THE NSS BULLETIN

QUARTERLY JOURNAL

OF THE

NATIONAL SPELEOLOGICAL SOCIETY

VOLUME 36 **NUMBER 3**

Contents

JULY 1974

(formerly: BULLETIN OF THE NATIONAL SPELEOLOGICAL SOCIETY)

NATIONAL SPELEOLOGICAL SOCIETY

The National Speleological Society is a non-profit organization devoted to the study of caves, karst, and allied phenomena. It was founded in 1940 and is incorporated under the laws of the District of Columbia. The headquarters of the Society now are located in Huntsville, Alabama. The Society is affiliated with the American Association for the Advancement of Science.

The Society serves as a central agency for the collection, preservation, and dissemination of information in fields related to speleology. It also seeks the preservation of the unique faunas, geological and mineralogical features, and natural beauty of caverns through an active conservation program.

The affairs of the Society are controlled by an elected Board of Governors, which appoints national Officers. Technical affairs of the Society are administered by specialists in fields related to speleology through the Society's Biology Section, Section on Cave Geology and Geography, and Research Advisory Committee.

Publications of the Society include the quarterly BULLETIN, the monthly NEWS, an annual SPELEODIGEST, SPECIAL PUBLICATIONS, and the CAVING INFORMATION SERIES. Members in all categories except Family receive the BULLETIN and NEWS.

A library on speleological subjects is maintained by the Society at its office. Material is available to Society members at a nominal fee to defray the cost of handling and to others through inter-library loan. An extensive file of information on caves of the United States is maintained by the Society and currently is housed in Reston, Virginia.

A Survey of Springs Along the Green and Barren Rivers, Central Kentucky Karst

John W. Hess *, Steve G. Wells †, and Thomas A. Brucker ‡

ABSTRACT

Study of the springs in the Central Kentucky Karst is an important link in developing an understanding of the hydrogeology and geochemistry of the ground water in that region. The locations and descriptions of 105 springs discharging along the Green and Barren rivers were compiled over a three-year period, disproving the assumption that the Central Kentucky Karst aquifer is drained by a few large springs. Temperature difference and electrical conductivity methods were used for the verification and location of the springs and are considered to have been valuable adjuncts to visual detection. Eighty-one springs along the Green River between Munfordville and Brownsville, Kentucky and 24 springs along the Barren River between Polkville and Bowling Green, Kentucky were located and described. Most of these springs are new to the literature on the area. They are classified as alluviated or as gravity springs, based on their morphology at pool stage of the rivers, and as large or as local springs, based on their discharge and specific conductance.

Introduction

The purpose of this paper is to report the locations of and to provide brief descriptions of springs along the Green and Barren rivers, Central Kentucky Karst, and to demonstrate the usefulness of the temperature difference and electrical conductivity methods of spring verification and location. The relationship of the springs to the drainage system of the Karst area also is discussed. Most of the springs were located visually by their characteristic modifications of the river bank and were then verified using a thermistor thermometer and a specific conductance bridge. This report represents a small segment of two separate, large projects by Hess and Wells dealing with the hydrogeology and geomorphology of the Central Kentucky Karst.

Springs located prior to this study were found by enquiring of local people who use the rivers and by making visual observations from boats. Only the larger, more obvious springs were found by these methods and it was assumed that the Central Kentucky Karst aquifer system is drained by a few large springs. This is not the case. A great many smaller springs supplement these few large springs in discharging the aquifer.

The study area (Fig. 1) is a part of the Central Kentucky Karst, which lies in the Interior Lowlands Province in the south-central portion of Kentucky, approximately 100 mi south of Louisville, Kentucky. The Central Kentucky Karst is developed in a sequence of Mississippian limestones 530 ft in thickness. These limestones have an average dip to the northwest of 30 ft per mile and lie at the southeastern edge of the Illinois Basin, in a broadly synclinal indentation on the regional structure. The area is divided into two physiographic units by the Chester (Dripping Springs) Escarpment. The Sinkhole Plain, a limestone plateau of low relief, lies to the southeast. To the northwest, are the rugged, dissected, clastic-capped ridges of the Chester Cuesta (Mammoth Cave Plateau).

* Department of Geosciences, The Pennsylvania State University, University Park, 16802

† Department of Geology, University of Cincinnati, Cincinnati, Ohio 45221

‡ 445 W. South College Street, Yellow Springs, Ohio 45387

The traditional boundaries of the Central Kentucky Karst are delineated hydrologically (White, *et al*, 1970). The northern boundary is the drainage divide between Green and Rough rivers. In the south, the divide is the interfluve between Green and Barren rivers. The rim of the Little Barren River basin near Munfordville is denoted as the eastern boundary. To the west, the boundary is near Brownsville, where the Green River has not yet dissected the clastic caprock. Recent investigations by Miotke and Papenberg (1972) and by Wells indicate that the traditional boundaries should be modified to include the subterranean tributaries of the Barren River. Dye tracing conducted by Miotke, Papenberg, and Wells in the spring and summer of 1972 revealed the existence of a large karst drainage basin feeding the Graham Spring complex on the Barren River.

The Green River, the master base level for the region, flows westward some 40 mi through the study area. The Barren River is a tributary of the Green River and flows across the southwestern portion of the Central Kentucky Karst. It serves as the base level for much of the southwestern portion of the Sinkhole Plain.

The location, description, and study of all springs are important in understanding the hydrogeology and geochemistry of a region. Spring water is a composite of all of the various types of waters making up the ground water system. Its physical and chemical characteristics provide information about the flow system behind the spring. Such parameters as catchment area, rock type, and residence time can be inferred by study of the spring.

Most of the springs described here have not previously been reported in the literature. On the Green River, Pike, Styx, Echo, and Turnhole springs have been mentioned by most previous authors discussing the springs of the area. In addition, Watson (1966) mentions Blue and Hix springs, but their specific locations are not designated; White, *et al* (1970) report a "Blue Spring" on the north side of the river; Cushman, *et al* (1965) point out that many large springs are situated at the mouths of the hollows leading to the river; Deike (1967) and Brown (1966) add Sand Cave Spring and Garvin Spring to the list of springs along the Green River. These are but a small percentage of the 81 springs located along the Green River between Munfordville and Brownsville, Kentucky (A-A′ on Fig. 1). Of those

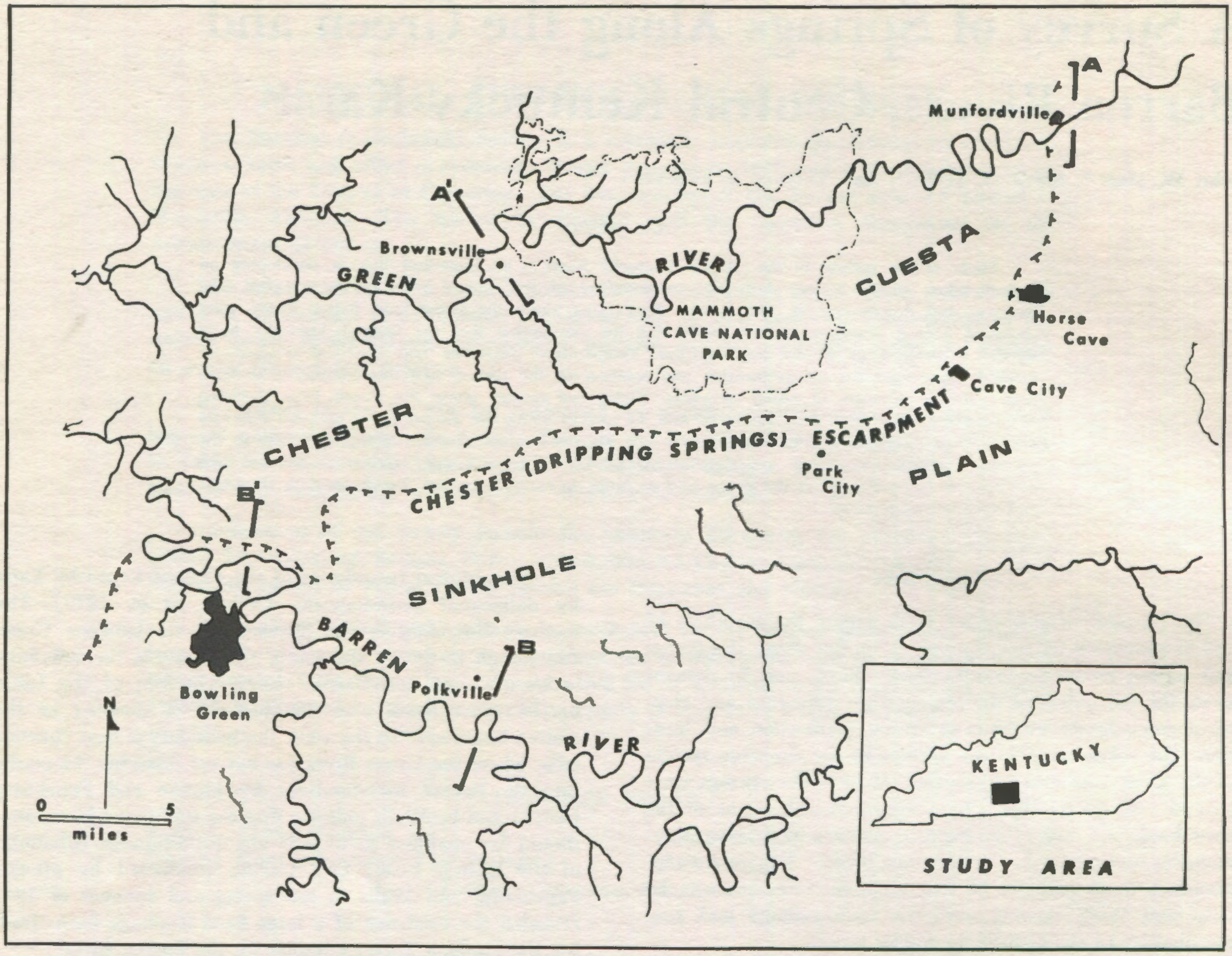


Fig. 1. Generalized map of the Central Kentucky Karst. The Green River, flowing from east to west, was investigated for springs from A to A′. The Barren River, flowing from southeast to northwest, was investigated from B to B′.

springs located on the Barren River, Graham Spring is mentioned in a report prepared by the Kentucky Department of Commerce in coöperation with the U. S. Geological Survey. To this, we add 23 springs between Polkville and Bowling Green, Kentucky (B-B′ on Fig. 1).

The flow directions of the springs are controlled by the direction of the difference between the hydrostatic head caused by the stages of the rivers and the head caused by recharge from the catchment areas. Thus, at the time of observation, a spring may be in a state of discharge to the river, of zero flow, or of recharge to the carbonate aquifer.

Spring Location and Verification Methods

The springs described in this paper were for the most part detected visually and then verified using either the temperature difference method or both the temperature difference and the electrical conductivity methods. Most of the springs modified the river bank in such a way as to indicate their presence. The temperature difference method uses a thermistor to detect temperature changes along the river and the electrical conductivity method uses a specific conductance bridge to detect variations in the electrical conductivity of the water. Spring water will differ in temperature and specific conductance (Spc) from those of the river and of surface water inputs to the river. A very small percentage of the springs were located on the basis of temperature changes only.

The field method used was to float with the current along the bank of the river and to watch for evidence of springs while towing a thermistor alongside the boat to detect any temperature changes. Evidence of springs ranged from small, subtle notches under trees along the river bank to resurgence rivers up to several thousand feet long. Some of the springs flow from open cave mouths, while others have rise pools ranging from a few feet up to 100 ft in diameter. In many cases, morphological evidence was not enough to verify the smaller springs. Small notches, for example, could be caused either by a spring or by bank slumpage due to a fallen tree. In these cases, the temperature difference and electrical conductivity methods came into use. Observations were made in summer or early fall, when the river water—spring water temperature contrast was at a maximum. River temperature ranged between 18° and 26°C and that of the springs between 11° and 20°C. Summer also is the low-water period for the rivers and the season at which morphological evidence for the springs is most visible.

Specific conductance as well as temperature was measured during the summer of 1972. It has been shown by Bray (1969) and by Stenner (1969) that, for areas with relatively pure carbonate waters, Spc is proportional to the hardness of the water. The springs, generally speaking, have Spc's different from those of the rivers, although they may be either higher or lower, depending on the hydrogeologic characteristic of the spring. Parameters such as flow path and residence time will affect the Spc. The river, being a composite of all types of spring and surface waters, would be expected to have an intermediate Spc value.

At the time a spring was discovered, an estimate of its discharge was made, its temperature and Spc were recorded, and a brief description of it was written. Most of the springs on the Green River were located before the Spc meter was obtained and the Spc of those was measured later.

Spring Classification

The springs located along Green and Barren rivers can be placed into the 2x2 classification shown in Table 1. Each is classified either as an alluviated or as a gravity spring on the basis of its morphology at pool stage of the river and as a large or as a local spring according to its discharge and Spc.

TABLE 1. Spring Classification

	ALLUVIATED	GRAVITY
LARGE	1. Discharges below pool stage of river. 2. High discharge. 3. Relatively high Spc.	1. Discharges at or above pool stage of river. 2. High discharge. 3. Relatively high Spc.
LOCAL	1. Discharges below pool stage of river. 2. Low discharge. 3. No characteristic Spc.	1. Discharges at or above pool stage of river. 2. Low Discharge. 3. No characteristic Spc.

Alluviated springs discharge below river level and have rise pools that are fed at depth. Pike and Turnhole springs, two alluviated springs on the Green River, have been examined by scuba divers. Both are fed from large conduits approximately 30 ft below pool stage. Graham Spring, on the Barren River, is another example of alluviated spring (Fig. 2). The large rise pool of Graham Spring is typical of this type of spring. Springs whose outlets are unconfined and which are free-flowing are gravity springs. Spring No. 14 on the Barren River (Fig. 3) is typical of gravity springs.

Fig. 2. Outlet 3 of the Graham Spring complex serves as the principal resurgence during both high and low discharge periods. It is a typical example of alluviated spring. The photograph was taken during a period of high discharge in the winter of 1972. Photo by Steve Wells.

The large springs have relatively high discharges and Spc's. They must receive their major recharge from the Sinkhole Plain, which is the only area capable of providing the catchment area necessary to maintain high discharge and the long flow path and residence time necessary to produce high Spc. The local springs have lower discharges and a wide range of Spc. The discharge of these springs is smaller due to their more limited catchment areas and their Spc is very variable because of the great variation in flow paths and residence times among the springs.

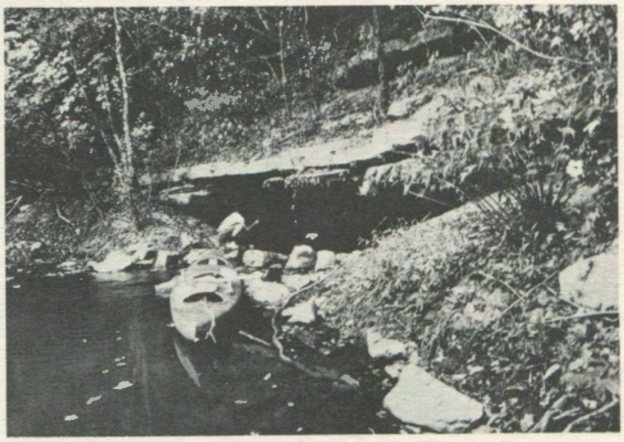


Fig. 3. Spring No. 14 along the Barren River flows from a open cave 3½ ft high and 15 ft wide. It is an example of free flowing gravity spring. Photo by Steve Wells.

Springs along the Green River

The search for springs along the Green River was carried out during the summers of 1970 and 1971; some of the larger springs were rechecked in 1972 in order to record their Spc. Fig. 4 indicates the locations of 81 springs along the Green River between Munfordville and Brownsville, Kentucky and their classification as gravity or alluviated springs. Table 2,[1] keyed to Fig. 4 by spring number, gives the name of the spring (when known), the date of observation, its estimated discharge (cfs), its temperature (°C), its Spc (μmhos at 25°C), and additional comments.

At this time, only five springs have been placed in the large spring classification: Gorn Mill (No. 3), Hicks (No. 11), Garvin (No. 14), Blue Spring South (No. 21), and Turnhole (No. 64) springs. Of these, only Turnhole Spring has been linked to the Sinkhole Plain by dye tracing (Miotke and Papenburg, 1972). Classification of the other four as large springs is based on their discharge and Spc. Future dye-tracing studies may very well lead to changes in the list of springs in this group. The primary sources of these springs are the sinking streams and sinkholes on the Sinkhole Plain and back-flood water from the Green River.

All other springs along the Green River are classified as local springs. These are springs the sources of which lie in the Chester Cuesta and in the valleys and hollows dissecting it. Their headwaters consist of vertical shaft complexes conducting water from the cuesta and its valleys down into the karst aquifer (Brucker, *et al*, 1972). These springs also are back flooded from the Green River.

[1] Unabridged copies of Tables 2 and 3 are available free of charge from: NSS Cave Files Committee, Cave Avenue, Huntsville, Alabama.

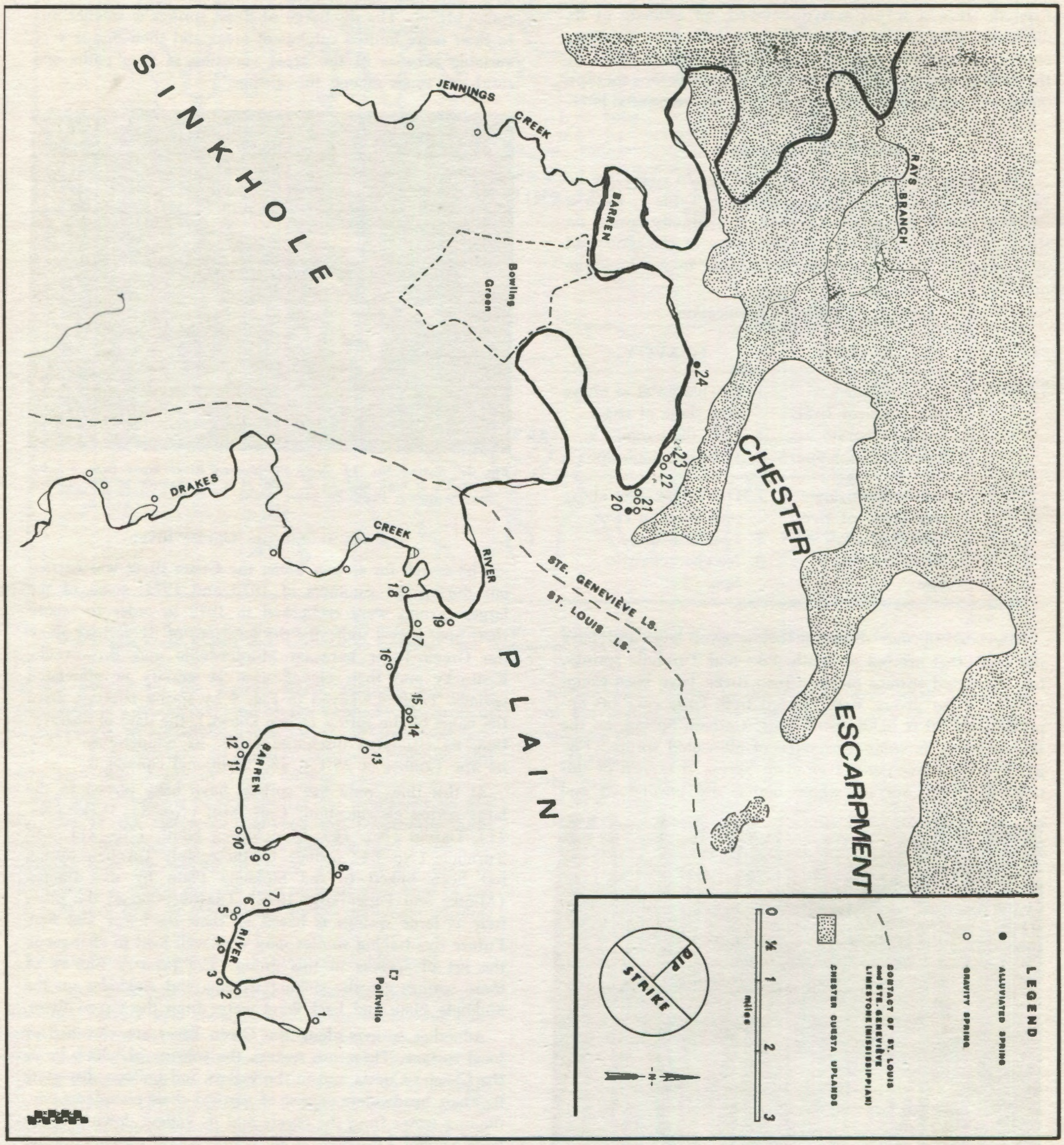


Fig. 4. Map of the Green River between Munfordville and Brownsville, Kentucky, showing the locations of 81 springs.

TABLE 2. Springs Along The Green River

Number	Date	Name	Discharge (cfs)	Temperature (°C)	Spc (μmhos) @ 25°C	Comments
3	26 November 1970	Gorn Mill	34	13.9	430	30′ diameter rise pool 70′ back from the river
10	24 July 1970	Hick's (Hix?)	—	—	—	small rise pools with cold air coming off rocks above
12	24 July 1970	Blow Hole	none	14.5	—	several thousand feet of cave at river level
14	26 November 1970	Garvin	3	13.8	422	1100 feet of cave passage back to a sump
21	26 November 1970	Blue Spring S	2	13.7	350	20′ diameter rise pool 600′ back from the river
27	26 July 1970	Blue Spring N	—	13.5	—	large rise pool at rock cliff, largest spring on the north side of river
39	11 August 1972	Pike Spring	8	13.5	225	from a 30′ wide by 6′ high conduit 25′ below pool stage of the river
54	22 September 1972	River Styx	.1	15.8	235	rise pool under rock cliff 500′ back from the river
55	22 September 1972	Echo River	2	14.1	225	rise pool at rock ledges 1200′ back from the river
64	20 October 1972	Turnhole	20	13.9	422	100′ diameter rise pool 30′ deep at river edge

Each of the hollows leading to the river has its own subterranean drainage system. Surface water sinks near the clastic rock—limestone contact, except in periods of very high runoff. This water then either enters the regional subterranean drainage network and is conducted to a large spring or remains isolated and is discharged at a local spring along the river near the mouth of the hollow. This tendency for each hollow to have its own subterranean drainage system accounts for the large number of small springs on either side of the Green River.

Springs along the Barren River

A detailed investigation of springs along the Barren River between Polkville and Bowling Green, Kentucky was conducted late in the summer of 1972. A total of 24 springs were located (Fig. 5). Their hydrologic and morpologic characteristics are given in Table 3;[1] the majority of these springs are gravity springs.

Twenty of the 24 springs along the Barren River are outlets for the ground water systems of the Sinkhole Plain. Only one of these, alluviated Graham Spring (No. 20), is a large spring. The other 19 springs draining the Sinkhole Plain are local, gravity springs. They have relatively small, isolated drainage basins compared to complex Graham Spring, the drainage basin of which has been calculated to be approximately 140 sq mi in area.

An interesting series of local springs occurs on the south and west sides of the Barren River southwest of Polkville. These, springs Nos. 6, 9, 10, and 11, are the only springs thus far reported in the Central Kentucky Karst which have considerable amounts of travertine deposited around their outlets. All have high Spc's (512 to 550 μmhos), indicating relatively high hardnesses.

The major spring system draining the Sinkhole Plain east and north of the Barren River is Graham Spring (No. 20). This complex system is described by Miotke and Papenberg (1972) as having five major outlets along a resurgence river approximately 1500 ft long and another outlet on the Barren River just downstream from the mouth of the resurgence

TABLE 3. Springs Along The Barren River

Number	Date	Name	Discharge (cfs)	Temperature (°C)	Spc (μmhos) @ 25°C	Comments
6	15 September 1972	—	.01	13.0	550	spring complex approx. 35′ above river, only 2 springs flowing, area has abundant travertine deposits
12	13 September 1972	Hardcastle	2.0	12.0	544	spring complex with major outlet 25′ above river, small cave beside spring with north-south trend, spring is 150′ down river from Number 11
13	15 September 1972	Gotts Spring	1.75	14.0	411	several spring outlets nearly 1500′ from river and 25′ above it, some outlets have collapse and breakdown obscuring the discharge
20	12 September 1972	Graham Springs	20			
		Outlet 3	—	15.0	494	spring complex with 5 outlets, 3
		Outlet 4	—	13.0	486	have resurgence pools, total dis-
		Outlet 5	.05	15.0	454	charge for outlets 1-4 is 20 cfs

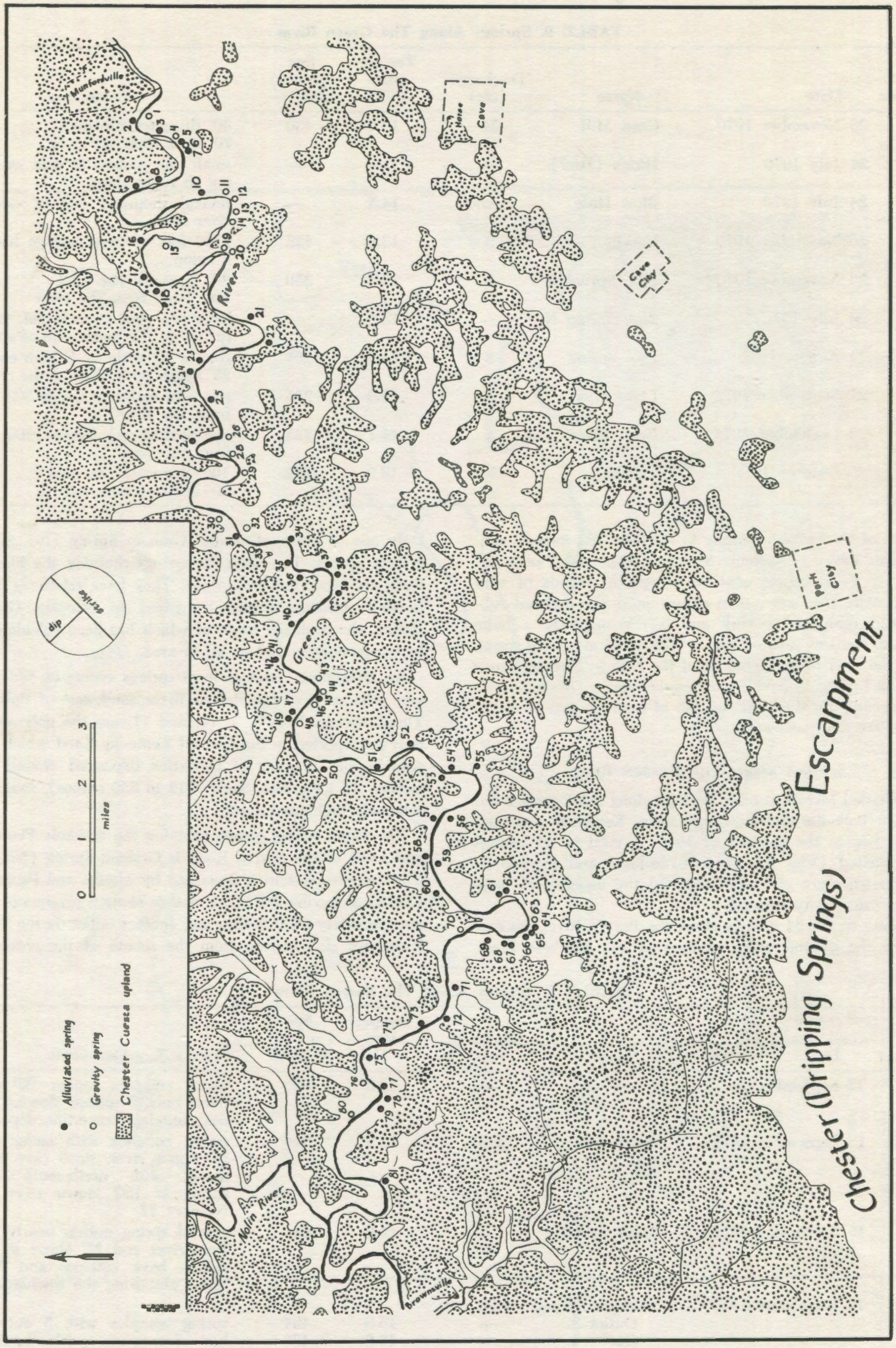


Fig. 5. Map of the Barren River and major tributaries between Polkville and Bowling Green, Kentucky, indicating the locations of 24 springs.

river. Two large, alluviated outlets, shown in Fig. 6, occur at the bases of large alcove bluffs at the head of the resurgence river. As pointed out by Miotke and Papenberg, the primary outlet is the southern one. However, the master outlet for Graham Spring is 600 ft downtream from the first two outlets (Outlet 3, Table 3 and Fig. 2). During low discharge periods, there is no flow from the first two outlets. Dye Tracing by Miotke and Papenberg and by Wells has shown the existence of a large subterranean drainage basin resurging at Graham Spring. It is comparable hydrologically to Gorn Mill and Turnhole springs on the Green River.

Downstream from Graham Spring are four other local springs, Nos. 21 through 24. Spring No. 24 is the only other alluviated spring on the Barren River. Its large "blue hole", approximately 30 ft in diameter, is at the head of a 200 ft long resurgence river.

Fig. 6. The two large springs at the head of the resurgence river of Graham Spring are typical of the alluviated or "rise pool" springs found along the Green and Barren rivers in the Central Kentucky Karst. During low discharge periods, water from these two outlets is diverted to the next outlet downstream through a subterranean conduit and the surface channel remains dry. Photo by Steve Wells.

Springs that drain to the Barren River or its tributaries can be arranged into the following hydrologic units: 1) The Sinkhole Plain north and east of the Barren River. Most drainage is to Graham Spring, except for 15 small, separate drainage basins upstream from Graham Spring. 2) The Sinkhole Plain south and west of the Barren River and east of Drakes Creek. Most drainage is north to the Barren River, oblique to strike. 3) West of Drakes Creek (south and west of the Barren River), karst ground water systems are tributary to Drakes Creek and to Jennings Creek, which serve as local base levels. 4) There exists a possibility that springs may exist on the south and west sides of the Barren River between Drakes Creek and Jennings Creek.

Summary

A census of springs discharging (and recharging) a karst aquifer is essential in interpreting the hydrogeology and geochemistry of the karsted area. The use of temperature difference and of electrical conductivity methods to verify the location of springs is superior to reliance on visual methods, alone. The temperature difference method uses a thermistor thermometer to detect changes in water temperature; the electrical conductivity method uses a specific conductance bridge to detect changes in the electrical conductivity of the water. These measurements should be made in summer or early fall, when the temperature contrast between springs and the rivers into which they discharge is at a maximum.

Using the above methods to verify new springs along the Green and Barren rivers in the Central Kentucky Karst, the authors were able to identify 81 springs along the Green River between Munfordville and Brownsville, Kentucky and 24 springs along the Barren River between Polkville and Bowling Green, Kentucky. This disproves the commonly made assumption that mature karst aquifer systems are drained by a few big springs. The Central Kentucky Karst aquifer is drained not only by a few large springs, but also by a large number of smaller springs. The springs were classified as alluviated or as gravity springs, based on their morphology at pool stage of the river, and as large or as local springs, based on their discharge and Spc.

Acknowledgements

The authors thank the Cave Research Foundation for field support and for use of the Flint Ridge Field Station. This work also has been supported by the Office of Water Resources Research of the U. S. Department of the Interior, The Pennsylvania State University, the Ohio Academy of Sciences, and the NDEA Program. Mammoth Cave National Park personnel are thanked for their help and coöperation and Roger Brucker is thanked for his critical reading of the manuscript.

LITERATURE CITED

Anon (ca. 1966)—Graham and Lost River Springs, a Potential Source of Industrial Water Supply: Ky. Dept. Commerce.

Bray, L. G. (1969)—Some Notes on the Chemical Investigation of Cave Waters: *Cave Research Group Great Britain, Trans.* **11**:165-173.

Brown, R. F. (1966)—Hydrology of the Cavernous Limestones of the Mammoth Cave Area, Kentucky: *U. S. Geol. Survey, Water-Supply Paper* 1837, 64 pp.

Brucker, R. W.; J. W. Hess; W. B. White (1972)—Rôle of Vertical Shafts in the Movement of Ground Water in Carbonate Aquifers: *Ground Water* **10**:5-13.

Cushman, R. V.; R. A. Krieger; J. A. McCabe (1965)—Present and Future Water Supply for Mammoth Cave National Park, Kentucky: *U. S. Geol. Survey, Water-Supply Paper* 1475-Q, pp. 601-647.

Deike, G. H. III (1967)—The Development of Caverns of the Mammoth Cave Region: Thesis, The Pennsylvania State University, 235 pp.

Miotke, F.-D.; H. Papenberg (1972)—Geomorphology and Hydrology of the Sinkhole Plain and Glasgow Uplands, Central Kentucky Karst: Preliminary Report: *Caves and Karst* 14:25-32.

Stenner, R. G. (1969)—The Measurement of the Aggressiveness of Water towards Calcium Carbonate: *Cave Research Group Great Britain, Trans.* **11**:175-200.

Watson, R. A. (1966)—Central Kentucky Karst Hydrology: *NSS Bulletin* **28**:159-166.

White, W. B.; R. A. Watson; E. R. Pohl; Roger Brucker (1970)—The Central Kentucky Karst: *Geog. Rev.* **60**: 88-115.

Manuscript received by the Editor 2 June 1973
Revised manuscript accepted 14 January 1974

Silica Deposits in Eastern Wyoming Caves

Paul L. Broughton *

ABSTRACT

A unique assemblage of silica stalactites, stalagmites, and related speleothems occurs in limestone caves in eastern Wyoming. Source of the silica is Oligocene ash beds overlying the Pennsylvanian Hartville limestone. The silica stalactites occur in conical and cylindrical forms up to 40 cm in length. The silica was deposited under low pH conditions and, at least partially, under ponded water.

Introduction

Silica deposits in limestone caves are rare. When present, the silica usually replaces collapse breccia and other cave fills. It is generally believed to be hydrothermal in origin (Quinlan, 1972; Dings and Whitebread, 1965). Silica speleothems in non-carbonate caves have been reported in the lava tubes of western United States (Peck, 1962; Anderson, 1930; Swartzlow and Keller, 1937) and in sandstone caves beneath outcrop ledges in Australia (Lassak, 1970). The only previously described silica speleothems in caves developed in carbonate rock are in Jewel Cave and Wind Cave in the Black Hills of South Dakota (Broughton, 1971; Deal, 1964; White and Deike, 1962), and a single opaline speleothem in an Argentine cave (Siegal, *et al.*, 1968). None of these have the conical or cylindrical form of carbonate stalactites. Instead, they are irregular, somewhat helical in form, and are thin encrustations on carbonate or detrital chert cores.

Silica speleothems recently discovered in several related limestone caves in Wyoming may be the most significant found to date and are the finest examples yet known of stalactitic silica morphology.

Eastern Wyoming Silica Speleothems

Several solutional caves in limestone of the Pennsylvanian Hartville formation in Platte County, Wyoming are characterized by a unique assemblage of silica speleothems. Silica stalactites were found by this author in more than a dozen small caves in nearly horizontal, gray, dolomitic limestone of the Meek member of the Hartville formation. For the general geologic setting, see Mallory (1972). The caves occur as linear, sub-parallel passages from 2 to 12 m in length and up to 6 m in diameter. They are developed approximately 8 m stratigraphically above the basal Meek sandstone, a calcite-cemented orthoquartzite.

Groundwater circulation along nearly vertical joints above the caves is indicated by the precipitation of silica along them. Vertical veins of silica along joint surfaces may be traced for as much as 14 m above the beds in which the caves are developed. The source of the silica is ash beds in the overlying White River formation. The silica speleothems occur where vadose circulation along joints encountered caves and deposited silica in stalactitic forms (Fig. 1), as flowstone along cave walls (Fig. 2), and as accumulations up to 30 cm thick on the floor (Fig. 3). No vadose seepage was evident when the caves were visited in April of 1972 and July of 1973. It is presumed that silica deposition is no longer taking place.

* Saskatchewan Department of Mineral Resources, Regina, Saskatchewan, Canada.

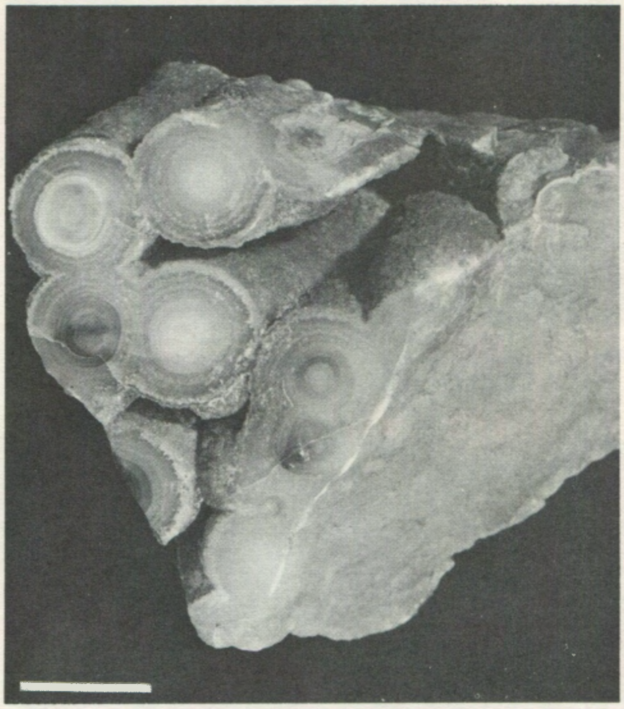


Fig. 1. Chalcedony stalactites showing internal concentric growth structures. Scale: 2 cm.

The vadose water responsible for the leaching of silica from the ash beds and its subsequent precipitation is thought to have been at normal (near-surface) temperature and pressure. Derivation of the silica from ascending hydrothermal water is discounted because of a lack of silicification below the basal sandstone and other physical field evidence. Examination of thin sections of the Meek limestone and basal sandstone does not indicate any hydrothermal silica veinlets or alterations.

The pre-existing clastic cave sediment is well preserved (Fig. 3). Most clasts are orthoquartzite, but relict carbonate clasts firmly cemented by the silica permeation are present. The source of the orthoquartzite clasts in these limestone caves is thought to be the subjacent basal sandstone. Nearly all of the original carbonate in the clasts has been replaced by silica that probably was amorphous (opaline) when originally deposited but which now is recrystallized to microcrystalline quartz and length-fast chalcedony. Only very corroded traces of this original carbonate cement remain.

Most of the silica speleothems are thought to have been directly precipitated. However, thin bands of calcite growth lines (Fig. 4), now partially replaced by silica, are in-

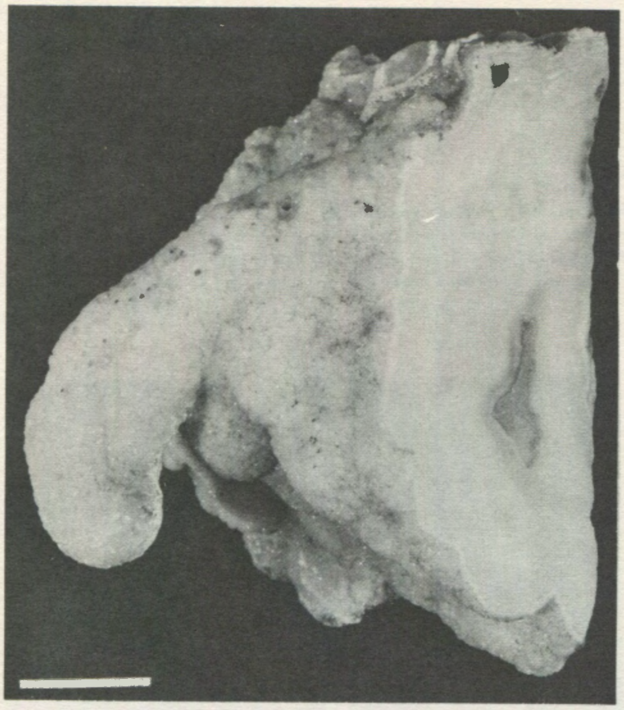


Fig. 2. A section of silica flowstone with stalactitic and botryoidal surface morphology. Scale: 2 cm.

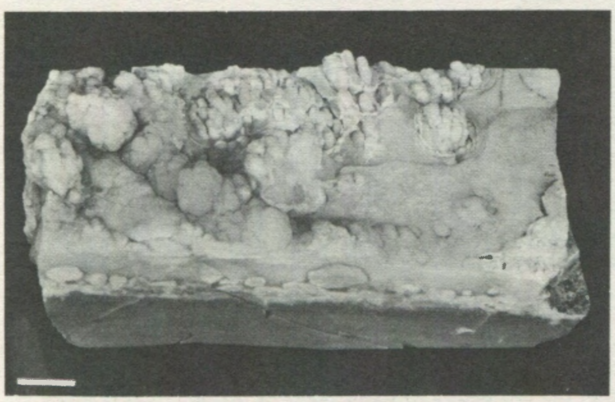


Fig. 3. Orthoquartzite clastic cave sediment. A silica permeation replaces the original carbonate framework cement. The botryoidal stalagmitic deposits on the surface strongly suggest that at least some silica deposition took place beneath ponded water. Scale: 2 cm.

terpreted as evidence of pH fluctuation. Less than 3 per cent of a given speleothem consists of carbonate growth lines. The replacement of silica after carbonate, or the coprecipitation of silica and carbonate, sometimes is cryptocrystalline, thus producing a diffuse Alizarin-red staining effect. Irregular contacts along calcite-chalcedony growth lines occur. Within the crystalline calcite aggregates enveloped by silica are very fine spherulites of chalcedony replacing the carbonate. The spherulites average 0.07 mm in diameter and are associated with diffuse silica swirls in the carbonate framework.

The stalactites occur as blunted conical and cylindrical forms. No central canal was found in any specimen. Some are as long as 40 cm, but they average about 10 cm long and 2 cm in diameter. Clusters of closely grouped individuals are common (Fig. 1). Approximately 500 stalactites were observed.

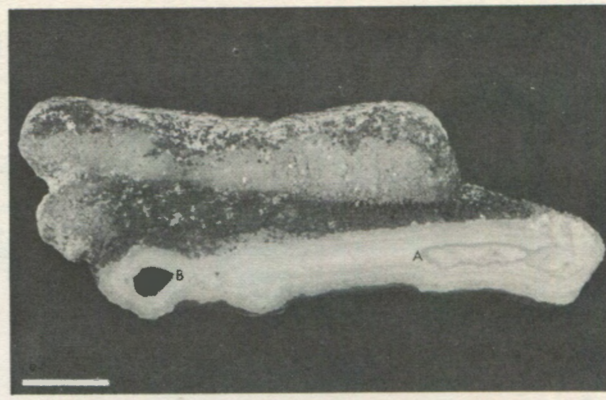


Fig. 4. A lengthwise slice of a silica stalactite, revealing carbonate growth lines (a) and a calcite crystalline aggregate (b). The speleothem surface is coated with euhedral quartz crystals 1 to 3 mm in diameter. Scale: 2 cm.

Petrographic study of thin sections (Fig. 5) of the stalactites reveals that they consist chiefly of radially-oriented, length-fast chalcedony fibers. During the last stage of silica precipitation, euhedral quartz crystals up to 25 mm in diameter developed on the surfaces of many of the stalactites. Examination of the stalactitic silica by Scanning Electron Microscope (SEM) techniques reveals its microstructure to be an aggregate of crystalline plates 5 to 20 μ in diameter, as well as larger, euhedral quartz crystals (Fig.

Fig. 5. Photomicrograph of a silica stalactite. Note the optically fibrous radial structure under crossed nicols.

6). The silica, being length-fast, was deposited under low pH conditions (see Folk and Pittman, 1971) and is interpreted to be a composite of spiral chains of silica tetrahedra in an interlaced fibrillar chain habit. The *c* axes lie tangentially to the surface and perpendicularly to the direction of growth of the fibers. This would explain the discrepancy between the fibrous appearance of the silica under optical examination and the lack of such under SEM examination (see Folk and Weaver, 1952; Folk and Pittman, 1971; Monroe, 1964). X-ray diffraction examination indicates that the polymorph is alpha-quartz.

Most of the stalagmitic deposits are on the silica flowstone, but some are isolated. The largest are about 3 cm in diameter and 1 cm high. They are assumed to be formed in the classical manner, by deposition from dripping water.

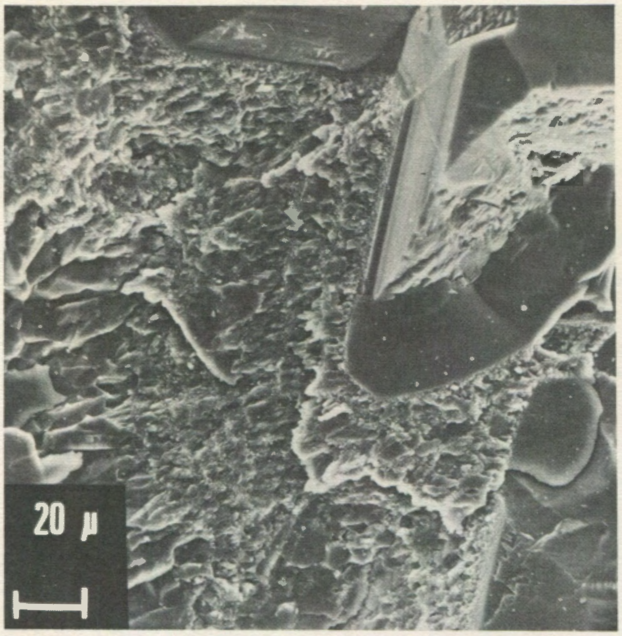


Fig. 6. Scanning electron micrograph of crystalline silica in a silica stalactite.

The botryoidal morphology of the silica encrusting the clastic cave sediments (Fig. 3) strongly suggests that some deposition took place beneath ponded water.

Conclusions

Stalactitic silica is an extremely rare secondary feature of solutional limestone-dolomite caverns. Overall chemical processes in the subsurface karst are relatively poorly understood (Broughton, 1972). This discovery provides a new insight into the variations of morphology and chemistry of mineral deposits in karst terraines.

Acknowledgements

The author thanks James F. Quinlan (Mammoth Cave National Park), William B. White (The Pennsylvania State University), and Dwight E. Deal (University of Texas) for critically reviewing early drafts of this manuscript and offering valuable suggestions; the Saskatchewan Department of Mineral Resources and the Department of Entomology of the University of Alberta for the use of their research facilities.

References Cited

Anderson, C. A. (1930)—Opal Stalactites and Stalagmites from a Lava Tube in Northern California: *Am. Jour. Sci.,* **5th ser. 20**:20-26.

Broughton, P. L. (1971)—Ubersicht der in Tropfsteinen Bekanntgewordenen Minerale: *Die Höhle* **22**:81-82.

——— (1972)—Secondary Mineralization in the Cavern Environment: *Studies Speleol.* **2**:191-207.

Deal, Dwight (1964)—Scintillites, a Variety of Quartz Speleothems: *NSS Bull.* **26**:29-31.

Dings, M. G. and D. H. Whitebread (1965)—Geology and Ore Deposits of the Metaline Zinc-Lead District, Pend Oreille County, Washington: *U. S. Geol. Survey, Prof. Paper* 489, 109 pp.

Folk, R. and J. Pittman (1971)—Length-slow Chalcedony: a New Testament for Vanished Evaporites: *Jour. Sed. Petrol.* **41**:1045-1058.

——— and C. E. Weaver (1952)—A study of the Texture and Composition of Chert: *Am. Jour. Sci.* **250**:498-510.

Lassak, E. (1970)—A Note on Some Non-Calcareous Stalactites from the Sandstones of the Sydney Basin, N.S.W.: *Royal Soc. N.S.W., Jour. Proc.* **103**:11-14.

Mallory, W. (1972)—Regional Synthesis of the Pennsylvanian System, IN: Rocky Mountain Association of Geologists, Geologic Atlas of the Rocky Mountain Region: Denver, Rocky Mt. Ass'n Geologists, pp. 111-127.

Monroe, E. A. (1964)—Electron Optical Observations of Fine-Grained Silica Minerals: *Amer. Mineralogist* **49**: 339-347.

Peck, S. (1962)—Lava Caves of Craters of the Moon National Monument: (Iowa Grotto, NSS) *Iowa Cave Book* **6**:B33-B35.

Quinlan, J. F. (1972)—Karst-Related Mineral Deposits and Possible Criteria for the Recognition of Paleokarsts: A Review of Reservable Characteristics of Holocene and Older Karst Terranes: Internat. Geol. Congr., 24th, Proc., Sec. 6, pp. 156-168.

Siegal, F. R.; J. P. Mills, and J. W. Pierce (1968)—Aspectos Petrograficos y Geoquimicos de Espeleotemas de Opalo y Calcita de la Cueva de la Brujo, Mendoza, Republica Argentina: *Ass'n Geol. Argentina, Rev.* **23**:5-19.

Swartzlow, C. R. and W. Keller (1937)—Coralloidal Opal: *Jour. Geol.* **45**:5-19.

White, W. B. and G. H. Deike (1962)—Secondary Mineralization in Wind Cave, South Dakota: *NSS Bull.* **24**:74-87.

Manuscript received by the Editor 20 March 1974
Revised manuscript accepted 13 April 1974

Note added in proof: The caves herein described are now under legal jurisdiction of mining claim law and, consequently, the "agate" speleothems are being extracted at an alarming rate by mineral and gem collectors.

Environmental Influences of the Glacieres of the Pryor Mountains, Montana

William B. Vincent *

ABSTRACT

Detailed examination of the glacières in the Pryor Mountains, Montana revealed several macroenvironmental factors which contribute to the freezing microclimates of the caves. The stored-up "cold" of winter and the structure of the cave are traditionally invoked as causes of subterranean ice and are of primary importance. Secondary factors, such as the location of the cave, the vegetation zone in which the cave is found, its associated plant cover, and the availability of moisture, also are important.

Introduction

Although a considerable literature has been produced describing the basic dynamics of ice formation in glacières,[1] comparatively little has been written relating these to the over-all external environment within which the caves exist. This paper relates the freezing microclimates of glacières to their over-all external environment, utilizing the glacières of the Pryor Mountains as examples.

The Pryor Mountains are in southern Montana, 30 miles south of Billings. They are a small, isolated range. The larger Big Horns lie to the east and the massive Beartooth lie to the west. Although the Pryor Mountains appear small in relation to their larger neighbors, they rise to an altitude of 8,776 ft and contain many areas as yet little touched by man.

Several of the glacières in the Pryor Mountains are considered by Halliday (1954) to be major limestone glacières. However, very little has been written about them. Brief descriptions are given by Aaberg (1954) and Halliday (1970). The geology of the caves is discussed by Elliott (1963). The lack of detailed information on these caves makes it difficult for one to judge the extent of their ice deposits in the past and to determine what are the dynamics of ice formation at the present.

The microclimates of glacières do not exist as isolated entities. They, like other microclimates, are related to the overall environment. In order to provide an understanding of freezing microclimates underground and to relate these to the macroenvironment of the Pryor Mountains, the first section of this article briefly reviews theories on the formation of ice underground and discusses the important environmental factors involved. The next section discusses the geology, physical setting, and climatology of the Pryor Mountain area.

Section three describes the glacières in the Pryors. Big Ice Cave, Crater Ice Cave, Little Ice Cave, and Red Pryor Ice Cave are included. Several minor neigières found in the area are discussed, also. The mechanism of ice formation and the effects of the local environment and the physical structure of the caves on the formation of ice are analyzed.

In the final section of this paper, the structure of the caves, vegetation zones, climatic variations, and other factors are compared. Conclusions are then drawn as to the importance of these environmental influences on the maintenance of freezing microclimates in the glacières.

Dynamics of Ice Formation Underground

Edwin Swift Balch published "Glacières or Freezing Caverns" in 1900. In this work, a classic of American speleology, he spelled out the factors which cause the formation and preservation of ice underground and pointed out the links between the general environment of a region with glacières and the existence of ice in caves. He also discussed the physical shape of the caverns and the movement of air and water into them. In the following paragraphs, those ideas which pertain to the glacières of the Pryor Mountains will be given a closer look.

The most important link between the surface environment and freezing microclimates in caves, according to Balch, is the "winters cold." Simply stated, this theory asserts that it is due to the cold of the winter that ice is found in some caves. Freezing caverns will be found only in regions where the temperature falls below freezing for at least part of the year. Evidence for this includes the fact that glacières are found only in temperate regions, or in areas at high elevations with temperate zone characteristics. Balch presents a list of snow and ice deposits, ranging over a spectrum from snow drifts to glacières, and concludes that deposits of ice in caves are but the longest-lingering traces of winter.

The ways in which the cold of winter penetrates caves and is preserved are various. An important factor is the relatively great density of cold air. Because of this, cold air flows down slopes and collects in low places; caves are no exception, provided that the structure of the cave is such to allow this. The method by which air enters and leaves caverns is the basis of one of Balch's classifications of glacières.

Dynamic glacières, as described by Balch, operate in the following way: Air flow is induced by density changes occurring between two entrances, one higher than the other. When the temperature of the cave is colder than that of the outside air, dense, cool air from the cave will flow out of the lower entrance. This creates a partial vacuum at the upper entrance, which induces warmer, outside air to flow into it,

* Research Associate, Rocky Mountain College, Billings, Montana 59102.

[1] The term "glacière" has been used by Balch, Halliday, and others to denote caves containing ice formed underground. The use of the term "ice cave" to describe such caves is common. "Ice cave" is properly used to describe a cave in a glacier. In this paper, "glacière" will be used most of the time; where the term "ice cave" is found, it will denote one of the glacières in the Pryor mountains.

warming the rock. As the air enters the cave, it is cooled by the surrounding rock and flows to the lower portions of the cave. In the winter, when the outside air is cooler than the air in the cave, the flow is reversed. Warmer air flows out of the top entrance, causing cold air to be drawn into the lower entrance, cooling the rock.

Static glacières are caves with downward-sloping passages containing pockets of cold, "dead" air. Active air circulation occurs in these passages only in the winter, thus making the air in them appear to be stagnant when observed in the summer. Balch (1900, p. 122) describes the movement of air into them in the following way: ". . . the air in summer is nearly still, while in winter there are distinct rotary movements of the air as soon as the temperature outside is lower than that within." The movement of air into any glacière by gravity flow is, of course, controlled by the structure of the cave.

Balch uses cave structure as the basis for another classification of glacières. According to this classification, there are pit caves and cliff caves. Pit caves are those with vertical entrances. In these, the ice usually is found under the entrance, as in the bottom of a sinkhole. The second type of cave, the cliff cave, has an entrance at the base of a cliff or in the side of a hill. In this type of cave, if it is to be a glacière, the floor must slope downward, usually from the entrance. If the floor does rise initially, it never becomes higher than the top of the entrance. The utility of a classification of freezing caves based on structure is, of course, problematical. Balch (p. 119) states ". . . the lines of transition between" forms of caves, "however, are so indefinite in nature, that it is often difficult to specify a cavern as belonging to any special type."

In addition to the winter's cold, the greater density of cold air, and the structure of the cavern, many other factors influence the development of ice underground. Some of these influences, such as elevation, amount of moisture, and the season at which moisture enters the cavern, were noted by Balch. He gave little attention to specifics, however, saying (p. 144): "They all follow the same general laws as to the origin of their contents, modified only in slight degree according to the varying natural local conditions . . ." Elucidation of their influence was left, to be completed later.

Other theories concerning the formation of ice in caves have been put forth. Many of these, in one form or another, were thoroughly discussed by Balch (1900) and found wanting. A few of the mechanisms may have some influence on ice formation, but none are major factors.

Halliday (1954) presents a summary of the theories on ice formation underground, including that proposed by Balch. In Halliday's paper are discussions of environmental factors which Balch discussed superficially or not at all. Moisture, barometric pressure changes, evaporative cooling, and other facets of speleometeorology are highly influential. They, to a large extent, determine the morphology of the ice deposits found in glacières.

Speleometeorology traditionally views caverns as homeothermic systems. Variations in the system are brought about by outside influences. The major external influences on glacières are discussed in this paper. Those interested in more information on spelean meteorology in general are referred to Montorial Pous (1951), Trombe (1949), Cropley (1965), Conn (1966), and Myers (1953).

Moisture entering a cave will either raise the temperature of the cave or lower it, depending on the type of cave and the form in which the moisture is introduced. A cave in which the temperature is above freezing will be cooled greatly by an entering stream in the winter and will be heated by it in summer (Cropley, 1965). Halliday (1954) notes that water entering a cave will act as a temperature stabilizing agent, if that water is in temperature equilibrium with the bedrock. This would seem to be the exception rather than the rule. Water entering caves from the surface usually is not in temperature equilibrium (Cropley, 1965). In a freezing cave, entering water can only raise the temperature because, as it freezes, it releases a large amount of heat. The introduction of snow into a cave produces the opposite effect, aiding in the cooling of the cavern (Halliday, 1954). Often, this is a major factor in pit glacières and other caves, the structure of which allows massive amounts of snow to enter in the winter.

Air flow into caverns is a major process in speleometeorology. Spelean air flow may be caused by several agents, such as variation of barometric pressure, thermocirculation, and entraining of air by water flow in narrow cavities (Trombe, 1949).

Air flow in glacières usually is caused by barometric pressure changes and by thermocirculation. Barometric changes cause an imbalance between the surface and underground air pressures. The time which it takes for the pressure to reach equilibrium, and the volume of air exchanged, depends on the size and shape of the cavern (Conn, 1966). Barometric air flow in glacières is detrimental to ice deposits when the air temperature is above freezing.

Thermocirculation is thought to be the major agent of heat exchange in glacières. In dynamic glacières, the "chimney effect" is responsible for the often noticeable draughts. In static glacières, the "rotary" pattern of air flow is not so easily observable and is seasonal in its occurrence.

Environmental factors such as plant cover play a little-known rôle in the formation of underground ice. While others have acknowledged its influence, they have presented little data on it. The influence of the foregoing environmental factors will be evaluated in discussing each of the glacières of the Pryor Mountains.

Geology, Climate, and Vegetation Zones of the Pryor Mountains

Geology

The Pryor mountains, one of the encircling ranges of the Big Horn Basin (Fig. 1), were formed initially by the Larimide revolution at the end of the Cretaceous (Mackin, 1937). These uplifts appeared as low hills (Elliott, 1963), compared to the height of the present day mountains. Extensive Tertiary sediments were deposited at the base of these hills. Sedimentation continued until the Big Horn Basin became filled with sediments to the level of the top of the Pryor Mountains. Regional uplift occurred during the late Pliocene, causing removal of less resistant sediments and formation of the present-day mountains (Schulte, 1962). Erosion still continues. Madison limestone (Mississippian) is the formation most widely exposed in the Pryor Mountains.

Erosion during the Pliocene and later epochs also is responsible for the re-excavation and enlargement of the older caverns of the area (Elliott, 1963). This is a continuing process, as evidenced by the underground drainage systems found in the mountains. Secondary mineralization has taken

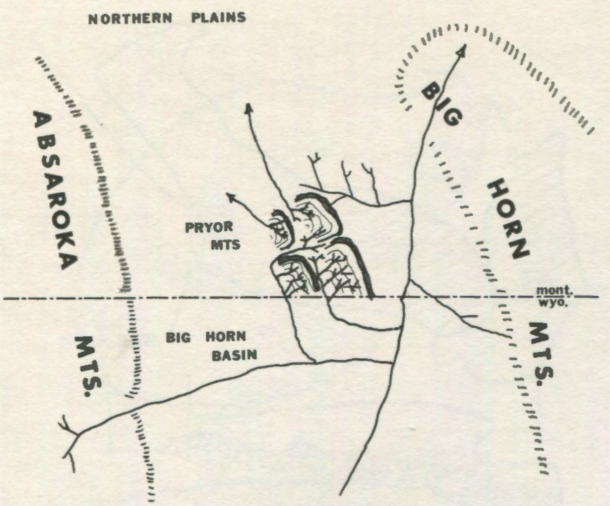
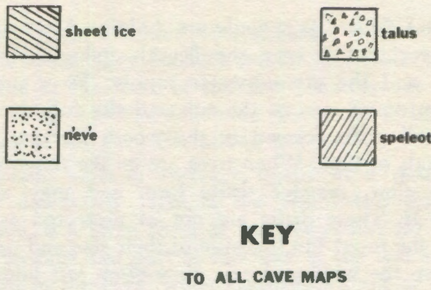


Fig. 1. Pryor Mountains Region.



Key to all Cave Maps.

place in some of the caverns, but it is not common in the glacières.

Climate

Climatic factors are among the most important influences which must be considered in a study of glacières. Because the Pryors form one of the boundary ranges between the Big Horn basin and the northern Great Plains, they exhibit a transitional climate. The summary for the Plains north of the Pryor Mountains is based on the station at Billings, Montana. The summary for the Big Horn basin is based on the station in Lovell, Wyoming.

The northern Great Plains in Montana are semi-arid. Annual precipitation (1935-1968) averages 13.88 inches. Winters are cold, though often interrupted by mild periods; the month with the lowest mean temperature (23.2°) is January. Summers are warm, days being hot but nights cool; July is the month with the highest average temperature (73.0°). Most precipitation comes as rain during spring (April-June) and fall (September-October). Snow accounts for only 20% of the yearly precipitation (U. S. Dept. Comm., 1968).

The Big Horn Basin is dry in comparison with the Great Plains. Portions of its northern end receive only six inches of precipitation yearly. The lack of moisture in the basin is due to its low elevation in relation to that of the Pryors and of the other mountains around it. Yearly average precipitation is 6.52 inches at the Lovell station. The month with the lowest average temperature is January, which has (1911-1941) a mean of 16°F. (U. S. Dept. Ag. 1941). The mean for July during the same period was 71.8°F. (U. S. Dept. Ag., 1941).

The climate of the Pryors does not, of course, simply correspond with that of the Great Plains or the Big Horn Basin. While the southern slopes of the Pryor Mountains have a climate similar to that of the Big Horn Basin and the lower northern slopes have one corresponding to the Plains climate, climate in the higher areas of the mountains is modified by increasing elevation. As one travels up the slopes of the mountains, the temperature drops noticeably and the amount of moisture increases. The normal lapse rate is about 3°F per 1,000 feet in the Pryor Mountains. It should be noted, however, that temperatures in winter often are not as low at high altitudes as on the adjacent plains, because cold, polar air masses often fail to rise as high as the summits. This may be observed in the records from Mystic Lake (Stillwater County, Montana) and Crandall Creek (Park County, Wyoming) (U. S. Dept. Ag., 1941). The 3° change per thousand feet can, however, result in low temperatures at high altitudes during the summer; an example is the hard frost of July, 1971 at Bear Canyon Guard Station, in the Pryors.

More precipitation occurs at high elevations in the Pryor Mountains than in the Big Horn Basin or the northern Great Plains. The following information is based on conditions in mountains near the Pryors (there is no weather station in the Pryors, themselves). Precipitation in the higher portions of the mountains is at least 18 inches annually. Snowfall exceeds four feet; all precipitation from September to mid-May is likely to be in the form of snow.

Vegetation Zones

The climatically controlled vegetation zones of the Pryors in turn influence the freezing microclimates of the glacières. Vegetation can affect snowdrift size, run-off volume, and wind velocity. The formation of ice underground is strongly influenced by these factors. The following zones are based on Cary (1917).

The lowest vegetation zone found in the section of the Pryor Mountains under consideration is the Transition zone. This zone is characterized by high, sage-covered plains in the northern portion of the Pryors and by slopes covered with mountain mahogany and juniper in the south, reflecting the climates of the northern Great Plains and of the Big Horn Basin, respectively. The most common plants found in this zone include: *Artemisia cana* (Silver Sagebrush), *Artemisa arbusula* (Black Sagebrush), *Juniperus utahensis* (Utah Juniper), *Pinus ponderosa* (Western Yellow Pine), and *Cercocarpus ledifolius* (Curl Leaf Mountain Mahogany). The vegetation found in this zone does not affect run off to any great extent, except by increasing the moisture-holding capability of the soil. Snow drifts do form, but fail to last long due to their exposure to the wind and sun.

The Canadian zone, found on the middle and higher mountain slopes, often appears abruptly out of the sagebrush and juniper slopes of the Transition zone. The Canadian zone is noted for its stands of evergreen trees, evidence for the increase in moisture with increasing elevation. Plant communities found in this zone are not limited to those dominated by conifers. There also are meadows, some of them quite large, in which grasses or sagebrush may be dominant. Trees found in this zone in the Pryor Mountains include: *Pinus flexilis* (Limber Pine), *Pinus Contorta* var. *latifolia* (Lodgepole pine), *Pseudotsuga taxifolia* var. *glauca* (Douglas-fir), and some members of the genus *Picea* (Spruce). Sagebrush in this zone includes *Artemisia tridentata* (Big

Sagebrush) and *Artemisia scopulorum* (Alpine Sagebrush). The plant cover in this zone significantly influences snow cover, runoff, and the ground-water supply. Trees protect the snow from direct rays of the sun and the full force of the wind. Small drifts formed in the woods last for considerable periods of time. When trees are on the down-wind side of a meadow, massive drifts form and may attain depths of 30 ft. These drifts are not as protected as are those deep in the forest but, because of their size and partial protection from the wind and sun, they often last into late spring. Openings in the forest are often wind-swept. Large drifts do not form in them unless an obstruction to the wind is present. Drifts in the open are the first to melt. Run-off from them is rapid.

The Hudsonian zone, the area of dwarfed, stunted conifers just below timberline, is found only in a few small areas of the Pryors. Here, the ground cover is composed primarily of grasses. Only a few small, stunted woody plants are present. The Hudsonian zone in the Pryors is characterised by large meadows bounded by the Canadian zone on the lower side and by wind-swept ridges with stunted spruce on the upper. These areas, while receiving massive amounts of snow, often drift clear. Large drifts, some of which last into August, do form along ridges and cliffs. Run-off from this area is rapid, but does not occur until late spring. Ground water in the caverns is affected by this run-off, receiving most of its annual recharge within a very short period of time.

Glacieres of the Pryors

Several authors have presented short descriptions of the glacières of the Pryor Mountains. None, however, have given detailed information on the ice deposits found therein. The following descriptions of the caves give attention to the types, extent, and other basic characteristics of the ice found in each cave. The caves discussed are: Big Ice Cave, Crater Ice Cave, Little Ice Cave, and Red Pryor Ice Cave. Another reported ice cave is mentioned, as are some neigières in the area.

Big Ice Cave

Big Ice Cave is on Cave Ridge, East Pryor Mountain at an elevation of 7600 ft in the Canadian zone. The top of the ridge and the canyon are thinly forested, but trees do not shelter the cave entrance. Big Ice Cave is near a good road and is the best-known cave in the area. The U. S. Forest Service placed a gate at the entrance and developed the cave for tours on weekends. Due to poor construction procedures, however, the gate has failed to stop vandalism.

The large entrance to Big Ice Cave opens at the foot of a twenty five-foot cliff. From the cliff face, a talus slope leads down to the sheet of ice which forms most of the floor in the first chamber. At the rear of this chamber, is a small pit which leads down into the second and final chamber.[2] This chamber is much smaller than the first and is set at an angle in relation to the upper chamber (see Fig. 2).

The ice deposits in Big Ice Cave are the most extensive of any in the Pryor Mountain glacières. The first and second chambers are floored by a massive sheet of ice. It is 16 ft thick in places. Most of the ice is much thinner, tapering to

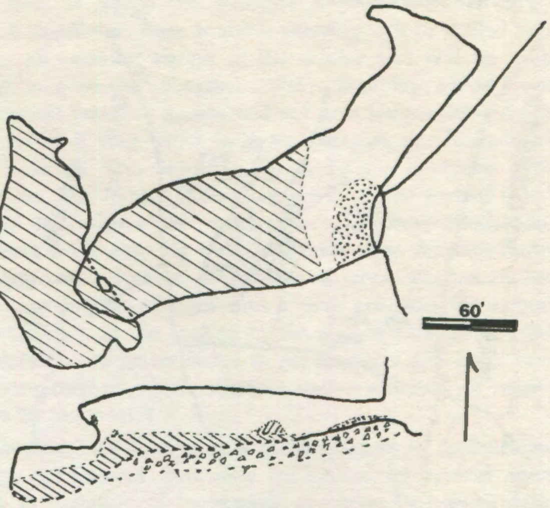


Fig. 2. Big Ice Cave.

one or two ft in thickness at the front of the cavern. Other types of ice deposits found in the cave include frost crystals and ice stalagmites. The formation of the ice deposits within the cave and their classification is as follows:

The sheet ice on the cave floor is a permanent feature. It has been said by local residents that the ice was much thicker in years past, but no quantitative measurements were made by which to establish this. One source of water for this sheet ice is snow which has drifted into the entrance. As the season progresses, the snow melts and the water flows down onto the ice floor where, at a lower level and temperature, it is frozen. The extent to which the gate influences snow entering the cave is not known. The second source of water for the sheet ice is snow melting on the land surface above the cave. This water percolates through the limestone, falls into the cave, and is frozen. Percolating melt water also is responsible for the stalagmites. These, in melting, add water to the sheet of ice.

Seasonal speleothems include frost crystals in the lower chamber and stalagmites in the upper chamber. Frost crystals are widely developed in the lower chamber. They last well into the summer but are much diminished by late August. Small stalagmites in the upper chamber are formed by dripping water, which is frozen when it encounters subfreezing air in the cave. These speleothems reach their peak of development in the spring and usually are gone by late fall.

Crater Ice Cave

Crater Ice Cave is on the main portion of Big Pryor Mountain, in the Hudsonian zone. The cave entrance is just below the rim of the escarpment at the head of Tibbs Hollow, at an elevation of 8,700 ft. While the top of the mountain at this locality is not wooded, the base of the cliff is shielded by wind-blown, stunted spruce. The cave entrance is in one of these groves.

Crater Ice Cave has two entrances, one a sink and the other a walk-in. These make a natural arch, through which the wind often blows. The cave consists of a single oval chamber, 200 by 130 ft. in diameter. From the crest of the talus beneath the sink, the floor slopes down to the rear of the chamber. (Fig. 3).

The cave contains relatively few ice deposits. However, those that it does contain are of some significance. The cave

[2] There have been persistent rumors that the ice in the lower chamber of Big Ice Cave has blocked openings leading to many other chambers. While none have been located, the possibility still must be considered.

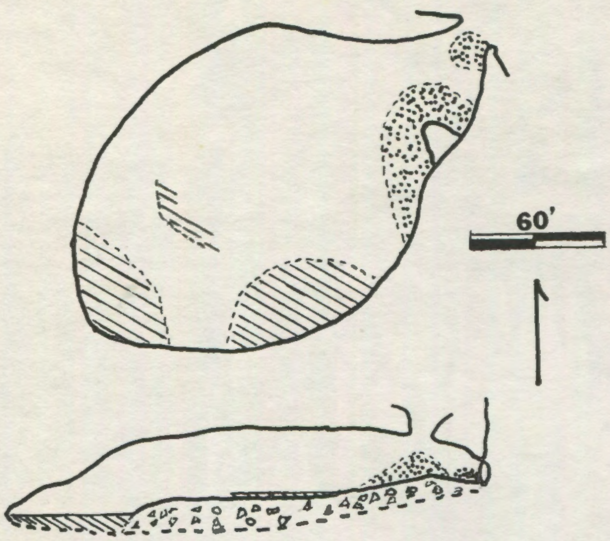


Fig. 3. Crater Ice Cave.

has been called a neigière (Halliday, 1954), but close examination reveals otherwise. The major deposit in the cave is a cone of snow which has been blown into the sink entrance. This cone lasts all summer and the water from its melting gives rise to sheet ice in several lower areas of the cave. The thicknesses of the ice sheets are not known, but they are of a permanent nature.

The pillar of ice which Schultz (1969) mentions is not a speleothem caused by drip, as it first appears to be. It is a remnant of a massive sheet of ice which once covered the floor. Remnants of this ice sheet are found under the talus which covers the floor. The total extent of the ice is not known, as it crops out in only a few areas of the cave. The presence of ice under the talus needs more study. Ice deposits in the cave, though small, may provide a key to many unanswered questions about past climatic conditions in the area. The buried ice indicates that this now minor glacière once contained extensive ice deposits.

Current ice formation is governed by several factors. The high elevation of the area, while providing some conditions favorable to ice formation, hinders it in other ways. Long, cold winters with heavy snowfall and cool summers are positive factors. The plant cover, typical of the Hudsonian zone, does not keep the snow from drifting. As a result, much of the area above the cave is blown clear. This lack of snow along with the fact that the cave lies under the crest of a ridge, limits the supply of water available to it. The major moisture source is the snowdrift in the sink entrance.

Little Ice Cave

This glacière, located near the top of Tillett Ridge on East Pryor Mountain, is the largest of the known caves in the Pryor Mountains (Schultz, 1969). It is the second-most important glacière in terms of the volume of ice contained. The entrance, located at an elevation of 8100 ft, is near the upper limit of the Canadian zone. Sheltered by trees, it is well protected.

The structure of Little Ice Cave is complicated when compared to that of the other caves in the Pryor Mountains. The cave consists of small passages on three levels, all developed within a limited vertical range. Neither the passages nor the rooms they connect contain ice. Only in the entryway and at the front of the first chamber are ice deposits found (Fig. 4). The following discussion is limited to the portions of the cave containing ice.

A small, walk-in entrance opens to a passage sloping gently downward. Here, a few feet below the lip of the entrance, is a large sheet of ice. Ice covers the floor of the entry passage and part of the floor in the first chamber. Ice deposits in the cave, while massive, are not as large as are those in Big Ice Cave. The sheet of ice which covers most of the floor is only four feet thick at its greatest exposure. Its average depth is much less. The sheet ice is permanent and does not recede to any great extent from year to year.

The floor of the chamber beyond the entrance passage also is covered with ice, in front. The ice here is much thinner than is that found in the entry passage. It is in this area that ice stalagmites and other seasonal ice deposits reach their greatest development.

The ice in Little Ice Cave has three different sources. The sheet ice which covers the floor is derived from snow blown into the entrance. Melt water from this snow flows down into colder areas and freezes. Only a small portion of the sheet ice is formed from water coming from other sources.

Percolating ground water finds its way into the cavern through small fractures. Where these are present, many ice

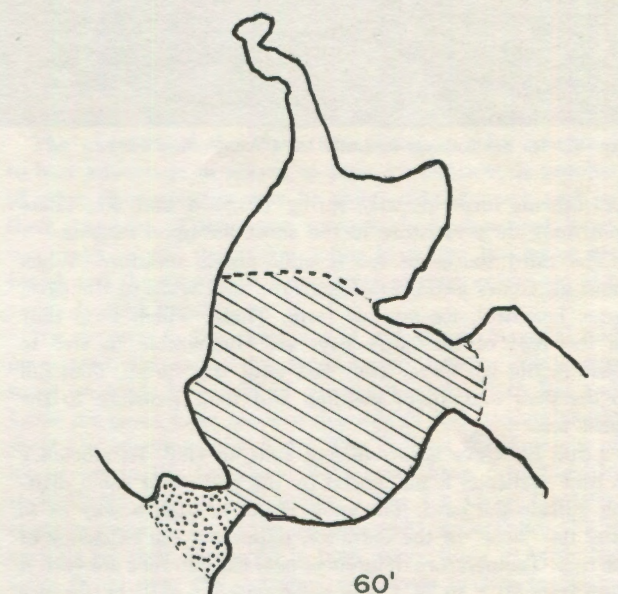


Fig. 4. Little Ice Cave.

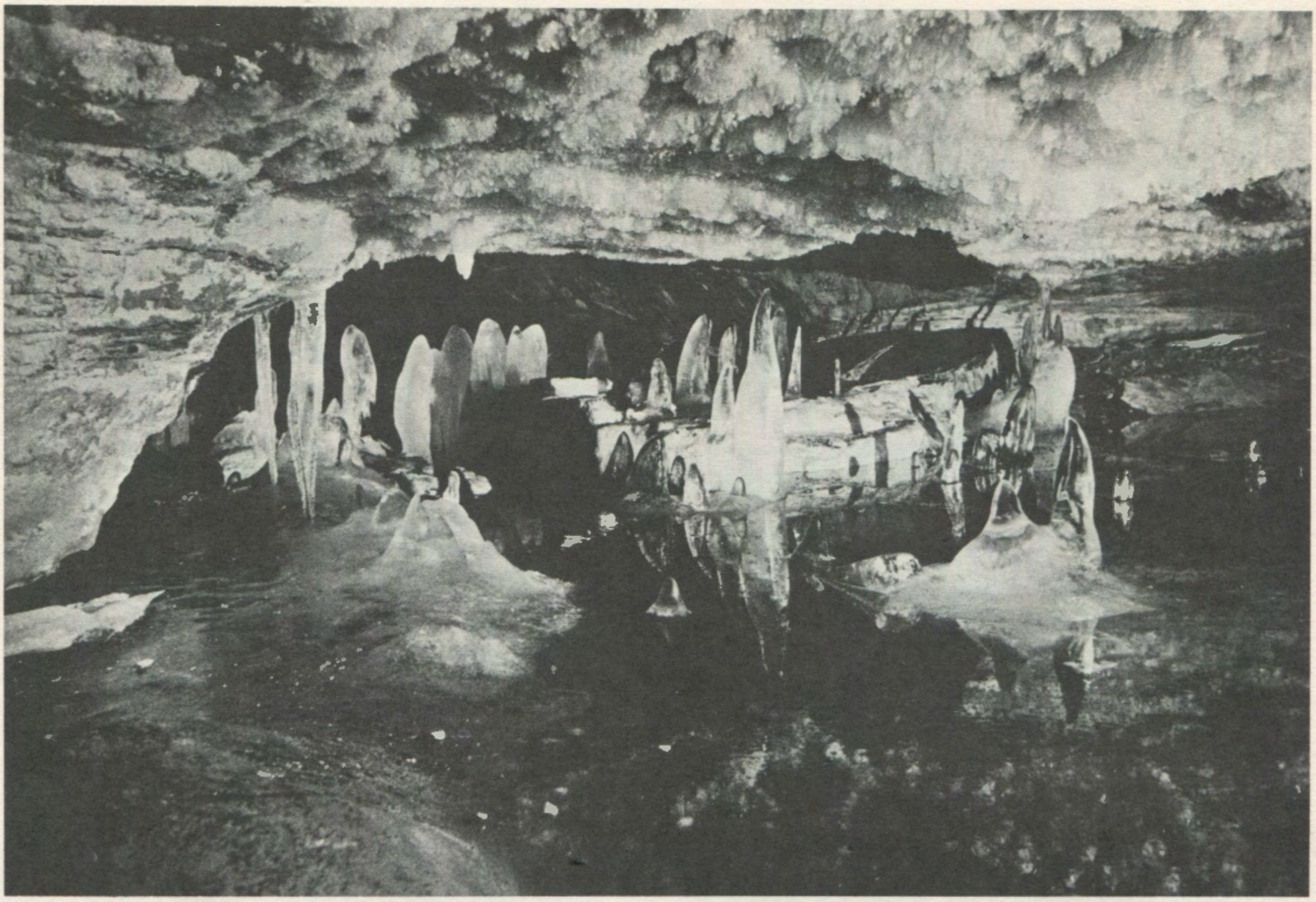


Fig. 5. Ice Speleothems in Little Ice Cave, Spring, 1971.

speleothems form in early spring (Figs. 4 and 5). These contribute their moisture to the sheet ice upon melting.

The third source of ice is atmospheric moisture. When moist air comes into contact with the cold walls of the cave, large, beautiful, ice crystals form. These, which form only on the roof of the entry-way, are comparable in size to those in Big Ice Cave. They last until late spring, then fall to the floor and, upon melting, add their moisture to the sheet ice.

Little Ice Cave is an efficient cold air trap. Its efficiency in heat exchange is influenced by the pattern of snow drifting outside the cave. The snow drifts in such a way as to raise the "floor" of the entrance, improving the efficiency of the trap. Temperature differences near the entrance are vast. A drop from 80°F to 32°F was noted within a vertical distance of 10 feet in the summer of 1971. The existence of passages beyond the ice deposits raises many questions. Barometric air movements influenced by them has an unknown effect on the ice deposits. A detailed study will be required to determine the extent of their influence on the ice deposits.

Red Pryor Ice Cave

Red Pryor Ice Cave is on the southern end of Crooked Creek Ridge (Red Pryor Mountain). This cave occurs in the Canadian zone, at an elevation of 7900 ft. The ridge in the vicinity of the cave is barren of trees, except for a few pines along the edges. The cave is entered through a sink about 40 ft deep.

The structure of the cave is that of a simple pit glacière. Two sinks open into a large, oval room 200 by 300 ft in diameter. The larger of the sinks is 15 by 20 ft wide at the surface. This sink, which is considered to be "the entrance", has a large talus pile at the bottom. This pile covers the entire floor of the cave, except for small areas of mud and ice (Fig. 6).

Ice deposits in Red Pryor Ice Cave qualify it as a glacière. A large cone of névé is found under the entrance sink. It and a smaller one under the second sink are replenished annually by snow. In early spring, the larger cone extends to the surface, making it possible to walk into the cave. As the season wears on, the cones decrease in size. The larger cone has a base of permanent ice, but the smaller one is a seasonal phenomenon.

The cones, in melting, provide the moisture from which most of the true cave ice is formed. Water from their melting flows down the talus slope and freezes at successively lower elevations. Although ice did not cover all of the talus in the spring of 1971, it was as much as two feet thick in some areas. It appears that much of this ice is temporary in nature, melting and forming anew each season. A flat sheet of ice occurs at the base of the talus slope. This sheet ice had a maximum thickness of about four feet in the spring of 1971. In the fall, it was covered by water.

Ice speleothems are found in some areas of the cave, but these are small and of short duration. Small frost crystals are found early in the year, but do not reach a high level of development. Thermostratification affects speleothems in the cave, causing the higher ones to melt first.

The primary source of moisture in the cave is the snow which drifts into the larger sink. The other sink is so small

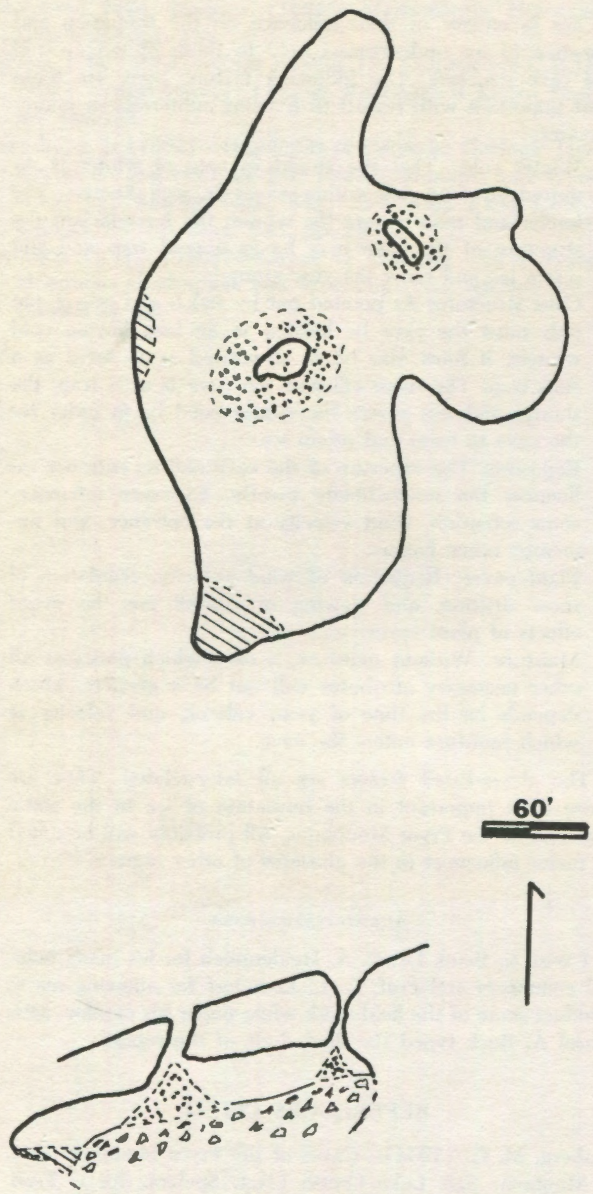


Fig. 6. Red Pryor Ice Cave.

that little snow is blown into it. The exposure and elevation of the ridge beneath which the cave is found and the lack of ground cover larger than sage brush prevents the formation of massive drifts above the cave. The amount of water which finds its way into the cave from the surface is, therefore, quite limited. As the two entrances are at the same elevation, there is no chimney effect.

Blackie Ice Cave

This cave, listed in "Caves of the Big Horn and Pryor Mountains of Montana" (Schultz, 1969), has not been relocated as of spring, 1971. An extensive search of old files produced a map, dated in the 1930's, which includes the location of a "Blackie Ice Cave." The mapped location and the written description in the file for this cave match those of Red Pryor Ice Cave. The cave listed as "Blackie Ice Cave" by Elliott (1963) is a neigière in the area of Crater Ice Cave.

Nameless Pryor Cave

This cave, found in 1968 by a member of the Treasure State Speleological Society, is near Little Ice Cave. It has a sink entrance and ice was seen in the bottom of the sink. Due to the lack of time and equipment, this cave was not entered at the time of its discovery. It has not been found since.

Neigières

The top of Big Pryor Mountain is peppered with sinks, many of which provide suitable conditions for the retention of snow during most of the year. Many of the sinks are borderline neigières. Others, such as the one called Blackie Ice Cave by Elliott (1963) and the one described in "A New Pit, Pryor Mountains, Montana" (Vincent, 1971) contain considerable snow the year around. The latter sink is interesting in that, during the spring run-off, it takes the entire flow of the stream with little effect on the snow in another portion of the sink.

East Pryor Mountain has sinks, but none are known at this time to contain snow all year in normal years. Many of them do have snow late in the fall, but it is gone before the first new snow flies.

All of the neigières are found at high elevations. Here, over 7500 ft above sea level and north of the 45th parallel, winters are long and cold while summers are short and cool. The sinks classified as neigières are found in the Canadian and Hudsonian zones. Most of these are in the open and snow drifts into them freely. It is no surprise that many of the sinks found in these zones are neigières.

Comparisons

The glacières of the Pryor Mountains may be compared to best advantage in regard to their ice deposits. A number of other characteristics may be compared, however, such as their structures. Comparison of variable environmental factors, such as the vegetation zones in which the caves are found and the density of plant cover at their entrances, also yields fruitful insights.

Ice deposits in the caves are of two types: Seasonal and permanent. Permanent ice tends to be massive, while seasonal ice tends to consist of smaller deposits. Both Big and Little ice caves have massive deposits of ice, predominantly sheet ice, covering their floors. The greatest development of ice occurs in Big Ice Cave. Both caves have dripstone-like speleothems such as stalagmites. Little Ice Cave, however, contains more by far of the dripstone-like speleothems. Frost crystals are well developed in both caves, those in Little Ice Cave being near the entrance and those in Big Ice Cave being found in the lower chamber. Both caves, at some time of year, do have snow in their entrances, but not to the extent that Red Pryor and Crater ice caves do.

Crater and Red Pryor ice caves contain less ice than do Big and Little ice caves. The largest ice deposits found in Crater and Red Pryor ice caves resemble those in neigières. These caves do contain ice deposits formed underground, however, thus permitting them to be classified as true glacières. The snow cones in both caves give rise to permanent, though small, sheets of ice lower down. As described previously, the other ice speleothems in these caves are quite small and are of a seasonal nature.

The glacières in the Pryors are of two structural types: Cliff glacières and pit glacières. Both Big and Little ice caves fall into the category of cliff glacières. These caves open at the base of a cliff, or on the side of a hill, and slope down into the hill. The slope into Big Ice Cave is steeper than is the one into Little Ice Cave, but this does not seem to be much advantage in developing and preserving ice.

Red Pryor Ice Cave is a pit glacière. The two sink entrances to this cave provide a suitable structure for the collection of cold winter air. The structure of Crater Ice Cave is intermediate between that of a pit and that of a cliff glacière. It has a sinkhole entrance, into which snow is blown in the manner of pit glacières; it also has a walk-in entrance at the base of a cliff. The structure of the cave, while suitable for the preservation of ice for a long period of time, is robbed of some of its efficiency by wind passing between the two entrances. The cave would be more efficient as a cold trap if only one entrance were present.

All glacières and neigières in the Pryor Mountains are at high elevations, where winters are long and cold. There, most precipitation occurs as snow, which may fall almost any month of the year. It should be noted that none of the caves developed at low elevations in the Pryors contain ice or snow except in the winter.

The only major climatic difference among the locations of the ice caves seems to be the relatively high wind velocities at Crater and Red Pryor ice caves. The glacières in the Pryor Mountains are all located within either the Canadian zone or the Hudsonian zone, the two highest vegetation zones on the mountains. In the lower and more widespread of the two (the Canadian zone), are Little Ice Cave, Big Ice Cave, and Red Pryor Ice Cave. The higher (Hudsonian) zone, which occupies only a small area, contains Crater Ice Cave.

Plant cover plays a major role in the formation of ice underground. Larger plants, such as trees, in the Canadian zone protect caves from the direct effects of high winds. Winds of up to 30 mph have been recorded in the clearing near Little Ice Cave at the same time that winds of only two or three mph were present in the forest at the cave entrance. Big Ice Cave, while not immediately protected by trees, receives some benefit from the sheltering forest. Red Pryor Ice Cave and Crater Ice Cave, being located on high, bare ridges, receive the full blast of the winds.

It is by retention of snow and regulation of run-off that plant cover plays its most important rôle in influencing the microclimates of the caves. Big and Little ice caves are so situated that massive snow drifts form above them. Water from these slowly melting drifts enters the caves over a period of time and builds up layers of ice, whereas a more rapid influx would run off without freezing. Red Pryor Ice Cave, situated on a treeless, wind-swept ridge, does not have this advantage. The area above the cave often is blown clear of snow and, consequently, the only moisture in the cave is from snow which drifts into the entrance. Crater Ice Cave receives little benefit from the massive amount of snow that accumulates in the trees just outside the walk-in entrance. The trees, hence the drifts, being down slope from the cave, the melt water flows away from the cave. Most of the moisture in Crater Ice Cave is derived from snow which drifts into the sink entrance.

Conclusions

This study has shown that ice deposits in the glacières of the Pryor Mountains are influenced by macroenvironmental characteristics as well as by cave structures. The factors of climate, exposure, plant cover, and amount and rate of run-off are secondary in their influence on the formation and retention of ice underground only to those of winter cold and cave structure. The following factors, then, are those most important with regard to freezing subterranean microclimates:

1. Winter cold: That the stored-up cold of winter is required to form ice within caves is well known. The longer and more severe the winter, the less efficient the structure of the cave may be as a cold trap and still retain ice and snow the year around.
2. Cave structure: As pointed out by Balch and others, not only must the cave be located in an area having cold winters, it must also be so structured as to serve as a cold trap. The more efficient the cave is as a trap, the shorter and less severe the winter need be in order for the cave to form and retain ice.
3. Exposure: The exposure of the cave and its entrance influences the microclimate greatly. Exposure influences snow retention, wind velocity at the entrance, and numerous other factors.
4. Plant cover: Reduction of wind velocity, regulation of snow drifting, and slowing of run-off are the major effects of plant cover.
5. Moisture: Without moisture, a cave which possesses all other necessary attributes will not be a glacière. Much depends on the time of year, volume, and velocity at which moisture enters the cave.

The above-listed factors are all inter-related. They are those most important in the formation of ice in the static glacières of the Pryor Mountains. All probably will be noted as major influences in the glacières of other regions.

Acknowledgments

I wish to thank Dr. C. A. Heidenreich for his many helpful comments and Prof. L. L. Loendorf for allowing me to conduct some of the field work while under his employ. Miss Carol A. Buck typed the final draft of the report.

REFERENCES CITED

Aaberg, M. C. (1954)—Caves of the Pryor Mountain Area, Montana: Salt Lake Grotto (Nat. Speleol. Soc.), *Tech. Note* 20, 2 pp.

Balch, E. S. (1900)—Glacières or Freezing Caverns: Philadelphia, Allen Lane & Scott, 337 pp.

Cary, Merritt (1917)—Life Zone Investigations in Wyoming: Biol. Survey Bur. *North American Fauna* No. 42, 95 pp.

Conn, H. W. (1966)—Barometric Wind in Wind and Jewel Caves, South Dakota: *NSS Bulletin* 28:55-69.

Cropley, J. B. (1965)—Influence of Surface Conditions on Temperatures in Large Cave Systems: *NSS Bulletin* 27:1-10.

Elliott, J. K. (1963)—Cave Occurrences in the Mississippian Madison Limestone of the Pryor Mountains, Montana: Billings, Billings Geol. Soc., 13 pp.

Halliday, W. R. (1954)—Ice Caves of the United States: *NSS Bulletin* 16:3-28.

——— (1970)—Introduction to Johnson Reprint Corp. Edition of Balch's "Glacières or Freezing Caverns": NYC, Johnson, xxxiii pp.

Mackin, J. H. (1937)—The Erosional History of the Big Horn Basin, Wyoming: *Geol. Soc. America, Bull.* 48:813-893.

Montorial Pous, Joaquin (1951)—Meteorologia Hipogea: *Urania* 238:225-246.

Myers, J. O. (1953)—Cave Physics, IN: Cullingford, C.H.D. (Ed.), British Caving: London, Routledge & Kegan Paul, pp. 144-169.

Schulte, J. J. (1962)—Highlights of Cenozoic Geologic History of the Big Horn Canyon Region, Montana and Wyoming, IN: Billings Geol. Soc., 12th Ann. Field Conf.: Billings, Billings Geol. Soc., pp. 13-24.

Schultz, R. L. (1969)—Caves of the Big Horn—Pryor Mountains of Montana and Wyoming: Nat. Speleol. Soc., Guidebook 10.

Trombe, Felix (1949)—Gouffres et Cavernes du Haut-Comminges (Paris) (extracts translated by M. H. Chantry): *Cave Research Group Great Britain, Trans.* 1(2):9.

U. S. Dept. Agriculture (1941)—Climate and Man: Washington, D. C., Government Printing Office, 1248 pp.

U. S. Dept. Commerce, Environmental Data Service (1968)—Local Climatological Data, Annual Summary with Comparative Data, 1968, Billings, Montana: Washington, D. C., Government Printing Office, 4 pp.

Vincent, W. B. (1971)—A New Pit, Pryor Mountains, Montana: *NSS NEWS* 29:109-110.

ERRATA

Minimum Diameter Stalactites, *Bull. Nat. Speleo. Soc.*, 1972, 34(4):129-136.

p. 131, Equation (4). Derivative should be in brackets, viz: $\left(\frac{\partial \hat{\theta}}{\partial \nu}\right)_{\theta-\theta_m}$

p. 134, Equation (6). Close brackets after -1.870).

p. 135, Equation (7) should read $tan(\theta_m/2) = \frac{dr}{dz} = -0.91\left(\sqrt{\frac{\rho g}{\sigma}}\, r - 0.935\right)$

p. 135, two lines below Equation (7): *Letting* r_m be the . . .

p. 135, col. 1, line 30: . . . different *than* those of . . .

Minimum Diameter Stalagmites, *Bull. Nat. Speleo. Soc.*, 1973, 35(1):1-9.

p. 6, Equation 9. Close brackets after $(-t/\tau)$

p. 6, Equation 11. Argument of exponential should be exp $(-t'/\tau)$

p. 8, Equation 8. First term should be $\delta \frac{dc}{dt}$.

Curl, R. L. (1974)—Deducing Flow Velocity in Cave Conduits from Scallops: *NSS Bull.* 36(2):1-5.

ERRATA

p. 3, Eqn (6).

$$\overline{L}_{32} = \frac{\sum_i l_i^3}{\sum_i l_i^2} \qquad (6)$$

7 lines further: . . . $Re_L = \rho\, \overline{u}(\overline{L_{32}})\overline{L}_{32}/\mu = \ldots$

Eqn (8). omit parenthesis after B_L.

4 lines further: . . . to calculate $\overline{Re}_L$.

p. 4, col. 1, line 8, . . . $\overline{Re}_D = \rho\overline{u}D/\mu$.

line 9, . . . constants Re^* and . . .

p. 5, col. 2, line 11 *up*: . . . measure $u(\overline{L_{32}})$ or . . .

col. 1, under Characteristic Velocities (Table 1)

Average conduit, $\overline{u}$

Literature Cited:

Blumberg, P. N. and R. L. Curl (1974)—Experimental and Theoretical Studies of Dissolution Roughness: *Journal of Fluid Mechanics.* 65:735-ff.

A Checklist of the Cave Fauna of Oklahoma: Reptilia

Jeffrey Howard Black *

ABSTRACT

Eleven species of reptiles (2 Testudines, 1 Saurian, 8 Serpentes) are reported from Oklahoma caves. All reptiles are ecologically classified as accidentals, even though some species are common in cave entrances. Most of these reptiles have not previously been reported from Oklahoma caves.

Introduction

Information on invertebrate and vertebrate faunas in Oklahoma is sparse. Only the single study by R. C. Harrel (1960, 1963) on Wild Woman Cave, Murray Co. includes information on the total fauna in an Oklahoma cave. The bibliography of cave fauna prepared by Thompson (1970) listed only 40 references, whereas over 100 references (not including those on Chiroptera) dealing with Oklahoma cave life actually exist (Black, 1971).

This paper is part of a continuing report on the cave fauna of Oklahoma. Several papers in this series already have been published (Black, 1971, 1973; Black and Best, 1972) and cover various groups of invertebrates and vertebrates in Oklahoma caves. This paper gives all available records for reptiles in Oklahoma caves and some notes on their habitat.

Reptiles are poorly known with respect to their occurrences in Oklahoma caves. Most records occur when individual reptiles wander or are washed into caves. They would thus be classified as accidentals. Rattlesnakes commonly are associated with Oklahoma caves and this results in little incentive to collect reptiles. Also, many reptiles are difficult to capture and preserve.

Methods

Extensive collections of vertebrates and invertebrates have been made in over 30 caves, while numerous other caves have been examined briefly for fauna. Limestone, gypsum, granite, conglomerate, and sandstone caves were included, to obtain a broad spectrum of the Oklahoma cave fauna. A standardization of Oklahoma cave names has not yet been completed. The names used in this report are those presently accepted by the Central Oklahoma Grotto. Detailed locations of these caves are available from the Grotto [1] and some are given in Glass and Ward (1959).

Terminology used to indicate the probable ecological classification of species is that defined in Barr (1963). No troglobites, troglophiles, or trogloxenes are recognized among the reptiles collected in Oklahoma caves.

Results and Discussion

Following is an annotated list of the turtles, lizards, and snakes observed in Oklahoma caves. Common names of reptiles follow the recommendations of the Committee on Herpetological Common Names (1956). Cave names and, in some instances, cave locations are listed after county names.

PHYLUM CHORDATA

Class Reptilia

Order Testudines

Family Testudinidae

Terrapene ornata ornata (AGASSIZ) Ornate Box Turtle
Oklahoma records. HARMON CO.: Reed Bat Cave.
Comment: Accidental. Collected and removed from 250 ft inside the cave, where it apparently had washed during high water.

Terrapene carolina triunguis (AGASSIZ) Three-toed Box Turtle
Oklahoma records. ADAIR CO.: Turtle Cave. DELAWARE CO.: Natural Funnel Cave.
Comment: Accidental. Three-toed Box Turtles frequently are found in the entrances of caves and, occasionally, at the bottoms of vertical pits where they have become trapped.

Order Squamata

Suborder Sauria

Family Teiidae

Cnemidophorous sexlineatus viridis LOWE Six-lined Racerunner
Oklahoma records. HARMON CO.: Reed Bat Cave.
Comment: Accidental. Found about 250 ft inside this gypsum cave, where it apparently had washed during high water in the cave.

Suborder Serpentes

Family Colubridae

Elaphe obsoleta obsoleta (SAY) Black Rat Snake
Oklahoma records. ADAIR CO.: Charley Owl Cave.
Comment: Accidental. A single Black Rat Snake was discovered in a crack on a limestone wall, about five ft off the cave floor.

Elaphe guttata emoryi (BAIRD and GIRARD) Great Plains Rat Snake
Oklahoma records. KIOWA CO.: cave 60 ft deep (Webb, 1970). WOODWARD CO.: J. Selman Cave System.
Comment: Accidental. The snake in Woodward Co. was found swimming in a small pool of water about 100 ft

* Department of Biology, Oklahoma Baptist University, Shawnee, Oklahoma 74801 *and* J. Willis Stovall Museum of Science and History, University of Oklahoma, Norman, Oklahoma 73069.

[1] c/o National Speleological Society, Cave Avenue, Huntsville, Ala. 35810.

inside the gypsum cave. It was very emaciated and soon died. In Kiowa Co., one was found in "a cave 60 feet deep" (field notes of A. I. Ortenburger in Webb, 1970).

Masticophis flagellum flagellum (SHAW) Eastern Coachwhip

Oklahoma records. SEMINOLE CO.: Gar Creek Cave.

Comment: Accidental. In January, 1974, while attempting to dig out a lower entrance into the above small, conglomerate cave, several hibernating Eastern Coachwhips were uncovered.

Lampropeltis calligaster calligaster (HARLAN) Prairie Kingsnake

Oklahoma records. MAJOR CO.: Nescatunga Cave.

Comment: Accidental. This Prairie Kingsnake was very weak when found deep within this gypsum cave.

Natrix sipedon pleuralis COPE Midland Water Snake

Oklahoma records. ADAIR CO.: Christian School Study Cave.

Comment: Accidental. Collected in the twilight zone in a small pool of water.

Family **Viperidae**

Agkistrodon contortrix (LINNAEUS) Copperhead

Oklahoma records. CHEROKEE CO.: Tahlequah Pit. DELAWARE CO.: Stansberry-January Cave. ADAIR CO.: Duncan Field Cave System, Sam's Pit, Three Forks Cave.

Comment: Accidental. The floors of cave entrances in northeastern Oklahoma frequently are covered with thick layers of leaves, thus offering habitat for the Copperhead. Some also have been found in small pits, where they had become trapped.

Crotalus atrox BAIRD and GIRARD Western Diamondback Rattlesnake

Oklahoma records. MAJOR CO.: Nescatunga Cave, Roadside Pit, Saloon Cave, Vickery Waterfall Cave, 1500 Foot Cave, Young Jud Cave. WOODWARD CO.: Hathaway Cave, Horseshoe Cave, Hot Cave, Murray's Rattlesnake Cave, J. Selman Cave System, Buell's Rattlesnake Cave.

Comment: Accidental or, perhaps, a trogloxene. Rattlesnakes are common inhabitants of cave entrances in western Oklahoma, especially where rocks and leaves are plentiful.

Crotalus horridus horridus LINNAEUS Timber Rattlesnake

Oklahoma records. ADAIR CO.: Three Forks Cave.

Comment: Accidental. A single Timber Rattlesnake was found in the entrance to this limestone cave.

Acknowledgements

This study was partially supported by a Grant-in-Aid of Research from the Society of Sigma Xi, a Grant-in-Support of Research from the Theodore Roosevelt Memorial Fund of The American Museum of Natural History, and a travel grant from the Oklahoma Biological Survey. I am grateful to members of the Central Oklahoma Grotto for their valuable assistance and support of this study.

LITERATURE CITED

Barr, T. C., Jr. (1963)—Ecological classification of cavernicoles: *Cave Notes* 5:9-12.

Black, J. H. (1971)—The cave life of Oklahoma: A preliminary study: *Oklahoma Underground* 4:2-53.

——— (1973)—A checklist of the cave fauna of Oklahoma: Amphibia: *Proc. Okla. Acad. Sci.* 53:33-37.

——— and T. L. Best (1972)—Remains of a Gray Wolf (*Canis lupus*) from northwestern Oklahoma: *Proc. Okla. Acad. Sci.* 52:120.

Committee on Herpetological Common Names (1956)—Common names for North American amphibians and reptiles: *Copeia* 1956:172-185.

Glass, B. P. and C. M. Ward (1959)—Bats of the genus *Myotis* from Oklahoma: *Jour Mammal.* 40:194-201.

Harrel, R. C. (1960)—A preliminary report on the invertebrate animals of Wild Woman Cave: *Proc. Okla. Acad. Sci.* 40:29-34.

——— (1963)—Further notes on the invertebrate animals of Wild Woman Cave, Murray Co., Oklahoma: *Proc. Okla. Acad. Sci.* 43:129-131.

Thompson, J. H. Jr. (1970)—Cave fauna of Oklahoma. I. Bibliography excluding Chiroptera: *Proc. Okla. Acad. Sci.* 49:102-104.

Webb, R. G. (1970)—Reptiles of Oklahoma: Univ. Oklahoma Press, Norman, 370 pp.

Manuscript received by the Editor 18 March 1974.
Revised manuscript accepted 29 April 1974.

The NSS Bulletin is published quarterly, in January, April, July, and October. Material to be included in a given number must be received at least 90 days prior to the first of the month in which publication is desired. The subscription rate in effect 23 March 1974: $7.00 per year.

Office Address:

NATIONAL SPELEOLOGICAL SOCIETY
CAVE AVENUE
HUNTSVILLE, ALABAMA 35810

Discussion of papers published in the Bulletin is invited. Discussions should be 2,000 words or less in length, with not more than 3 illustrations. Discussions should be forwarded to the appropriate editor within three months of publication of the original paper.

THE NSS BULLETIN

QUARTERLY JOURNAL

of the

NATIONAL SPELEOLOGICAL SOCIETY

VOLUME 36, NO. 3 JULY 1974

CONTENTS