whipple Cave and Goshute Cave recorded radon daughter concentrations that were quite low (Fig. 1, 2, 3, 4). Radon W.L. in Whipple Cave ranged from a low of .048 directly beneath the large entrance to a high of .073 at the far end. Although in places twice as high as in Whipple Cave, Radon W.L. in Goshute Cave was still low, from a .056 at the deepest and farthest point to .128 closer to the entrance. The variation in W.L. for each cave should also be noted. An average for each cave, however, would be misleading because of the variation in cave size, number, and locations of sampling sites. These concentrations should only be interpreted as an approximation of W.L. For comparison of radon W.L., Lehman Caves in the same general vicinity of eastern Nevada (in Great Basin National Park) frequently records radon W.L. of .500 to above 1.000 in the summer, depending on the day and location in the cave. Yarborough (1977) and the NPS recommends that caves with radon WL of less than .100 need not be monitored for radon, and that caves with radon W.L. between .100 and .200 be monitored annually. Providing that these one-time measurements are representative, the implication is that the casual caver who spends limited time in either Whipple Cave or Goshute Cave need not be concerned about encountering excessive radon. The limited data suggest that even on a hot summer day when the radon W.L. should be at its highest, the concentration in either cave may be considered insignificant.

Suggested Hypotheses

Although there are many similarities between Whipple Cave and Goshute Cave, the differences may be significant enough to influence the radioactivity levels in the two caves. The minor differences in radon daughter concentrations could be caused by anything that aids or hinders air flow. Recorded concentrations of radon daughters were generally lower in Whipple Cave. This lower W.L. may be tentatively attributed to the large west-facing entrance open to prevailing winds and also to the straight, uncomplicated passageway which could encourage airflow. Therefore, it is more likely that radon daughters in Whipple Cave would be more frequently diluted by fresh outside air. The generally higher W.L. recorded in Goshute Cave may be tentatively attributed to specifics about its location and configuration that could be expected to retard air flow. Its location on the lee side of the mountain, its small entrance, and complex network of passages may prevent air from entering and circulating as freely. Thus, radon levels in Goshute Cave may be higher because the cave air would stagnate and radon daughters would accumulate. The lower radon W.L. at the entrances of both caves may be tentatively explained by a freer air flow at these locations.

There is an anomaly, however, that cannot be explained by these hypotheses. The lowest W.L. of radon daughters recorded in summer 1984 occur toward the back of Goshute Cave. If the presence of higher radon levels can be explained by inefficient airflow, then it follows that the least airflow and highest radon would occur toward the back of the cave. This is obviously not the case in Goshute Cave, where the highest radon W.L. concentrations occur in the middle and nearer the entrance.

CONCLUSION

First-time radon daughter measurements in Whipple and Goshute Caves show low concentrations, levels at which BLM personnel and recreational cavers need not be too concerned. Whipple Cave, however, appears to have slightly lower radioactivity than does Goshute Cave. Two hypotheses are offered that may help explain and predict radon daughter concentrations on the basis of environmental factors other than exterior temperature and/or barometric pressure. It does appear that exposure to prevailing winds, entrance size, length and complexity of passageways may have an effect on radon daughter concentrations. However, much more study is needed before significant correlations and conclusions can be made.

ACKNOWLEDGMENTS

There are several persons and agencies without whose help this project could not have been completed. The National Park Service at Great Basin National Park (formerly Lehman Caves National Monument) is to be thanked for its generous loan of radon-monitoring equipment. The Bureau of Land Management in Ely readily provided maps and information concerning the caves. Technical assistance was provided by Norm Hiestand and Ed Wood, both of the NPS. The maps are based on surveys and sketch maps done in 1954 by W. R. Halliday of the NSS Salt Lake Grotto. Ed Wood also deserves special credit for creating rough sketches of the cave profiles.

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BOOK REVIEWS

Sedimentology, Dolomitization, Karstification, and Mineralization of the Leadville Limestone (Mississippian), Central Colorado. Richard H. DeVoto (ed.), Guidebook for Field Trip No. 6, Society of Economic Paleontologists and Mineralogists, Rocky Mountain Section Midyear Meeting, Golden, CO, August, 1985. 180 pp.

This publication has been overlooked by the cave and karst research community because of its limited distribution. Masquerading as a field guide and proceedings of an SEPM section meeting, it is much more than a field guide; it is a benchmark publication on an important mineral district in west-central Colorado. If you wonder why this is of importance to cave and karst practitioners, it become apparent when one realizes that the region is composed of limestones and dolomites and the mineral deposits are found in paleokarst features. This is an excellent example of the applied nature of karst research, showing there is an economic benefit in studying caves and karst processes. Although much has been written about the Leadville District and its fabulous mineral wealth, little has been published explaining its karst origin. The editor has assembled the leading experts on the topic and convinced them to publish their ideas in one place. Unfortunately, he was not able to publish it in a mainstream archived journal so that the work could reach a larger audience. It is an important scientific contribution that will undoubtedly stimulate more research into the relationship of paleokarst and mineral deposits not only in Colorado but elsewhere.

The fieldguide portion by Richard H. DeVoto is a very small part of the volume and is tacked onto the end of eight excellent chapters detailing the background, current status, geology, and theories about the mineralization of the Leadville District. In the first chapter, Richard H. DeVoto introduces the Mississippian stratigraphy of Central Colorado and shows evidence of extensive karstification in the past. It sets the stage on which the latter chapters are firmly based. This is followed by a detailed stratigraphy of the Leadville Formation by Rebecca A. Dorward describing the formation of the dolomites, processes acting upon the formation, formation of the zebra-textured dolomites, paleokarstic solution-erosion episodes, and the impact of Devonian and Mississippian faulting on subsequent deposition. The third chapter by Robert A. Horton classifies the dolomites into five types and discusses their formation, characteristics, and distribution. His detailed analysis of the supratidal environment, marine transgressions, hydrothermal saline solutions, and the impact of the hydrothermal activity associated with the Laramide and Tertiary igneous activity leads the reader through a complex stratigraphic record.

Chapter four by David W. Beatty shifts gears and uses oxygen and carbon isotope analysis to study 119 carbonate samples. Results show similar isotopic compositions with early signatures of the dolomitization having been obliterated by contact metamorphism, hydrothermal activity, and flushing with acidic solutions. While not giving any definitive answer about the staging of events, this chapter shows the complex processes acting upon the ore bodies. Next, Richard J. Tschauder and Gary P. Landis show the staging events from paleokarst formation to hydrothermal deposition by analyzing karst deposited breccias, dolomite sands, and dolomite flowstones. Richard H. DeVoto, in chapter six, discusses the incision of the major valleys in late Mississippian and the creation of a karst system in which one today finds the majority of lead, zinc, and silver deposits. These deposits are shown to be correlated with paleokarst fill rather than following the regional joint pattern as previous researchers have assumed. He contends that the mineralization occurred when hot basinal brines, transported by sandstone aquifers, mixed with meteoric water. The cooling, dilution, increase in oxygen fugacity, and change in pH associated with contact with the carbonate cave sediments caused the ore deposition.

The last two chapters do not fit into the otherwise smooth synthesis of the other chapters into an integrated work. Chapter seven by David W. Beatty et al. is simply a case study of field observations at the New Wyoming Decline in the Red Cliff District, showing further evidence of the link between paleo-cave systems and mineralization, but adding nothing new to the volume. The last chapter is an abbreviated contribution by Tommy B. Thompson et al. that can be perceived as a disclaimer of much of the previous work in the volume because the authors state that there has been extensive replacement in bedded dolomites as well. It is unfortunate that this section is not further developed so the reader could compare Thompson's results with those of DeVoto, Tschauder, and Landis with whom there is some disagreement.

In general, the volume suffers from the inconsistency of a conference proceedings, ranging from well developed scientific articles to short contributions that only whet one's appetite. The references at the end of each section give one an insight into a whole new arena on paleokarst which is often overlooked by mainstream cave and karst researchers. Although it may be difficult to find a copy of this publication, it is recommended for anyone interested in applied karst, paleokarst, or mineralogy. This information with some updating could be the basis for a special issue of the

NSS Bulletin or some similar journal where it could be available to a larger audience which it deserves.

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Resource Management in Limestone Landscapes: International Perspectives. David Gillieson and David Ingle Smith (eds.). Proceedings of the International Geographical Union Study Group on Man's Impact on Karst. Sydney, Australia, 1988. 260 pp.

This is the proceedings from the IGU Study Group on Man's Impact on Karst held at the IGU in Sydney, Australia and contains papers from the USA, Australia, Japan, China, Yugoslavia, England, and Italy. It is a fitting topic for the IGU since geography is a science interested in the interrelationship between people and the physical environment, a world in which over one-fifth of the land surface is susceptible to karst processes. Much of the karst research in the world is done by geographers and the IGU has a long and distinguished history of contributing to karst studies.

Half of the volume was written by Australian karst specialists and is about Australia, an area many American karst experts are unfamiliar except for the writings of Joe Jennings. Contrary to popular opinion, Australia is not a major karst locale because less than 3% is karst, much of that in aeolian calcarenites and deposits far from population centers. The Precambrian shield contains few carbonates and many large carbonate areas such as the Nullarbor Plain are located in arid areas; resulting in minimal karst development. With a low population density, Australia does not have the same management problems facing humid karst regions in Europe and North America, with the exception of the intrabasinal karsts of eastern Australia. Many cave and karst management problems facing Australia are unique to its environment.

The first section of the book, an overview of Australian karst with Andy Spate introducing the various karst regions, does a disservice to the regional approach by giving only a paragraph or two on each region; just enough to arouse one's interest but not enough to satisfy. A companion chapter by David Ingle Smith reviews the Australian karst aquifers on a regional basis. It is very well written, organized nicely, and contains excellent illustrations. David Gillieson's contribution on the effects of land use on karst is a good survey of environmental problems facing Australian karst regions; presenting examples from mining, forestry, recreation, grazing, and water supply development. An abbreviated article by Andy Spate on cave use and management from Aboriginal time to the present concludes the section. Several case studies on Australian karst regions appear next, the best being a thorough treatment of Tasmania by Kevin Kiernan showing how the physical geography has been modified by human settlement patterns and how karst management strategies can be used for both government and non-government areas. This is the best example of what a chapter should contain in a karst management book. A very short paper by Susan White on aeolian calcarenites, one of the largest Australian karst assemblages, provides a cave classification for management purposes and little else. This is followed by an interesting case study of the Tietkens Plain by Julia James; a case study on how nuclear testing in the 50's and 60's caused changes in this unique region, rather than a management plan or experimental design. Another contribution by Kevin Kiernan on karst management issues at the Jenolan Caves makes the problems at Mammoth Cave, Kentucky look pale in comparison; while American cave conservationists complain about roads over Mammoth Cave, Jenolan Caves has a major provincial highway running through the cave! This is an excellent analysis of the impact of surface activities and pollution sources on the cave and the impact of visitors in the cave. The last contribution in the section, stalagmites as monitors of environmental change by A. Goede, is an excellent piece of research using electron spin resonance (ESR) to date speleothems; but, it does not belong here, it is pure research with no application to karst management.

The second section of the book on overseas experiences in cave and karst management is a mish-mash with no unifying theme tying the articles together, as one would expect from a conference proceedings of volunteered papers. Yuan Daoxian of the People's Republic of China looks at the karst environmental system as an interdisciplinary effort that is a prerequisite for planning and management decisions, and proposes a karst classification based on climate regions. Ivan Gams presents a new definition of karst which is much more liberal than previously used standards, resulting in 27% of the total land surface being classified as karst. Limestone mining as an agency of landform change is discussed by John Gunn and Peter Gagen with particular reference to quarries in the Peak District of England. Erosion rates of identification of anthropogenic karst landforms are discussed, but only one paragraph addresses karst management. Kazuko Urushibara-Yoshino presents a case study on the red residual limestone soils of Japan's Nansei Islands showing how mismanagement of sugar cane plantations has resulted in soil compaction and fertilizer rich, sediment laden runoff that is killing coral. Continuing the theme of karst soils, David Gillieson reviews karst soil types and processes in the rain forests of the New Guinea highlands, showing how cultural practices have had beneficial and negative impacts. Michael Day and Carol Rosen discuss soil degradation and other environmental problems associated with the population pressure of Guatemalan and Salvadorian immigrants on the Hummingbird Karst of Belize; but no discussion of government mitigation techniques nor suggestions of how to manage the area are proposed.

While Japan is not noted for its karst regions, it is difficult to imagine that Kazuo Mitsui can discuss over 3000 caves, the variation of karst in Japan, and give special consideration to the problems in the Akiyoshi Plateau in only two pages of text, four maps, and two pages of photos. Equally short, the next section by Sauro Ugo and Meneghel Mirco is another set of legend symbols for cave and karst maps. Much of the text explains a color scheme for the symbols, but unfortunately the illustrations are printed in black ink. The final two papers are a joint effort by George Huppert, Betty Wheeler, E. Calvin Alexander, and Russell S. Adams, Jr. on the Coldwater Cave groundwater basin, Iowa. The first reports on a survey of farmer perception of the groundwater problem and how management practices have changed over the years. The small sample indicates that farmers perceive a problem with farm pollution and erosion but not on their farm. The second selection reports on groundwater quality monitoring at Coldwater Cave, especially for nitrate and atrazine since this is a corn belt karst region. Both studies are data collection projects to lay the base for future management of the area.

Most of the contributions in the volume do not address resource management in the traditional sense. They are case studies of environmental degradation in karst areas with few suggestions of how to ameliorate the problems. The organization of the book leaves much to be desired because the topics often have nothing in common. The quality of the works range from two-page extended abstracts to well documented scientific works. With the material at hand, the editors did a good job of editing and publishing the book with the exception of the missing colors in the Ugo selection. The metric size may cause some consternation to American readers for 8 inches by 11.5 inches is too large for many bookshelves. Paper and money could have been managed better if a smaller format and smaller point size had been chosen. In general, it is an interesting book that should be on library bookshelves for use by dedicated conservation types, although it is not suggested for purchase by all geoscientists.

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The Invertebrate Cave Fauna of Virginia and a Part of Eastern Tennessee. John R. Holsinger and David C. Culver. *Brimleyana,* No. 14. North Carolina State Museum of Natural Sciences, Raleigh, NC. 1988. 164 pp.

Available from the *Brimleyana* secretary, Sally Turner, NC State Museum of Natural Sciences, P.O. Box 27647, Raleigh, NC 27611.

This book-length monograph is the final contribution of the "Cave Biogeography of the Central Appalachians Study Group," a former N.S.S. research project. It summarizes a tremendous amount of work, including near twenty years' worth of field collections from over 500 caves visited by the authors, and an additional 47 caves in Virginia and Tennessee where data were obtained from the literature or from other researchers. These caves represent approximately 21% of recorded caves in the study area which lies within the Appalachian Valley and Ridge Province. Most collections were made directly from the substrate and, not surprisingly, most of the specimens are macroscopic invertebrates that are at least 1 mm in size. The authors indicate that investigation of microscopic forms would probably be worthwhile. The paper also includes a few comments about cave vertebrates, in particular the plethodontid salamander, *Gyrinophilus porphyriticus,* which is a major predator in some cave-stream communities in the study area.

The monograph is all I expected, and more. The central focus is the data which are arranged alphabetically by family, genus, and species within the phylogenetic taxa of phylum, class and order. Some troglobites are not yet completely classified within their genus or species group. Whenever possible the designation of troglobite, troglophile, trogloxene and accidental has been indicated, realizing that some designations may change. All known cave records are given by county for each species. This wealth of information is made more valuable by the critical comments preceding each major group. All of this takes up 91 of the 162 pages of the monograph.

The remainder of the monograph is devoted to a general review of cave ecology followed by a detailed zoogeographic analysis of each drainage basin in the study area as a regional cave faunal unit. The frequency of species exchange between basins is influenced by proximity; adjacent basins generally have more species in common. Species exchange is also influenced by the geologic structure and geographic extent of drainage divides; carbonate rock in divides is easier to penetrate than non-carbonate rock.

A discussion of the distributional pattern of troglobites in the study area showed that cave species diversity and ecological complexity increased in areas with many caves and extensive karst development. A regression of species density against cave density per unit area for each basin showed that 92% of the variation in the data can be explained by cave density.

An interesting preliminary application of the terrestrial species data to island biogeography is included. Can the number of species in a karst area or cave be determined by an equilibrium of immigration and extinction rates applied to individual caves, given an evolutionary time scale rather than a standard ecological one? With many caveats, the basic island analogy holds. More work is promised on this subject.

The origin, evolution, and dispersal of aquatic versus terrestrial species receives much more attention than is indicated in the abstract. The authors have good evidence for derivation of aquatic troglobites both directly from surface ancestors and indirectly from ancestral lineages living in subterranean waters.

The authors support the Pleistocene climatic effect paradigm for the evolution of terrestrial troglobites. A spirited discussion of alternatives follows, including Howarth's evidence for adaptive shifts of preadapted ancestors into newly opened niches, a theory which is especially attractive since it can be applied to caves worldwide. Barr's allopatric speciation process for troglobitic trechine beetles provides strong support of the Pleistocene climatic-effect model, but the authors point out cases where it could apply to Howarth's adaptive shift theory.

Some discussion of how dispersal may occur underground outside of caves, based on recent discoveries in Japan and Europe is included. Ueno found an extensive troglobitic fauna in mines and fissures in non-calcareous rocks in Japan. Juberthie and Delay discovered cracks and fissures in a shallow underground compartment which may be connected to caves or deep cracks. Some troglobites were characteristically found in this habitat. Further work outside of Europe is needed to determine the extent and importance of this shallow underground compartment.

Holsinger and Culver suggest that their study area may have been an important geographic center for the ancestors of troglobites in much of the Eastern United States, based on nearly continuous distribution of some troglobitic genera as evidence of a generalized distributional track.

Lastly, the authors indicate that some common cave species like *Meta menardi* spiders and *Enhadenoecus fragilis* crickets, are well-adapted as troglophiles and show no evidence of evolving into troglobites. A few other species in the study area do show signs of evolving into genuine troglobites.

A common theme throughout is the need for more work. The authors continually mention areas needing more study. For example, more than 50% of the cavernicolous millipedes in the eastern United States are undescribed. The authors indicate (with remarkable restraint) that this situation "makes zoogeographic analysis difficult."

The editing job was extremely well done for such a complex manuscript. Typos are nearly nonexistent. Some of the maps suffer from too much reduction, especially Figure 2 which shows the major drainage basins in the study area. The distributional maps would be more valuable if they all covered the same area and were the same size. Most of the pictures are good, but a few, like the amphipod and snail in Figure 13, lose a bit in the printing. The main problem with the paper is its publication in *Brimleyana,* an irregularly published journal of the North Carolina State Museum of Natural Sciences. Researchers outside the southeastern United States may be unfamiliar with this journal and miss Holsinger and Culver's fine monograph.

This monograph has much to recommend it. Any researcher in the study area will find it useful, but the value of the analysis and commentary extends to anyone interested in ecology, evolution, and zoogeography.

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Stygofauna Mundi: A Faunistic, Distributional, and Ecological Synthesis of the World Fauna Inhabiting Subterranean Waters. Lazare Botosaneanu, ed. E. J. Brill, 1986. 740 pp. $204.

In this book, Botosaneanu and 64 contributors review our knowledge of the organisms inhabiting subsurface waters, including marine sediments, caves, and a host of other habitats such as seeps and thermal springs. In the introduction, Botosaneanu argues that the inclusion of the marine interstitial fauna is justified because of its similar fauna, physiographic continuity, and probable role as a gate to colonization of freshwater subterranean habitats. Furthermore, he points out that all these habitats share the characteristics of little or no light, temperature variations of low amplitude, and impoverished trophic rsources. He also provides a useful summary of the 30 or more subdivisions of the subterranean habitat that has been proposed by one or more authors. This is an especially esoteric topic, including such terms as hypotelminorheic (perched water tables), and Botosaneanu wisely limits his discussion to an inclusive list with equivalent French, English, and German terms, when they exist.

Each chapter begins with a general overview of the morphology of the subterranean members of the group, illustrations of their morphology, a bibliography, and a table of all known subterranean species categorized according to both habitat and geographic area. About a third of the chapters are in French, a scattering in German, and the rest in English. To the nonspecialist, the overview and the illustrations are of most interest. Taken together they provide a veritable bestiary of "minor invertebrates" such as Kinorhyncha and Gnathostomulida, as well as more familiar groups such as beetles and fishes. The predominance of Crustacea in subterranean water is reflected in the fact that nearly half the book concerns the Crustacea. Many apparently primitive groups of Crustaceans are found largely (orders Syncarida and Mictacea) or exclusively (class Remipedia and orders Thermosbaenacea and Speleographacea) in subterranean waters. The quality of the line drawings is very high, with the unfortunate exception of the fish. The overviews of the various groups echo the controversies concerning the nature of adaptation to the cave environment, including the role of preadaptation and the importance of isolation from surface populations. Especially thoughtful discussions of these and other points can be found in the sections by Danielopol, Holsinger, Karaman, and Ruffo.

For the specialist the species lists should prove to be most useful. The editor and his contributors are to be commended for gathering all this information in one place. Unfortunately, several lists are rendered considerably less helpful by the use of taxonomic names that are no longer valid. In spite of this quibble, I think the book succeeds both as an introduction to a fascinating fauna and as a handbook for the specialist.

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Index to Volume 52 of the National Speleological Society Bulletin

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This index contains references to all articles and other items of importance published in volume 52 parts 1 and 2.

The index consists of three parts. The first of these is a **keyword index** which starts on **page 116**. Keywords include: unique words from the article title, cave names, geographic names, and descriptive terms. The second part is a **biologic names index** begining on **page 123**. These terms are Latin names of organisms discussed in articles. The third part is an alphabetical **author index** starting on **page 124**. Articles with multiple authors are indexed under each author.

Citations are of the following form: names of all authors in the order which they appear in the journal; title of the article; volume number and part number (separated by a colon); beginning and ending page (separated by a dash); and year of publication from the cover of the issue. Volume number and year are included in the citation so that their format will match that of the cumulative indices of earlier volumes. Within an index group, such as Archaeology, the earliest article is cited first, followed by consecutive articles.

Index data was input on an IBM-PC using the SDI-Soft front-end program designed by Keith Wheeland. The index was prepared on an IBM 4341 computer running a VM/CMS operating system. Indexing was performed by the IBM KWIC/KWOC program as modified by William H. Verity at The Pennsylvania State University Center for Academic Computing. Formatting was accomplished using the SCRIPT text formater, and Generalized Markup Language, with camera-ready copy produced on a Xerox 2700 laser printer.

The author wishes to thank William H. Verity for his assistance. Computer funds were provided by the College of Earth and Mineral Sciences, The Pennsylvania State University.

Keyword Index

Biologic Names Index

Author Index

The *NSS Bulletin* is a multidisciplinary journal devoted to speleology, karst geomorphology, and karst hydrology. The *Bulletin* is seeking original, unpublished manuscripts concerning the scientific study of caves or other karst features. Authors need not be associated with the National Speleological Society.

Manuscripts must be in English with an abstract, conclusions, and references. An additional abstract in the author's native langauge (if other than English) is acceptable. Authors are encouraged to keep in mind that the readership of *The Bulletin* consists of both professional and amateur speleologists.

For general style refer to the present *Bulletin* and the following guides: "Suggestions to Authors" (U.S. Geological Survey), "Style Manual for Biological Journals" (American Institute of Biological Sciences), and "A Manual of Style" (The University of Chicago Press). For assistance in writing an abstract see "A Scrutiny of the Abstract" by K. Landes, Bulletin of the American Association of Petroleum Geologists, vol. 50 (1966), p. 1992. Because good figures are an essential part of any paper, authors are encouraged to see what bad figures look like in the editorial on figures by K. Rodolfo in the *Journal of Sedimentary Petrology*, vol. 49 (1979), p. 1053-60.

Each paper will contain a title with the author's name and address. This will be followed by an abstract and the text of the paper. Acknowledgements and references follow the text. References are alphabetical with senior author's last name first, followed by the date of publication, title, publisher, volume, and page numbers. See the current issue of *The Bulletin* for examples.

Authors should submit two copies of their manuscript (include only copies of the illustrations) to the appropriate specialty editor or the senior editor. The manuscript must be typed, double space on one side of the page. Authors submitting manuscripts longer than 15 typed pages may be asked to shorten them. All measurements will be in Système Internationale (metric). Other units will be allowed where necessary if placed in parentheses and following the SI units.

Figures and lettering must be neat and legible. Figure captions should be on a separate sheet of paper and not within the figure. Most figures will be reduced, hence the lettering should be large. Once the paper has been accepted for publication, the original drawings (with corrections where necessary) must be submitted to the editor. Black-and-white photographs must be sharp, high contrast, and printed on glossy paper. Color prints will be printed at authors expense only.

All submitted manuscripts are sent out to two specialists for review. Reviewed manuscripts are then returned to the author for consideration of the referee's remarks and revision (where necessary). Revised manuscripts are returned to the appropriate editor who then recommends acceptance or rejection. Upon acceptance, the author should submit all photographs and original drawings to the editor.

Once the paper has been typeset and laid-out, the senior author will be sent one set of proofs for review. Any corrections other than printer errors will be done at the author's expense. A reprint order form will be sent with the proofs. At this time all authors will be requested to contribute page charges of $25 per page to help defray the cost of publication. The actual cost to the society is about $100 per page. Acceptance of manuscripts for publication is not contingent upon payment of page charges.

A new speleological journal has been produced by Institute of Geography of the Czechoslovak Academy of Sciences. The journal is entitled *Studia Carsologica*. Its primary language will be English. Its focus is the publication of articles etc. on the effects of human activity on caves and karst. However, other karst topics will not be rejected out of hand.

Three volumes have already been published and Volume 4 is expected out soon. I have been appointed to the Editorial Board representing North America. If you have appropriate articles, book reviews, or other items, please send them to me for consideration.

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Since the economy in Czechoslovakia is not stable as of yet, they have not determined a price for the journal. At this time they are trying to see what the economy does and what the demand is. They are looking for subscribers. Typically, items coming out of eastern Europe have been quite reasonable. To receive the first three volumes and get on the mailing list, write to:

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CZECHOSLOVAKIA

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