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NATIONAL SPELEOLOGICAL SOCIETY

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The Greenbrier Caverns

John M. Rutherford * and Robert H. Handley †

ABSTRACT

The Greenbrier Caverns is the largest-known American cave east of the Central Kentucky Karst. Modern-day exploration of the cave began in the very late forties, when a small group from Charleston, West Virginia headed by Bob Handley became interested in "Organ-Hedricks" Cave. One discovery followed another in rapid succession. By 1958, the Greenbrier Caverns was known to have seven interconnected entrances and, with recent reports of over 44 miles of mapped passages and two more entrances, the system is surely among the 10 largest caves known in the world. The cave has a long history, dating back to the days of President Jefferson. Two distinctly different types of saltpeter workings are present in the cave. The more recent and extensive of the two is of Civil War age, or older. Collection of biological specimens from the system began in the thirties and a diverse fauna is known today. The cave is the type locality for four of the 14 troglobitic species known to occur in it. Bone collections from three different areas of the cave include (extinct) Pleistocene nine-banded armadillo, mastodon, Jefferson's ground sloth, and peccary. Geologically, the cave is very complex and the effects on cavern development of a variety of geologic factors can be seen. Joint control, bedding plane control, tilted and folded strata, numerous faults and thrust planes, and the contact between the Greenbrier limestone and the underlying Maccrady shale are among the geologic factors illustrated. The hydrology of the cave is correspondingly complex and consists of two (or three) parallel drainage lines developed along a structurally complex syncline.

Introduction

The Greenbrier Caverns, ranked fifth in the most recent edition of the long caves list of the United States (White, 1970), is the largest known cave in West Virginia. Included among its many interconnected entrances is Organ Cave, one of the five commercial caves in West Virginia. A long history is associated with Organ Cave. About one-half mile inside the Organ entrance, along the Upper Stream Passage, several 1924 dates have been clearly scratched onto breakdown littering the floor. Extensive saltpeter workings in the cave are known to date from the Civil War. A room containing remains of about 37 hoppers (Fig. 1) worked by Confederate troops under General Lee is one of the highlights of the commercial tour in Organ Cave. Some of these hoppers are remarkably well preserved. Pickmarks in the fill in the nearby Organ Stream Passage are abundant.

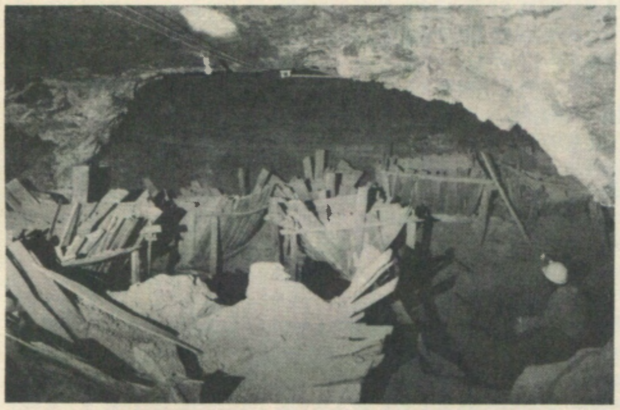

Fig. 1. Civil War saltpeter hoppers, worked by Confederate troops.

Less well known aspects of the historical situation, however, include the facts that Organ Cave was shown as a tourist cave as early as 1835 and that saltpeter miners' tally marks and inscriptions dating from the early 1800's are also found in this part of the cave (Fig. 2). Hence, it is clear that saltpeter mining in this section actually dates from the War of 1812, or earlier. Writing in an old, flourishing script, including a definite date of 1802, has been found along the Upper Stream Passage one-half mile inside the cave.

A set of still older and much more primitive saltpeter workings has been found along the Old Saltpeter Route, much deeper in the cave. These workings consist of bathtub-sized basins in dirt fill which are lined with wooden shingles. Originally, an all-wood shovel and wooden wheel were found in association with these workings (Fig. 3). Pieces of charcoal scattered about

Fig. 2. Saltpeter miners' tally marks, dating from the early 1800's.

* West Virginia Association for Cave Studies, 46 Cabot Lane, Kinnelon, N. J. 07405

† West Virginia Association for Cave Studies, Rt. 1, Box 647, St. Albans, W. Va. 25177

Fig. 3. Primitive, basin-type saltpeter working, with wooden shovel, on the Old Saltpeter Route.

and knotted strands of bark found there are both indicative of the antiquity of these workings, which are presumed to date from before the War of 1812. Modern explorers obtained access to the Old Saltpeter Route by removing rocks from a breakdown. This fact, combined with the remote location of the old workings, raises the possibility that early saltpeter miners may have reached the older workings through an entrance which is closed today. The existence of very old saltpeter workings in the cave is not too surprising, because the Greenbrier Caverns is located within a few miles of Union, West Virginia, the western terminus of the Old Virginia Saltpeter Route (Faust, 1964).

The oldest known account presumably of Organ Cave is that of President Jefferson, who reported the discovery of bones of a "quadruped of the clawed kind" in a cave long believed to have been Organ Cave (Jefferson, 1799). According to Jefferson's account, these bones were found by saltpeter miners. Recent historical research by Peter Hauer and Miss Blanche Humphreys, however, indicates that these bones were probably removed from a nearby saltpeter cave which has not been connected with the system (Hauer, 1975).

Exploration and Mapping

Despite these indications of fairly deep penetration into the cave in early times, accounts by these explorers are lacking. The exploration and mapping on which modern knowledge of the Greenbrier Caverns is based has taken place over the last 25 years and includes five fairly distinct phases.

Systematic exploration of the Greenbrier Caverns began in December of 1948, when members of the Charleston Grotto became interested in the area. Based on the rumored large extent of Organ Cave, they began investigating some of the other small caves in the area in the hope that one of these would lead into an extensive system. On a photography trip for *Life Magazine* two years before, they had visited Hedricks Cave, a small, pretty cave nearby. There, a caver named Alice Williams (in bathing suit for the camera) negotiated a wet crawl and returned to report that the cave opened up and continued. Bob Handley and Bob Flack returned to Hedricks Cave in 1948 for a short photography trip. Checking a crawl which bypassed the water passage, Handley found the continuation of Hedricks which Alice Williams had reported (Flack, 1949). Handley, accompanied by Bob Flack and Bob Barnes, pushed downstream a mile to the end of the Hedricks Stream Passage and made the connection with Organ Cave, lying to the east.

Thus, "Organ-Hedricks" was born. One discovery now rapidly followed another. First, the North Entrance was found, leading to a passage that paralleled and then intersected the Hedricks passage. The Sarver Room, the Flack Room, the Handley Room, the Bone Room, and the Room-Without-A-Name were quickly found in a complex of passages lying above and to the west of the Hedricks Stream. At its downstream terminus, the Upper Stream Passage ended in breakdown at the floor of the Waterfall Room, a 40-ft high room bringing in a third level of drainage at the ceiling. Immediately to the east of the Upper Stream Passage, the Revak Room was found and, farther to the east via crawls and canyons, was discovered Cyclops Hall, a segment of another major trunk channel draining toward the lower-level Hedricks Stream Passage.

After this flurry of discoveries, exploration settled down to a less spectacular pace. A fourth entrance, the Sively Entrance, was found in a passage off the upper part of the Organ Stream Passage. The stream in nearby Foxhole Cave, which lies above the Organ-Hedricks system, was dye-traced into the high passage in the Waterfall Room, although a humanly passable connection was not found.

The second phase began in 1951, when Handley, this time accompanied by John Rutherford, made more major discoveries. First in Humphreys Cave (for several years erroneously referred to as "Erwin Cave") and then in Lipps Cave, lying to the west of Organ-Hedricks, they found drops which led to 40-ft-high stream passages. The Humphreys and Lipps trunk passages were quickly found to form a second major drainage net roughly parallel to, but a considerable distance west of, the Organ-Hedricks drainage. A second Lipps entrance was found nearby. Upstream exploration in Lipps-Humphreys was blocked by massive breakdown in each passage. At about this same time, additional passage, continuing beyond the Waterfall Room, was found in Organ-Hedricks by David Bowen. This passage is a continuation of the Foxhole Cave drainage route, which formerly crossed the Waterfall Room at ceiling level. It quickly led to the Bowen Room, containing a 100-ft drop down the face of a huge flowstone cascade. At the bottom of the drop, the Big Canyon was found. This was a 40-60 ft high and 6-10 ft wide passage which headed south for about 3500 ft until it was blocked by a siphon. Near the end of Big Canyon, a long rimstone crawl, named the Meatgrinder, led to the Rutherford Room and crawls beyond. Although the promising leads began to dwindle once again, the concept of the Greenbrier Caverns had now been born—the connection of Lipps-Humphreys with Organ-Hedricks.

The third phase of the work began with mapping in Organ-Hedricks by Earl Thierry. Because many of the early explorers had moved away from the area, the search for the connection proceeded slowly until Hugh Jones and Conrad Revak, two budding young cavers from Charleston, had to be rescued from the drop in Lipps after becoming stuck. Undaunted by their awkward introduction to the system, they eagerly joined Handley in the search for the connection by a systematic check of all leads in the northwest part of Organ-Hedricks. Working off the Handley Room, they ultimately pushed a series of rocky crawls westward into Jones Canyon, another stream passage which lay between and parallel to the two parts of the presumed system. Then, working through a maze of crawls, the three cleared a breakdown choke and finally made the connection with Lipps in September of 1958. The Greenbrier Caverns became a reality (see Fig. 4). Mapping in the system continued for a while longer, but was soon suspended when complications developed.

Through the years, traffic in the system had steadily increased as tales of "big cave" circulated through the caving community. The Hedricks Entrance was closed in the late fifties as a result of a series of incidents which is almost a classic in the deterioration of caver-owner relations. The owners were being besieged by hosts of cavers from the east and northeast. Fences were being ridden down and, on two occasions, the owner had to retrieve calves which had fallen into the 25 ft entrance pit after careless cavers left it uncovered. On one occasion, a party of unwelcome cavers refused to vacate the premises until the owner's wife fired a shotgun into the air. Accordingly, it was decided to suspend work in the system for a while, in the hope that this rapidly worsening situation would cool down.

The fourth phase of the work began in 1964 when Charles Maus and Henry Stevens quietly resumed surveying in the less heavily travelled Lower Lipps portion of the cave. More passage was found to the west of the Lower Lipps Stream. A connection was found from the lower end of Jones Canyon to the Rutherford Room and the Big Canyon. This phase continued well into 1970. As additional leads were checked, the total mileage of surveyed passages increased to 17 (Fig. 4).

The fifth phase began late in 1970, when the D.C. Grotto launched a massive effort to explore and survey the cave themselves. WVACS then discontinued further mapping and exploration in the system and gave their survey data to the D.C. Grotto. Although major new discoveries do not appear to have been made, two more entrances and some additional passages have been found, and a surprising claim of nearly 45 miles of passages has been made (Wells and DesMarais, 1973). If this is substantiated, then the Greenbrier Caverns, to which the D.C. Grotto recently has been referring as "the Organ Cave System," possibly ranks among the five largest caves known in the world.*

Biological Findings

The greatest part of the biological work that has been done in the Greenbrier Caverns has been in taxonomy. Many interested people, beginning with Reese (1934), have collected specimens and have forwarded them to a variety of specialists in systematics for identification. Although some of this material has not yet been described, the Greenbrier Caverns is thus far recorded as the type locality for four troglobitic species: the snail *Fontigens tartarea* (Hubrict, 1963), the isopod *Asellus holsingeri* (Steeves, 1963), the amphipod *Stygonectes emarginatus* (Hubrict, 1943), and the pseudoscorpion *Kleptochthonius hetricki* most recently described (Muchmore, 1974). While some species known to be present in other caves in this general area do not seem to have been recorded from the Greenbrier Caverns, it is apparent that this cave system has a diverse fauna. The faunal list (Table I) provides a résumé of the forms known from the system (Holsinger, 1971).

Interestingly enough, crayfishes do not appear to have been collected from the cave. In the early days of the exploration, they were plentiful in the Hedricks and Organ streams. During recent years, however, crayfishes have not been observed in the course of periodic visits to the cave by one of the authors (J. M. R.). Also apparently lacking in recent years are the creek chubs which were formerly observed with some frequency in the Hedricks Stream near its confluence with the Organ Stream. Whether this represents an actual loss of these forms from the cave or, rather, a population decrease and/or a migration of these forms to the more remote portions of the cave cannot be determined without a more rigorous investigation.

Although the majority of biological findings in the Greenbrier Caverns has been in the area of systematics, the Greenbrier Caverns has also been included in areal studies of cave plankton (Starr, 1968) and morphological variation in the amphipod *Gammarus minus* (Holsinger and Culver, 1970). Directly related to cave biology are the effects of external environmental factors on the so-called "constancy" of the cave environment. The influence of external conditions on cave temperatures has been shown to extend for thousands of feet into this system (Cropley, 1965).

Pleistocene bones have been reported from at least two and, possibly, three different areas of the cave. Jefferson's "quadruped of the clawed kind" was described (Wister, 1799) as a "three-toed" ground sloth, *Megalonyx jeffersonii*, more commonly known today as Jefferson's ground sloth. However, it is highly unlikely that these bones came from the relatively remote areas of recent finds, if, in fact, they actually were removed from this cave system.

Bob Handley and Bob Flack discovered pleistocene remains in the Bone Room (R. Handley, 1950), including the peccary *Mylohyus nasutus* (C. Handley, 1956). Subsequent investigation by the Carnegie Museum found, in addition to peccary, remains of horse, bear, and armadillo *Dasypus cf bellus* (Guilday and McCrady, 1966). A bone collecting program by WVACS personnel (Rutherford and Holsinger, 1969) has also turned up Pleistocene material including mastodon (*Mammut americanum*) and the first Appalachian record of the grizzly bear (Guilday, 1971) in the vicinity of the Waterfall Room. A summary of the bones which have been identified from the Greenbrier Caverns is given in Table II.

Geology

The geology of the Greenbrier Caverns is extremely complex and, rather than attempting comprehensive explanation, some of the important structural and stratigraphic aspects of the system, only, will be briefly noted and described. The cave is developed in the lower parts of the Greenbrier formation, a sequence of Mississippian limestones which crops out in the eastern portion of the state in a northeast-southwest oriented band approximately parallel to the West Virginia-Virginia border. The Greenbrier formation reaches its maximum thickness (about 1000 ft) in Monroe County, to the south. It ranges from 500 to 750 ft thick in Greenbrier County. The two lowest members of the Greenbrier are the Hillsdale and Sinks Grove limestones, which correspond (approximately) to the St. Louis and Ste. Geneviève limestones of Kentucky (McCue, Lucke, and Woodward, 1939). The Organ, Hedricks, and Humphrey (erroneously called Erwin in early reports) entrances are reported to be in Hillsdale limestone, at the bottom of the Greenbrier formation, while the Lipps entrances are reported to be in the overlying Sinks Grove limestone (Davies, 1969). The Hillsdale limestone is a gray, massive, crystalline limestone containing numerous nodules and layers of black chert up to two inches in thickness. The Sinks Grove limestone is a blue, massive, crystalline limestone which occasionally is oölitic and contains only scattered nodules of chert. Since these limestones were originally described (Price and Heck, 1939), there has been some debate over the proper disposition of the Sinks Grove and overlying Patton limestones, with the suggestion that a distinction should not be made between these two and that they should be merged into a single unit, termed the Denmar formation (Wells, 1950).

The Greenbrier formation rests upon the Maccrady series, a sequence of about 250 ft of red shales and weakly bedded sandstones. The contact of the Hillsdale with the underlying Maccrady shale is an unconformity of some magnitude in Monroe County to the south (Reger, 1926) and in Randolph County to the north (Reger, 1931) as well as in Greenbrier County. At the

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* Editor's Note: The D.C. Grotto is currently conducting a study of the "Organ Cave System." Persons desiring additional information on the status of work in the system or who wish to participate in this project should contact the D.C. Grotto, the address of which may be obtained from the NSS Office, Cave Avenue, Huntsville, Alabama.

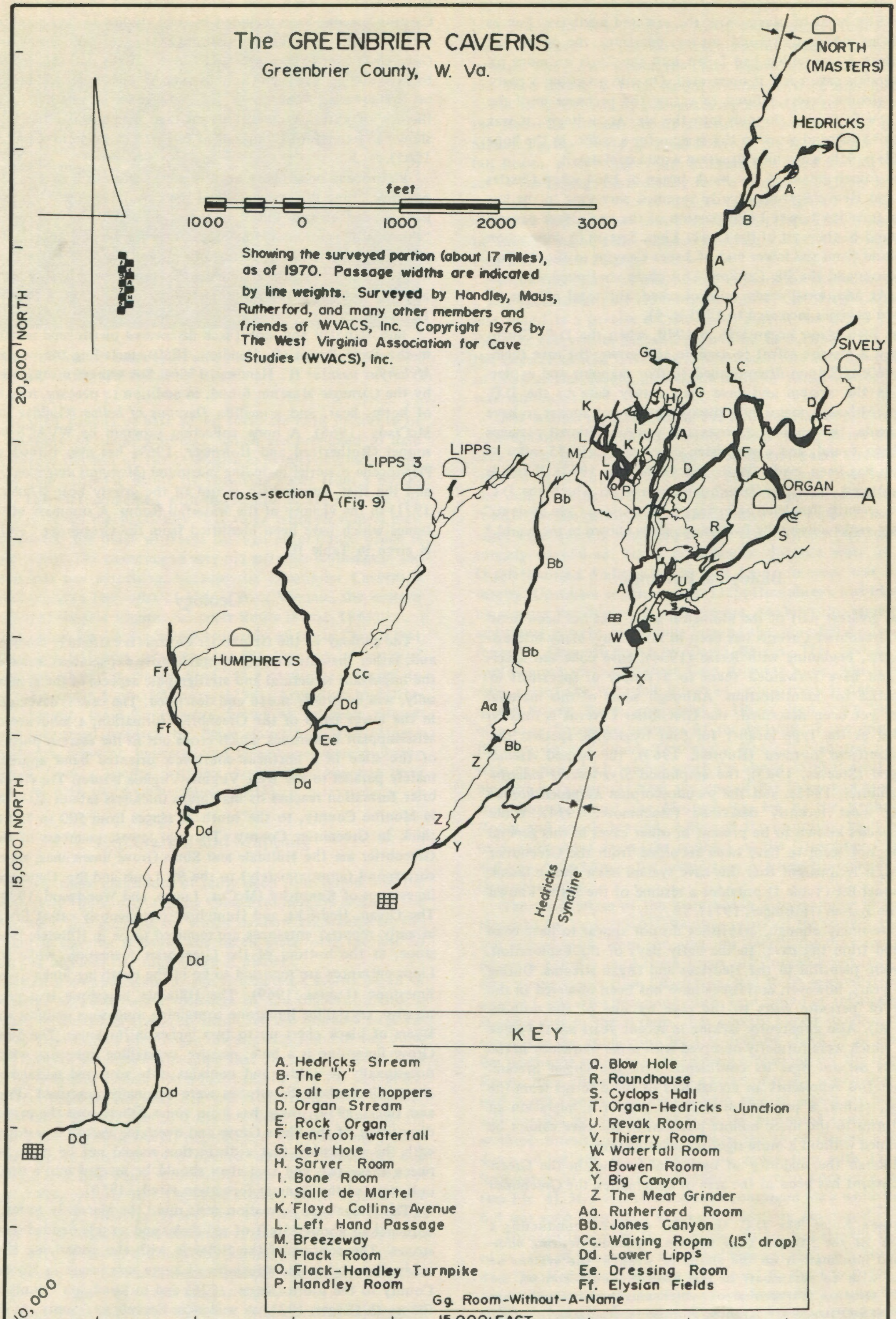

Fig. 4. Surveyed passages in the Greenbrier Caverns, as of 1970. Passage size is indicated by line width.

TABLE I. Fauna of the Greenbrier Caverns
(Compiled mainly by David Culver and John Holsinger).

Class or Order/Genus and Species	Troglobite	Troglophile	Trogloxene	Remarks
	Degree of Adaptation			
Tricladida (planarians, etc.)				
Sphalloplana sp.	X			Probably an undescribed species. Not common.
Gastropoda (snails)				
Fontigens tartarea	X			Type locality. Common under rocks in parts of Organ Stream.
Copepoda (copepod crustaceans)				
Bryocamptus nivalis		X		
B. sp.		X		Little is known of these species.
Attheyella illinoisensis		X		
Isopoda (isopod crustaceans)				
Asellus holsingeri	X			Type locality. In streams under rocks and in gravels.
Amphipoda (amphipod crustaceans)				
Stygonectes emarginatus	X			Type locality. Found in stream gravels and small pools.
S. spinatus	X			
Gammarus minus		X		Common in streams of caves of the area.
Crangonyx sp.		?	?	Uncommon and poorly known. Probably undescribed.
Decapoda (crayfish, etc.)				
Cambarus sp.		X		Not collected, but probably *Cambarus bartoni.*
Araneae (spiders)				
Anthrobia sp.	X			Probably, an undescribed species occurs here.
Linyphiidae (family)	X	X		Several other species of this family have been observed.
Acarina (mites and ticks)				
Rhagidia sp.	X			Possibly, an undescribed species of mite.
Pseudoscorpionida (pseudoscorpions)				
Kleptochthonius hetricki	X			Type locality.
Diplopoda (millipedes)				
Ophyiulus pilosus		?		
Pseudotremia sp.		?		A common but undescribed, purple to brown species.
Trichopetalum (=Zygonopus) packardi	X			Tiny and white.
T. weyeriensis	X			Also tiny and white, possibly intergrading with *T. packardi* in this area.
Collembola (springtails)				
Pseudosinella gisini	X			Common in area caves.
Tomocerus sp.		X		A larger, gray-colored form has also been observed.
Coleoptera (beetles)				
Pseudanophthalmus grandis	X			Common in area caves.
P. fuscus	X			Relatively rare compared to *P. grandis.*
Orthoptera (crickets, etc.)				
Hadenoecus sp.		X		Undescribed species common in several areas of the cave.
Amphibia (salamanders and frogs)				
Gyrinophilus porphyriticus		X		Common in stream caves of the area.
Desmognathus fuscus			X	
Eurycea lucifuga			X	Rather common near entrances.
E. bislineata			X	Not common.
Plethodon glutinosus			X	Not common.
Rana pipiens			X	Tends to be found along the upper stream.
Rodentia (rodents)				
Neotoma			X	Seen near entrances.
Chiroptera (bats)				
Myotis lucifugus		X		
M. subulatus leibii		X		
M. sodalis		X		
M. keenii		X		
Pipistrellus subflavus		X		
Eptesicus fuscus		X		

TABLE II. Bones recovered and identified from the Greenbrier Caverns.

Species	Common Name	Status*	Remarks
Didelphis virginiana	Virginia opossum	P	
Myotis grisescens	gray myotis bat	A	Present nearby, in Clinch Valley of SW Virginia
Myotis keenii	Keen's myotis bat	P	
Lasiurus borealis	red bat	P	Found in float with *M. nasutus.*
Megalonyx jeffersonii	Jefferson's ground sloth	E	From a different cave?
Dasypus bellus	nine-banded armadillo	E	
Peromyscus sp.	white-footed mouse	P	
Canis familiaris	domestic dog	D	
Ursus americanus	black bear	A	Found in the higher mountains within 50 miles.
Ursus arctos-horribilis	brown or grizzly bear	A	First Appalachian record.
Ursus sp.	?	A or E	Fragmentary remnants in float with *D. bellus, M. nasutus,* and *Equus.*
Mammut americanum	mastodon	E	
Equus sp.	horse	D or E	Fragmentary remnants in float with *D. bellus, M. nasutus,* and *Ursus.*
Mylohyus nasutus	long-nosed peccary	E	
Odocoileus virginianus	white-tailed deer	P	
Bos taurus	domestic cow	D	
Ovis aries	domestic sheep	D	

*A=absent from the region today but not extinct D=domesticated E=extinct P=present in the region

signment, which was based mainly on the presence of St. Louis fossils in the bed.

Irrespective of its age or origin, this "reworked Maccrady" layer has exerted a substantial influence on cavern development near the Greenbrier-Maccrady contact (Rutherford, 1967). Being incoherent, very readily attacked by groundwater, and located at the interface between the limestone and the underlying clastic rocks, it is not surprising that passages tend to develop here. Rooms and large passages including Floyd Collins Avenue, the Salle de Martel, Cyclops Hall, Jones Canyon, and parts of the Organ Stream Passage have been developed in this layer (see Figs. 5 and 6). Typically, the ceiling or upper parts of the passages will be in limestone and the walls and floor will consist of the fractured, friable, and crumbling reworked Maccrady layer.

A major structural factor which can be seen in the development of the Greenbrier Caverns results from the interaction of folded and tilted strata with bedding plane control of passage development.* On the small scale, and in the presence of level and undisturbed strata, the perching effects of bedding plane control are frequently easily recognizable. Along the Upper Stream Passage, the perching effects of a chert horizon one to two inches thick and of a nearby shaley layer are evident for over 1,000 ft, in the form of a well-developed ceiling (bedding-plane) anastomosis with tubes 12 inches in diameter (Fig. 7). This anastomosis becomes degraded to sculptured and eroded

Fig. 5. The reworked Maccrady bed in the Organ Stream Passage, showing its contact with the overlying Hillsdale limestone.

remnants which can be traced hundreds of feet farther downstream (Fig. 8). While readily evident on the small scale, perching effects on the larger scale are less obvious because, when displayed in level strata, they are frequently indistinguishable

* Although the reworked-Maccrady-layer effects mentioned above could, perhaps, be considered as a special case of bedding-plane control, the association of the reworked Maccrady with the underlying Maccrady shale and the highly characteristic weathering behavior of the reworked layer suggest that it belongs in a special category of its own.

Fig. 6. Detail of the reworked Maccrady bed. Steep fracture planes dissect the bedding planes.

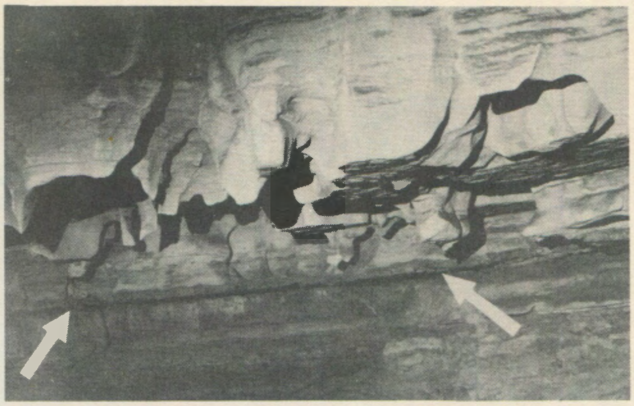

Fig. 8. Sculptured, eroded remnants of the anastomosis shown in Fig. 7, much farther down the Upper Stream Passage. Arrows indicate the chert layer exposed in the wall.

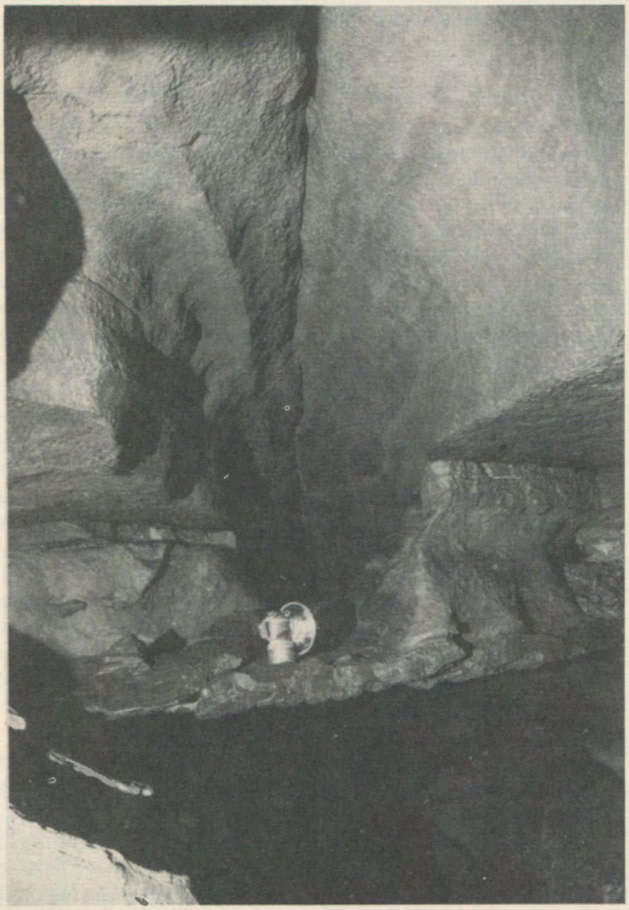

Fig. 7. One tube of a large ceiling anastomosis, perched on a chert layer, in the Upper Stream Passage.

from effects of stable ground-water levels. In the Greenbrier Caverns, however, where large-scale structural factors are present, these larger scale perching effects can also be seen.

The Greenbrier Caverns has been formed along the Patton syncline (Reger, 1926), the axis of which is oriented N20E and plunges gently to the southwest. Casual examination of the large-scale structure displayed in the system indicates that this syncline is not a simple, symmetrical structure. Although the Hedricks Stream follows the major structural trough, the Organ-Hedricks part of the system is developed along both limbs of this structure. Passages are developed down the dip into the trough of the syncline on both sides (Fig. 9). An additional route leading to the synclinal trough is that of the Left Hand Passage into the Handley Room and thence beyond, via the Silo, to the Sand Room on the Upper Stream Passage, which occupies the synclinal trough at a higher level. The zigzag in the Organ Stream Passage at "A" in Fig. 9 appears because, there, the passage turns to the south and follows the strike for about 360 ft before resuming its generally down-dip course toward Hedricks.

This kind of bedding plane control is also evident in a longitudinal section along the Hedricks trough. Three separate stream systems are found on different levels along the trough of the syncline, as shown in Fig. 10. The Hedricks Stream is at the lowest level, very close to the contact with the Maccrady. The Upper Stream lies approximately 50 ft above the Hedricks Stream. The stream in Foxhole Cave, directly atop the Greenbrier Caverns, is approximately 150 ft higher; the Foxhole drainage level is also indicated in the transverse section of Fig. 9. The apparently greater separation between the Upper Stream (Sand Room) and the Hedricks Stream in Fig. 9, compared to that in Fig. 10, is due to the facts that the trough plunges SW and that all of the passages shown in Fig. 9 are not in a common transverse plane. The Sand Room is actually located 400 ft up the trough from the Hedricks Stream cross-section shown in Fig. 9. Thus, a greater separation appears. X's in Fig. 10 indicate the passage segments shown in Fig. 9.

Davies (1960) tried to correlate multi-level cavern development in the folded Appalachian limestones with multiple stream terraces on the surface, because the latter probably represent periods of stable ground water levels which should have enabled extensive cavern development under base level control. Evidence in support of this concept has been found in the Flint-Mammoth System (Miotke and Palmer, 1972), although there the strata are relatively level and unfolded.

In the Greenbrier formation in Pocahontas County (ca. 30 miles north), Wolfe (1964) has shown that the levels of major horizontal passage development are correlated with remnants of and terraces associated with the old "Harrisburg Peneplain". But, while noting a tendency toward concordance of two of base of the Greenbrier, near the contact with the Maccrady shale, there is frequently found a layer of calcareous shale up to 30 ft in thickness which appears to be reworked Maccrady material (Price and Heck, 1939) redeposited with the initial accumulation of Greenbrier sediments. The precise nature of this bed has been debated and Wells has questioned Price's original as-

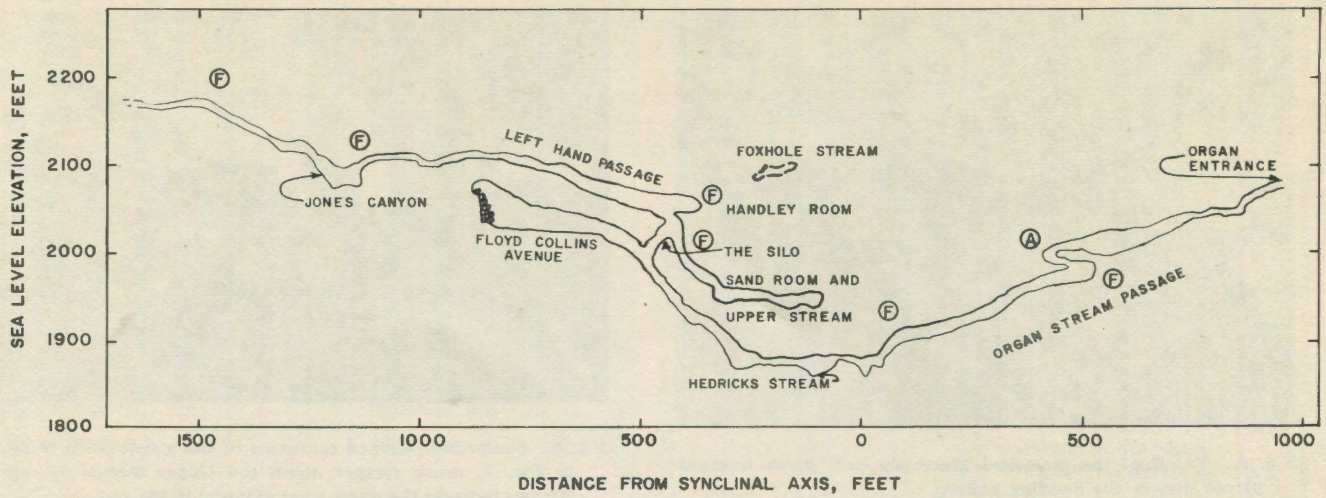

Fig. 9. Transverse section across the Hedricks syncline, looking northward along the strike. Note the different horizontal and vertical scales. At "A", the Organ Stream Passage turns south, following the strike for 360 ft. Areas shown by "F" have larger-scale faulting effects.

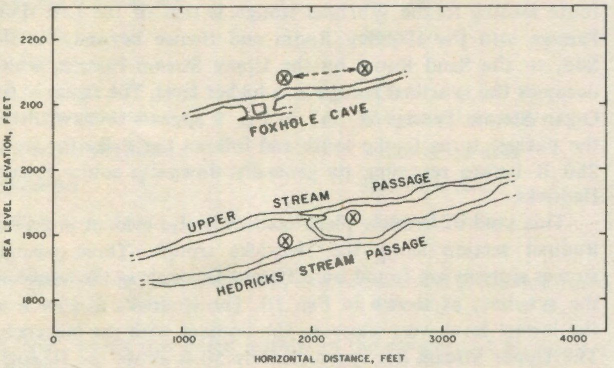

Fig. 10. Longitudinal section along the Hedricks syncline, looking westward. Note the different horizontal and vertical scales. "X's" show location of the cross-section plane of Fig. 9.

these cavern levels with the Hillsdale-Sinks Grove and the Maccrady-Hillsdale contacts, he concluded that lithologic and structural controls were substantially absent.

In the Greenbrier Caverns, however, passages often do not conform to Davies' prerequisite condition of uniform slope and the possibility that lithologic variations here cause substantial perching effects must be considered (see, for example, Fig. 9). Close scrutiny of the Hillsdale limestone has revealed the presence of both shaley and dolomitic layers within it. Ludingtons Cave, 12 miles north of the Greenbrier Caverns, although not located in this synclinal fold, is also developed in Hillsdale limestone near its contact with the Maccrady (Davis, 1963). Arthur Palmer has been able to trace several shaley and dolomitic layers throughout Ludingtons Cave and to relate nearly all of the vertical drops in that cave to the breaching of these resistant layers. The access routes to the major stream passages in the Lipps and Humphreys portions of the Greenbrier Caverns are reminiscent of the situation in Ludingtons Cave. It is tempting to assume analogous perching effects in the western portions of this system, also.

A further complication to the problem of correlating cave development in the Greenbrier Caverns with geologic structure is the frequent presence of faults. High angle normal faulting accompanied by drag folding has been reported only five miles to the north (Wells, Renton, and Perkins, 1966). Major roadcuts of recently constructed Interstate Highway 64 contain graphic evidence of this only two more miles northward. Although faults were dismissed by Davies (1965) as not exerting any significant influence on West Virginia caves, observations throughout the Greenbrier Caverns and in nearby caves have indicated that thrust planes, slickensides, and fracture zones are common (Fig. 11). Indeed, they are so common in the Greenbrier Caverns that it is sometimes difficult to assess their importance, especially when their effects are superimposed on other geologic factors such as lithologic contacts, perching layers, folds, bedding planes, and joints. Most readily apparent are the slickensides which abound on the walls and ceilings, and on the breakdown strewn along the passages. In many parts of the system, such as the downstream portions of the Organ Stream Passage and the Upper Stream Passage, signs of faulting are ubiquitous. They also can be found in many of the breakdown areas in the Greenbrier Caverns. Various factors involved in the occurrence of cavern breakdown have been discussed by Davies (1951) and by White and White (1969), but the influence of faults on the occurrence of breakdown seems to have gone largely unrecognized. In the Greenbrier Caverns, however, the association between faults and breakdown is frequently unmistakable. In many places, passage cross-sections display the imprint of planes of weakness that were generated by faults. Similar effects on a lesser scale have been noted by Werner (1972) in smaller and younger caves in the Cloverlick Valley of Pocahontas County, West Virginia.

Less obvious are some larger scale effects related to faults. There are several locations where these seem to have been especially influential in controlling speleogenesis. A few such places are indicated by "F's" in Fig. 9. Their influence is revealed in at least three different forms. One is the association of downstream passage terminations with major faults, at the end of the Organ Stream Passage and the end of the Upper Stream Passage. Another is the frequent association of room development with faulted and drag-folded strata. A third is the association of larger-scale faults with connections between the different levels in the Organ-Hedricks portion of the system. The "easy way" from the Organ entrance to the Upper Stream Passage follows a curving fault surface which becomes vertical as it nears the level of the Upper Stream (Fig. 11). In the vicinity of the Waterfall Room, where large breakdown and slickensides are prevalent, the three drainage lines stacked vertically along the

Fig. 11. Passage connecting the Organ Stream level with the Upper Stream level. Right wall has vertical slickensides which have been modified by groundwater.

Hedricks trough abruptly come together. The nature of the known cave downstream from that point is quite different from that of the upstream portion of the system. Interconnections between levels also occur near the Handley Room where both faults and folded strata are abundant.

Gregg (1974) has commented on the variability of the ease of solution along fault surfaces, but in the Greenbrier Caverns there is little doubt that faults have been followed where some of the connections between levels have developed. Eddy and Williamson (1967) noted the effect of a fault which breached three stacked cave levels in Cassell Cave, West Virginia.

Joint control is frequently present in the cave, although it tends to be obscured by other geologic processes. However, in parts of the system where the lithology is uniform and the beds are level and undisturbed, the effects of jointing can readily be seen. Probably the best such areas are the Gypsum and Discovery passages (Fig. 12), which tend to be tall and narrow. A chert horizon frequently at ceiling level contains a joint oriented parallel with the passage. Statistical analyses of cave survey data from this and other area caves suggest that many sets of controlling joints are present (Rutherford, 1968).

Hydrology

No significant surface streams overlie the Greenbrier Caverns. There are, however, numerous small sinking streams along the periphery of the limestone outcrop, at its contact with the Maccrady shale. As a result, recharge to the system comes in two forms. One form includes the myriad diffuse drips and seeps through the various joints, fissures, and fractures in the limestone. The other form consists of the many small sinking streams at the limestone-shale contact, which provide more concentrated recharge points.

The subterranean capture of surface streams is most apparent in the development of the Hedricks Stream drainage pattern, since the Hedricks Stream is developed along the trough of the syncline near the contact with the Maccrady shale. Here, tributaries are brought in by both limbs of the syncline and a typical dendritic underground network occurs. This is in distinct contrast to the Upper Stream which, although following the trough of the same syncline, is located above the contact and, hence, does not receive water from the sinking streams. Instead, recharge to the Upper Stream consists primarily of diffuse drips

JOINT CONTROLLED PASSAGE DEVELOPMENT

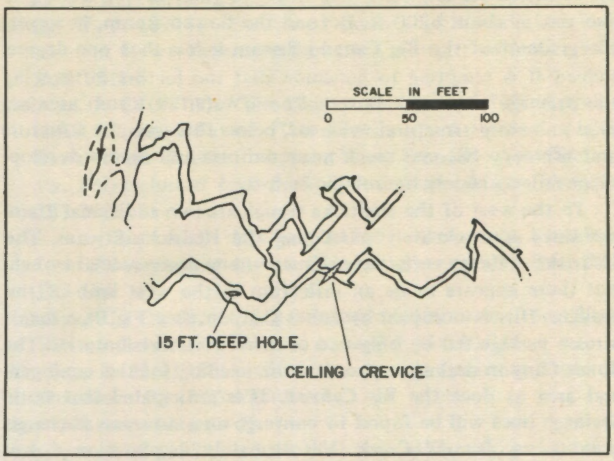

Fig. 12. Joint-controlled passage development in parts of the Discovery Passage and the Gypsum Passage.

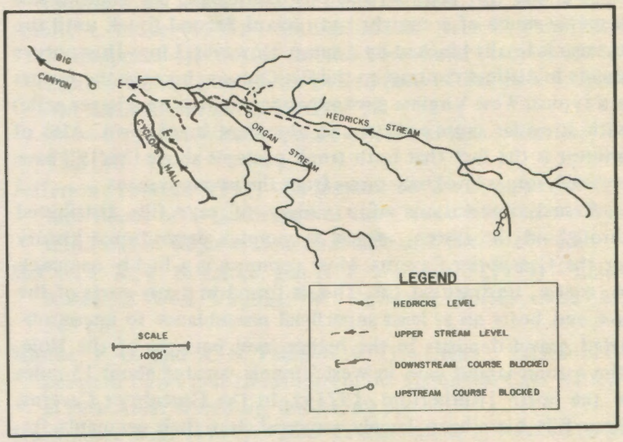

Fig. 13. Comparison of active drainage nets on the Hedricks and Upper Stream levels.

and seeps, which fail to form a developed drainage network. This difference in the two drainage network patterns is illustrated in Fig. 13.

In the vicinity of the Waterfall Room, however, the drainage along the Hedricks trough undergoes a dramatic change at all levels. The Hedricks Stream Passage is blocked by a sump as the ceiling becomes progressively lower. At the intermediate level, the Upper Stream Passage abruptly ends in the bottom of the Waterfall Room. The Upper Stream disappears into the breakdown. At the top of the Waterfall Room, the Foxhole drainage also enters, falls into the breakdown, and disappears. Stream tracing has shown that all three of these drainage lines reappear nearby, at the bottom of the Bowen Room. Beyond the Bowen Room, drainage continues to the southwest, toward its presumed outlet at Second Creek. The passage is finally blocked by a sump.

In addition to the fact that the three drainage lines have merged, there are two other differences between the cave downstream from the Bowen Room and that upstream from it. First, the three stacked drainage lines leading toward the Bowen Room are "typical" West Virginia cave stream passages—irregular, varying in cross-section, and generally littered with breakdown. Beyond, however, is a single high canyon which is free from breakdown.

Second, there is a definite change in gradient. For example, the Hedricks Stream has a 4 to 5 degree gradient over a straight line run of about 6200 ft. Beyond the Bowen Room, however, the gradient of the Big Canyon Stream is less than one degree. Indeed it is tempting to conclude that the factors influencing the drainage above the Bowen Room/Waterfall Room area are predominantly structural, whereas, below this area, the structure and lithology become much more uniform and cavern development reflects mainly hydrologic factors.

To the west of the Hedricks trough are two additional drainage lines approximately paralleling the Hedricks Stream. The Hedricks drainage certainly occupies the major synclinal trough, but there appears to be an inflection in the west limb of the trough. This is occupied by Jones Canyon, (see Fig. 9), a major stream passage fed by only two or three western tributaries. The Jones Canyon drainage appears to be heading for the same general area as does the Big Canyon. It is anticipated that both drainage lines will be found to converge on a common discharge point along Second Creek. Unfortunately, exploration downstream in both passages is blocked by sumps.

Still farther to the west lies the Lipps-Humphreys drainage system with its two 40 by 40 master trunk passages. These converge at the "Y" and, beyond this confluence, the drainage line is pretty much of a straight run toward Second Creek until the passage is finally blocked by a sump. However, Lipps-Humphreys stands in distinct contrast to the Big Canyon, because the former is a typical West Virginia stream passage, (albeit on a larger scale) with irregular cross-section and abundant breakdown. Also of interest is the fact that both trunk passages above the "Y" have perched, upper level tributaries from the two entrances.

Casual observations of a variety of cave fills distributed throughout the system suggest a complex depositional history for the Greenbrier Caverns. Most common is a highly compacted, coarse, unstratified fill. This is found in many parts of the cave and bears an at least superficial resemblance to unconsolidated gravel deposits in the higher level passages of the Hole, (the second largest cave in West Virginia, located about 15 miles to the north [Rutherford, 1971]). In the Greenbrier Caverns, these fills have been largely removed, but their remnants frequently can be found in pockets in the walls and ceilings of various passages. It is apparent that these fills once were both very extensive and very thick, with accumulations probably approaching 40 feet in some places. They are common in many parts of the Organ-Hedricks complex, including the Upper Stream Passage. They are also found in Jones Canyon and in the Lipps-Humphreys part of the system. Although generally coarse and unstratified, at least one instance is known where they are interbedded with two different layers of very fine, varved clays. No indications have been found of the age of these sediments. Because all Pleistocene bone deposits have been found at the surface and are of Wisconsinan age, it is assumed that the gravel beds predate the Wisconsinan glaciation.

The Foxhole level contained extensive unconsolidated fills of considerable thickness in the early days of the exploration. When the cave was re-entered for surveying in 1968, large changes were found in apparent passage sizes and cross-sections. Upwards of 30 ft of fill had been removed in some places. Similar changes have been noted in other small caves in the area. It appears that the small, once-filled caves associated with the Foxhole drainage level are presently being re-excavated. Hurricane Camille in 1969 produced some remarkable examples of deposition and reworking of cavern deposits in this area (Wolfe, 1970; Doehring and Vierbuchen, 1971) but the alterations referred to above predate Camille and appear to be part of an ongoing process independent of such dramatic events. In fact, the fills in Foxhole Cave do not appear to have been affected by Camille.

Summary

The history of the Greenbrier Caverns extends back to the days of President Jefferson. Inscriptions from the early 1800's have been found in two places in the cave and saltpeter miners' tally marks are associated with one set of these inscriptions. Two sets of saltpeter workings are also known. The older, more primitive set is presumed to date from the days of the American Revolution, whereas a set of about 37 more recent hoppers is known to have been worked by General Lee's troops during the Civil War. They constitute one of the highlights of the commercial tours in Organ Cave, one of the nine known entrances to the system.

Systematic exploration began in 1948, with the discovery by Bob Handley of a connection between Hedricks Cave and Organ Cave. An extensive complex of multi-level major passages and rooms was found in the Organ-Hedricks part of the system. Then, large trunk passages were discovered in Lipps and Humphreys Caves to the west and, in 1958, the two portions were connected. The surveyed distance in the Greenbrier Caverns has been reported to be over 44 miles, possibly placing it among the five longest caves known.

Biological collecting in this system started in the 1930's. A diverse fauna, including 14 troglobites, has been found. This cave is listed as the type locality for four of these species. Bone collections date from Jefferson's time and include many mammals no longer present in the area. Extinct Pleistocene remains included in these are nine-banded armadillo, mastodon, long-nosed peccary, and, possibly, Jefferson's ground sloth.

Geologically, the system is complex with effects of both lithologic and structural factors frequently being evident. In the Organ-Hedricks portion of the system, major passage development has occured on three levels stacked along the axis of a large syncline. Major development has also occured across the trough on the lower two levels, thus showing the combined effects of lithologic control and large-scale structure.

The contact of the Hillsdale limestone with the underlying Maccrady shale is not conformable. Frequently associated with it is a "reworked Maccrady" bed which is variable in thickness and readily attacked by groundwater. Throughout the lowest level, the effect of this bed is frequently manifested in the development of large rooms and passages.

Large numbers of faults are exposed in the Greenbrier Caverns, and their imprint is frequently found in slickensided walls and breakdown. Interconnections between levels are often associated with these faults. In addition, faults many times appear to be implicated in folded strata and room development, and in terminations of major passages.

There are many places, however, where other structural factors, such as folded, faulted, or tilted strata, co-exist with apparent joint control. In such cases, these other structural factors usually dominate, with the effects of joints only giving rise to short-range modifications and details in the over-all patterns. Joint control is also evident and a large number of joint sets seems to be involved.

The Greenbrier Caverns contains a wealth of source material for speleological studies dealing with the co-existing effects of variable lithologic, hydrologic, and structural factors. The system is sufficiently extensive so that these effects can be observed in many places. The large number of entrances makes it relatively easy for speleologists to obtain access to most parts of the system. Travel throughout the system is further simplified by an abundance of large passages and by the absence of problems related to seasonal flooding. Thus, the Greenbrier Caverns presents an unusual opportunity for study of the effects of diverse factors on cavern development.

The most elaborate active drainage network in the system is

in the lowest level of the Organ-Hedricks synclinal complex. A dendritic pattern has been formed along the trough here, fed by a multitude of surface streams that sink as they encounter the Maccrady-Hillsdale contact. In contrast, the higher levels in this complex, while following the same axis, have few tributaries because of the lack of recharge from capture of surface streams at the contact. In the vicinity of the Waterfall Room, the active drainage from all three levels has been dye-traced to the stream in Big Canyon, at the bottom of the Bowen Room. Downstream from here, the stream gradient is less and the breakdown and structural effects evident throughout most of the Organ-Hedricks complex are absent. West of this complex, lie Jones Canyon and Lipps-Humphreys, two simpler drainage lines which parallel the structure.

Acknowledgements

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Note Added in Proof

"Adverse caver-owner incidents in 1974 and 1975 have resulted in the closing of the North and Lipps entrances to the system, raising to three the number of entrances closed for such reasons (the Hedricks entrance was sealed for that reason in the Fifties). Mileage claims for the system recently have been revised downward to 32 (*cf* H.W. Blanchard, "The Twenty-Five Longest", *NSS News* 34:29-30 [1976])."

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Pollen Analysis and the Origin of Cave Sediments in the Central Kentucky Karst

Gilbert M. Peterson*

ABSTRACT

Pollen analyses of cave and surface sediments in and near the Central Kentucky Karst show that pollen spectra from cave sediments do not reflect regional pollen spectra, but, instead, are similar to local pollen spectra from the points where sediments enter the cave. Pollen spectra from passages draining the wooded Mammoth Cave Plateau are characterized by high percentages of arboreal pollen, while spectra from passages subject to river flooding show low percentages of arboreal pollen. Thus, pollen analysis is a possible means to distinguish cave sediments from different source areas. Pollen is presently transported into cave passages for distances of at least one-half mile, although no pollen is preserved in older cave sediments.

Introduction

An important question about the Central Kentucky Karst concerns the origin and source of cave sediments in the Flint-Mammoth Cave System. Collier and Flint (1964) stated that the Green River is an important source of sediments for the cave. Watson (1966), on the other hand, believed that the Sinkhole Plain is the major source of sediments for the Flint-Mammoth System. Mineralogical analyses of cave sediments have not enabled previous workers to distinguish between sediments from different source areas (Davies and Chao, 1959; Carwile and Hawkinson, 1969). Pollen analysis is a technique which can be useful in tracing cave sediments to their source areas.

Pollen analysis is the study of fossil pollen and spores from various geological environments. In recent decades, it has proven to be an important analytical technique in the reconstruction of past vegetation and climate (Faegri and Iverson, 1964). Lakes, ponds, or bogs are the most common sites for pollen sampling. Bottom sediments from such sites often provide a continuous record of sedimentation. In the present study, I compare pollen spectra from cave and surface sediments. The pollen spectra are then used to distinguish sediments from different source areas and to trace cave sediments to their respective source areas.

A review of European and American literature revealed that few pollen studies have involved cave sediments; of those pollen studies on cave sediments, the following are representative: Anderson, 1955; Bryant and Holz, 1968; Bryan and Larson, 1968; Derville and Firtion, 1951; Donner and Kurtén, 1958; Lüdi, 1941; Martin *et al.*, 1961; Prošek, 1958; Rivière, 1904; Schütrumpf, 1939; Van Campo and Leroi-Gourhan, 1956. I found no reference among these to pollen analyses involving a large cave system in a major karst area.

The Central Kentucky Karst lies approximately 100 miles southwest of Louisville. Within its boundaries lies the Flint-Mammoth Cave System, the largest known in the world. The geological literature on the Central Kentucky Karst spans several decades. It is discussed in two recent review articles (White *et al.*, 1970; Quinlan, 1970) and in a recent geomorphic study (Miotke and Palmer, 1972).

Only one previous pollen study has been undertaken in the study area. Wright *et al.* (1966) found high values of ragweed in cores from sinkhole ponds, but ragweed accounted for 20 to 65% of the total pollen in all samples. Because studies of short cores show that ragweed attains a high percentage after forested areas are cleared for cultivation (see McAndrews, 1966; Davis *et al.*, 1971; Webb, 1973), the pollen from sinkhole ponds was deposited after settlement of the region by white men and is therefore less than 150 years old. Although Wright *et al.*, offered discouraging prospects for use of pollen analysis for studying the last 15,000 years B.P. in karst areas, their finding pollen in the sediments encouraged the present attempt to use pollen data to trace the origin of sediments within caves.

The Cave Sediment Controversy

The first significant study of sediments in Mammoth Cave was carried out by Davies and Chao and was described in an unpublished Administrative Report to the National Park Service (1959). They made field descriptions of several sediment profiles exposed naturally in Mammoth Cave and undertook reconnaissance sampling and size and mineralogical analyses of cave and surface samples. They found that cave sediments are physically and mineralogically similar to surface sediments and believed that the sediment in the lower part of the cave comes from the Green River.

Collier and Flint (1964) investigated sedimentation rates in lower levels of Mammoth Cave, chiefly the Echo River/River Styx area. They claimed that the Green River is the chief source of sediment and flood water. They reported that, in 2½ years, one-half foot of sediment accumulated in the lower levels, but was subsequently removed by high floods. They found that, after every flood, a thin layer of clay is deposited throughout the submerged areas.

R.A. Watson (1966), taking issue with Collier and Flint, concluded that the amount of sediment moving from the Sinkhole Plain to the Green River must be considerably greater than the amount of sediment deposited in the cave by backflooding and that cave sedimentation must result from underground flow from the surface to the Green River. Thus, contrary to the inference of Collier and Flint, the Green River is not the chief source of sediment and floodwater to the cave, but the cave systems are the chief local sources of sediments and floodwaters for the Green River.

A detailed study of modern sedimentation in a base-level passage near the Green River was undertaken by Carwile and Hawkinson (1969). They made size and mineralogical analyses of sediments from several locations in Columbian Avenue, in Crystal Cave. Sediments captured were thinly laminated silt and clay, reflecting ponding during floods. X-ray diffraction analyses revealed no significant differences between the mineral content of sediments from Columbian Avenue and that of sediments from the Sinkhole Plain.

We can draw two conclusions from the above studies. First, there is disagreement with regard to the source of cave sediments in the Flint-Mammoth system. Second, mineralogical analyses of cave sediments have failed to yield information on the sources of clastic sediments. The present study shows that pollen analysis is a potentially useful tool for distinguishing cave sediments from different source areas.

* Center for Climatic Research, Institute for Environmental Studies, University of Wisconsin, 1225 W. Dayton St., Madison, Wisconsin 53706.

Surface Pollen and Vegetation

The studies of Davis (1967), Lichti-Fedorovich and Ritchie (1968), and Webb (1974) show that pollen spectra from surficial lake deposits generally reflect the regional composition of vegetation. Pollen spectra also respond to local differences in vegetation. If the local vegetation is different in two different source areas of sediment, the pollen spectra of their sediments will reflect these differences. If the pollen-bearing sediments are carried underground, these differences will enable us to trace cave sediments to their source areas. Also, if the characteristic pollen spectrum of sediments from a given source area is unchanged during transport along the cave passage, the pollen spectrum can be used to determine the distance of transport from the source area. To apply pollen-analytical techniques to the question of the origin of cave sediments, I sampled and analyzed cave and surface sediments in and around the study area. Samples were designed to obtain regional pollen spectra and to obtain local pollen spectra from wooded uplands and from the Green River floodplain within Mammoth Cave National Park.

Field and Laboratory Methods

Field work included obtaining samples of ancient sediments from dry upper cave passages, samples of modern cave sediments from shaft drains and base-level passages, local surface samples from within Mammoth Cave National Park, and a series of regional samples from ponds and lakes in Central Kentucky. Sample locations are shown in Figure 1; sampling sites are described in Table 1.* The older sediments were sampled from an upper level, about 600 ft above sea level, in Upper Salts Cave (Samples 16-21, 25), Broadway in Mammoth Cave (44, 46), and Symme's Pit and Blackall Avenue in Mammoth Cave (38-40); they also were sampled in a lower level, about 550 ft above sea level, in Rose's Pass and Rogers Avenue in Mammoth Cave (1-4, 6-8, 10-12, 14, 15), and in Great Onyx Cave (31-36).

Modern cave sediments were sampled in shaft drains, backflooded passages near the Green River, and base level conduits. Sampling areas include shaft drains in: Pohl Avenue (41-43, 74, 75, 82-84), Eyeless Fish Trail (23, 28, 29, 56-58), and Charon's Cascade in Mammoth Cave (59, 63) and in Colossal Cave (54, 55, 71-73, 78-81). Backflooded passages sampled include Columbian Avenue in Crystal Cave (22, 24, 26, 27, 30) and Mammoth Cave River Hall (13, 47, 48, 50, 63, 65).

Base level conduit samples are from Owl Cave (51, 52, 64, 76), Cedar Sink (66), and Mill Hole (70) (not shown in Figure 2).

Surface samples were divided into two groups: local samples from within Mammoth Cave Park and regional pond samples from outside the park. Local samples were from the Green River at Dennison Ferry (45, 49, 67, 68, 69, 77) and the Mammoth Cave Plateau (60, 61, 62). The locations of regional pond samples (121-133) are given in Figure 2.

Sediment samples were treated in the Pollen Laboratory in the University of Wisconsin Geology Department, by means of standard techniques (Faegri and Iverson, 1964).

Approximately 5cc of sediment were treated in most cases. Since the samples were quite rich in mineral matter, the treatment included flotation. After treatment with 10% KOH, the sediment was made acidic with 10% HC1 and transferred to a folded 11-in.-long piece of Tygon tubing, enough $ZnC1_2$ solution (1.9 s.g.) being added to fill the tubing within an inch of the top. The tube was clamped, unfolded, and the contents were mixed by hand. Then the tube was unclamped, filled to within one-half inch of the top with $ZnC1_2$ solution, and centrifuged at 2500 rpm for 20 minutes. This process, including centrifuging, was carried through twice. After each centrifugation, the Tygon tube was decanted into a large plastic centrifuge tube, pliers being used to pinch the Tygon tube about an inch below the surface of the liquid. The decantate was diluted with 95% ethanol and centrifuged. Treatment with HF, acetolysis, and tertiary butyl alcohol followed, after which the samples were stored in silicone oil.

Fig. 1. Sample locations within Mammoth Cave National Park (Base map after Quinlan, 1970). Note: samples 51, 52, 64, 66, 70, 76, and 85 are not shown.

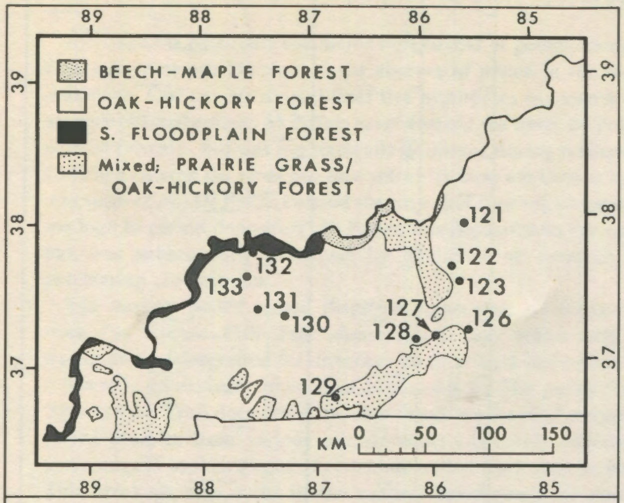
Slides were made at a pollen density suitable for counting and were examined at 600x magnification. Approximately 300 grains per sample were counted. Pollen percentages and 0.95 confidence limits (Maher, 1971) appear in Figure 3.

Results

Surface Samples

Regional surface samples (Samples 121-133) provide pollen spectra (Fig. 3) from small ponds and lakes in central Kentucky (Fig. 2). In general, the spectra show high percentages of oak (average: 23%), ragweed (average: 33%), and grass (average: 18%). The percentage of arboreal pollen averages 39%. Although the percentage of arboreal pollen ranges from 6.1% to 69.7%, 7 of the 11 surface sample AP values occur between 34 and 54%. The vegetation map of Küchler (1964) depicts two major vegetation zones in central Kentucky, namely: oak-hickory forest and a mosaic of forest and prairie. Although agriculture has restricted the oak-hickory forest to small, isolated groves throughout the region,

* Table 1 and Appendix are available free of charge from: NSS Cave Files Committee, Cave Avenue, Huntsville, Alabama.



SITE LOCATIONS

121. 37°55′N 85°40′W	128. 37°09′N 86°06′W
122. 37°39′40″N 85°50′40″W	129. 36°47′20″N 86°50′20″W
123. 37°33′20″N 85°44′W	130. 37°19′15″N 87°13′W
126. 37°13′N 85°43′40″W	131. 37°21′N 87°28′10″W
127. 37°09′N 85°56′35″W	132. 37°47′30″N 87°32′40″W
133. 37°36′N 87°35′W	

DJB 74

Fig. 2. Locations of regional surface samples.

the pollen spectra are typical of the eastern deciduous forest (Peterson, in preparation). These pollen spectra give a regional picture of the modern pollen rain and will afford a basis for comparison with pollen spectra from cave sediments.

Local surface samples from within Mammoth Cave National Park include numbers 45, 49, 60, 61, 62. Although sample 128 is from within the Park, it is grouped with the regional samples because it, also, is a pond sample. The local surface samples are intended to reflect a more localized pollen rain than that shown by pond samples. Samples 45 and 49, from floodplain sediments of the Green River, average 22.2% arboreal pollen, somewhat below the average from the regional samples; they also contain a high percentage of ragweed, higher than all but one of the ragweed values from the regional samples.

Samples 60, 61, and 62 are, respectively, moss from the Salts Cave Sink, soil from the Flint Ridge ranger station, and moss from the Dixon Cave Sink. These samples are all from the forested Mammoth Cave Plateau and are characterized by high percentages of arboreal pollen (average: 64%), higher than those of most regional samples (nos. 121-133) and considerably higher than those of the floodplain samples (nos. 45 and 49).

We may conclude that samples 60, 61, and 62 (from the wooded Mammoth Cave Plateau) are characterized by high percentages of AP. Samples from the Green River floodplain are characterized by low percentages of AP. AP values from the regional samples range widely, but most regional AP values fall between those of the Plateau and of the floodplain. Thus, we may distinguish two local pollen spectra in addition to the regional average, namely "wooded upland" and "floodplain."

Modern Cave Sediments

Pollen samples were obtained from two areas in Mammoth Cave, both of which have large drainage areas and are less than a mile from the Green River. The first sample area is River Hall (Samples 13, 47, 48), which has been discussed as a major area of sediment transport and deposition (Collier and Flint, 1964). The second sample area extends from the Black Onyx Waterfall in Pohl Avenue, along the Eyeless Fish Trail to its junction with Columbian Avenue (Samples 41, 42, 56, 23, 28). The Eyeless Fish Trail is a shaft drain complex with several small, upstream tributaries not shown on the map; its overall length is approximately 2500 ft. It contains a small, perennial stream which carries organic matter, sand, and finer sediments from the Mammoth Cave Plateau toward the Green River.

Pollen spectra from River Hall (see Fig. 1, Samples 13, 47, 48) have a uniformly low percentage of arboreal pollen (average: 28.6%) and a high percentage of ragweed (38 to 47%); pine averages 10.5%, which is considerably above the pine values for regional surface samples. The samples from River Hall (13, 47, 48) have pollen spectra very similar to those from the Green River floodplain (45, 49), the presumed source of sediments for River Hall.

Pollen spectra from the Eyeless Fish Trail (41, 42, 56, 23, 58, 28) have a high percentage of arboreal pollen (average: 80.1%), high pine percentages (40.1%), and relatively low ragweed values (13.3%). The unusually high pine percentages contrast sharply with the pine values from the regional surface samples, but the high pine percentage and the high AP percentage are similar to values from the Mammoth Cave Plateau (Samples 60, 61, 62, Fig. 1). The Mammoth Cave Plateau is the presumed source of sediments for the Eyeless Fish Trail.

No pollen was recovered from the older cave sediments. Some possible explanations are given in the discussion section.

Discussion of Modern Cave Sediments

Of the 43 samples from modern cave sediments, only 9 contained sufficient pollen for counting; the likelihood of successfully sampling pollen in modern cave sediments appears to be considerably less than that of successfully sampling surface sediments. Pollen was obtained from all sediment sizes from clay to sand; many of the successful samples were associated with organic matter, such as leaves and small twigs. The presence of such organic matter in cave passages implies that a more-or-less direct opening to the surface exists, and that water and sediment entering such passages probably is not subject to filtering through pore spaces or narrow joints in the overlying bedrock.

Davies and Chao (1959) and Carwile and Hawkinson (1969) found that modern cave and surface sediments in the study area are mineralogically similar. Pollen analysis has enabled us tentatively to distinguish between sediments from two source areas: the Mammoth Cave Plateau and the Green River.

Watson (1966) identified five means of supply to the underground networks that lead to the Green River. He emphasized the importance of sediment transport from the Sinkhole Plain toward the Green River through base-level conduits and de-emphasized the importance of backflooding. In the present study, a number of samples were taken from base-level conduits, namely: Mill Hole (70), Cedar Sink (66), and Owl Cave (51, 52, 64, 76). None of these samples contained sufficient pollen for counting.

I have, however, found evidence which suggests that the Mammoth Cave Plateau and the Green River are sources of sediment for the cave system. Pollen spectra from Eyeless Fish Trail are similar to those from surface samples on the Mammoth Cave Plateau. Pollen spectra from River Hall are similar to spectra from the Green River floodplain. The pollen spectra suggest that sediments from these two sources are transported into cave passages for distances up to one-half mile or more. Without denying the importance of drainage and sediment transport through base-level conduits, I would agree with Davies and Chao (1959) and with Collier and Flint (1964) that sediments, at least in lower levels near the Green River, are probably largely derived from the Green River.

As mentioned in the Introduction, Wright *et al.* (1966) found that pollen spectra from the Sinkhole Plain yielded only modern,

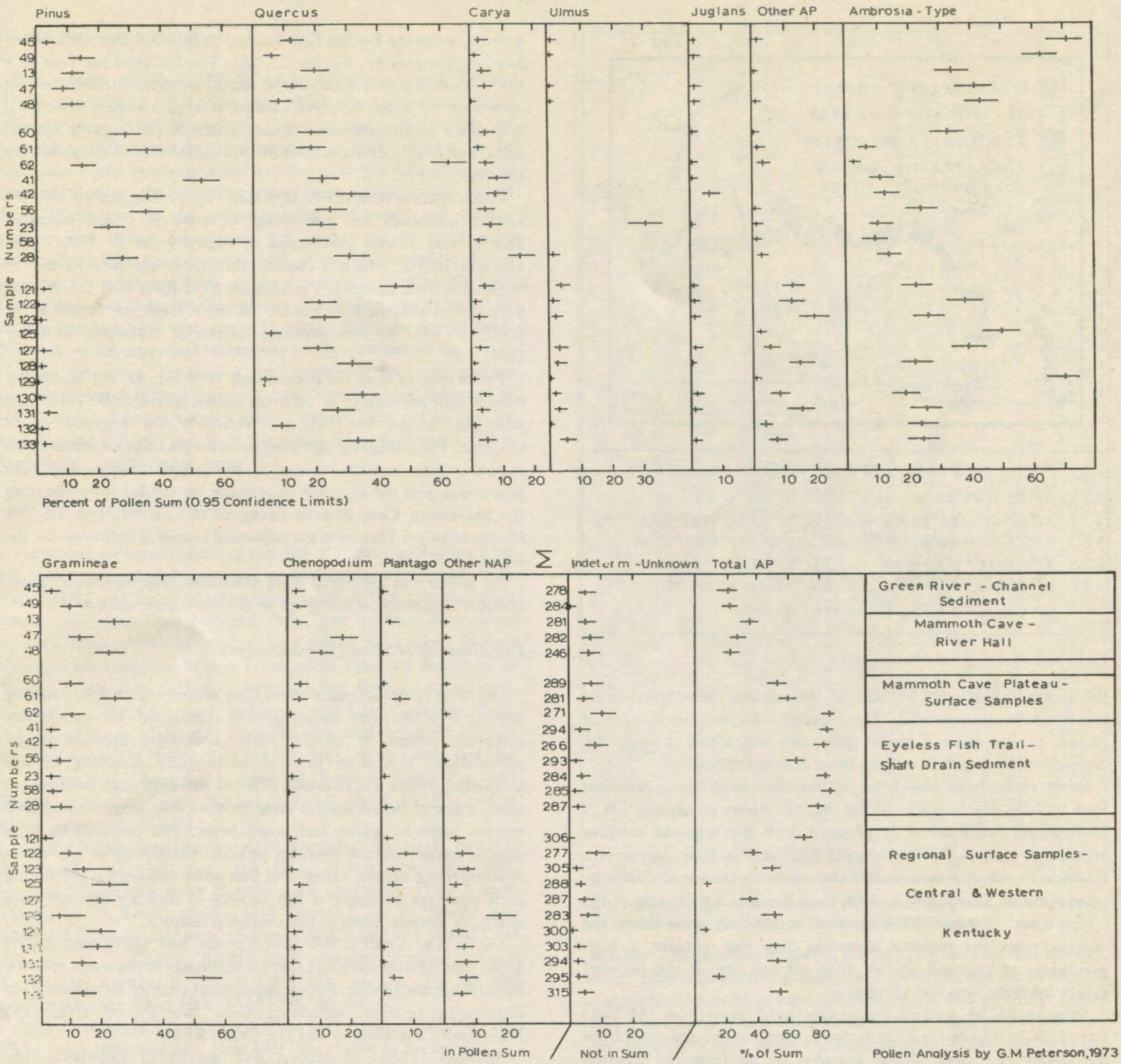


Fig. 3. Pollen spectra.

post-settlement pollen spectra. The present study, also, failed to find evidence of pre-settlement pollen in underground sediments. I did not find evidence that pollen is deposited with underground sediments derived from the Sinkhole Plain, but I did find evidence that pollen enters the cave by shorter routes from the Mammoth Cave Plateau and the Green River. I have shown that, contrary to the suggestion of Wright *et al.*, mechanical abrasion of pollen is not one of the factors limiting the preservation of pollen in the cave. The depositional irregularities of fluvial cave sediments are the chief obstacle to obtaining an underground stratigraphic pollen record. In stratified, finely laminated cave sediments, which might preserve a chronological record, as in Columbian Avenue, water velocities are apparently too low to move pollen. Although the Eyeless Fish Trail contains pollen, it is being transported rather than deposited. River Hall also contains pollen, but sediments there are subjected to much reworking during annual floods (Collier and Flint, 1964), which removes any possibility of obtaining a time-stratigraphic record from them. Thus, in cave passages containing pollen, the flow regime is either too high or too irregular to preserve a pollen-sedimentary record.

Traverse and Ginsburg (1966) observed that, in surface streams, moving water produces some size sorting of suspended pollen types. They noted that the percentage of pine pollen is a function of current conditions rather than of distance from source areas. The present study suggests that hydraulic sorting of pollen is not important once pollen enters the cave. Pollen spectra from Eyeless Fish Trail and River Hall match well spectra from surface source areas. Also, there is little variation in the spectra along the length of a passage. In Figure 3, the cave pollen spectra are shown in order of increasing downstream passage length. No significant changes in the spectra are noted along a given cave passage.

The cited authors, Muller (1959), and Brush and Brush (1972) found pollen associated with silts or fine sands. In the study area, pollen occurs with sands, silts, and organic matter. The local occurrence of pollen is not restricted to a particular sediment size, but pollen only occurs in cave passages where currents are strong enough to transport sand.

Older Cave Sediments

Although the older cave sediments were devoid of pollen, modern cave sediments provide clues to the absence of pollen in the older sediments. One can envision at least five hypotheses to explain the absence of fossil pollen: A) Pollen never entered the cave; B) Pollen entered the cave, but was mechanically destroyed during transport; C) Pollen entered the cave, but flow velocities were too slow to keep it in suspension; D) Pollen entered the cave, but flow velocities were too high to permit deposition; E) Pollen was deposited in the cave, but was subsequently destroyed by oxidation or removed by percolating groundwater.

The modern pollen record disproves hypothesis A. Sediments from the Eyeless Fish Trail demonstrate that pollen can be transported underground for distances up to at least one-half mile.

There is no evidence to support hypothesis B. The pollen from Eyeless Fish Trail does not show significant mechanical abrasion. Pollen samples from various environments commonly contain a percentage of broken grains of vesiculate pollen types such as pine. This breakage may occur in the natural environment or during laboratory processing. To measure the effect, if any, of mechanical abrasion in Eyeless Fish Trail, the following counts were made to determine percentages of broken pine grains:

TABLE 2. Mechanical abrasion of pollen grains.

	Surface Samples		Eyeless Fish Trail Samples			
Sample no.	60	61	41	58	23	56
Pine wholes	60	39	55	60	44	31
Pine halves	8	11	8	6	5	2
Total	68	50	63	66	49	33
Percent pine ½'s	11.8	22	12.7	9.1	10.2	6.0

The samples from Eyeless Fish Trail are arranged in order of increasing downstream distance from the sediment source areas. The percentages of pine halves do not increase downstream. Only one of the samples from Eyeless Fish Trail contains a higher percentage of broken pine grains than do the surface samples; most contain a lower percentage. The evidence suggests that, along the 2500-ft. length of Eyeless Fish Trail, mechanical abrasion of pollen grains is not significant. Thus, contrary to suggestions of Wright *et al.* (1966), mechanical abrasion apparently is not responsible for the lack of pollen in cave sediments.

Hypothesis C, that flow velocities in the cave are too slow to transport pollen, is true for at least some areas of the cave. Columbian Avenue, discussed in Carwile and Hawkinson (1969), is close enough in elevation and horizontal distance to the Green River to be affected by annual floods. Unlike River Hall, however, Columbian Avenue accumulates sediment at an extremely slow rate, perhaps a few millimeters per year. Sediments consist of clay and silt, which settle out of the ponded water. None of the sediment samples from Columbian Avenue (22, 24, 26, 27, 30) contained sufficient pollen for counting. As stated previously, all successful cave pollen samples come from areas where sand is also deposited.

Apparently, areas which accumulate only finer sediments (clays and silts) do not accumulate large quantities of pollen. This is unfortunate in two respects: 1) Clays and silts are less permeable than are sandy sediments; thus, pollen deposited with them is less subject to oxidation or physical removal by groundwater; 2) The finer cave sediments tend to be laminated; thus, providing a better time-stratigraphic record than sandy sediments.

Hypothesis D, that flow velocities were too high to deposit pollen, is certainly true for Eyeless Fish Trail. The passage contains a small perennial stream which continuously moves a thin layer of sediment over the bedrock passage floor. Sediments in Eyeless Fish Trail generally cannot attain a great thickness before they are flushed out of the cave.

Hypothesis E, that pollen was deposited in the older sediments but subsequently was removed or oxidized, cannot be proven or disproved in the present study. It is obvious that pollen enters the cave in certain modern environments, and it is reasonable to assume that it did so in the past. Edwards Avenue in Great Onyx Cave (Samples 31, 32, 33, 34, 35) may have had in the past an environment similar to that of River Hall, which is currently subject to backflooding. Both Edwards Avenue and River Hall are one mile (or less) from the Green River, although Edwards Avenue is now above flood level. Both passages contain sand and finer sediments. Near the end of Edwards Avenue, a natural exposure reveals interbedded gravels and sands grading upwards to laminated silt. No pollen was recovered from the site, although it is entirely possible that pollen may have been present originally. The high permeability of the sandy sediments could have permitted the oxidation of any pollen formerly present. It is also possible that water circulating in the sandy lower layers may have mechanically removed pollen along with other sediment fines. Carwile and Hawkinson (1969) cite an analogous situation in Columbian Avenue. After flood waters recede from the laminated silts and clays of the surface sediment, water continues to circulate in the coarser sands and gravels underneath; the cited authors suggest that clay and silt are selectively removed from the underlying sediments.

In summary, I cannot favor any single hypothesis to explain the absence of pollen in older cave sediments. I can eliminate hypotheses A and B; pollen is transported into the cave system for distances of more than 2,000 ft, with little or no mechanical abrasion. Hypotheses C and D are true for certain modern cave environments. Water velocity in Columbian Avenue, although it produces a laminated sedimentary record, is apparently not sufficient to transport pollen. Pollen occurs in Eyeless Fish Trail, but it is apparently being flushed through the cave rather than being deposited. Even if pollen were to be trapped in Eyeless Fish Trail after diversion of the stream to a lower level, it might later be oxidized or physically removed by flood waters. The lack of pollen in older sediments in a similar environment (Edwards Avenue in Great Onyx Cave) suggests that, if pollen were originally present (which is at least possible), it has probably been oxidized or removed by flood waters.

Summary

Pollen spectra from modern cave sediments do not reflect the regional pollen rain as analyzed from pond sediments; they reflect the local vegetation at surface points of input. The similarity of pollen spectra in cave sediments to local surface pollen spectra suggests that pollen analysis is a potentially useful tool for determining the source areas of cave sediment units the provenance of which cannot be identified on the basis of mineralogic composition. These local biases suggest caution in interpreting fossil pollen records from karst cave sediments. Pollen is currently transported into cave passages for distances up to at least one-half mile. The presence of leaf litter in these passages suggests that pollen enters through direct surface openings. Pollen spectra from a given sample locality are relatively homogenous and do not show evidence of hydraulic sorting along a given passage. The pollen occurs with sand, silt, and organic matter in passages with a flow regime sufficient to transport sand. Pollen-bearing cave sediments are subjected to continual reworking or removal. Modern cave sediments which are clearly stratified do not contain pollen. A systematic sampling of older cave sediments revealed that they contained no pollen.

Acknowledgements

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Doline Densities in Northeastern Iowa

Robert C. Palmquist*, Gary A. Madenford†, and J. Nicholas Van Driel*

ABSTRACT

Variation of doline density in the karsted areas of three counties in Iowa indicates that doline density is controlled by carbonate distribution, thickness of clastic overburden, and bedrock topography. Dolines are limited to areas with a thickness of till and loess less than 25 ft (7.6 m) and are common along the drainageways of bedrock uplands. Average doline densities for the karsted area within a county range from 45 per square mile (17.3/km^2) in Nebraskan drift to 4.6 per square mile (1.8/km^2) in late Wisconsinan drift. Mean doline density (D) increases with time (T) according to D = 18.8 log T — 72.3 in areas of thin drift. Dolines show preferential development along low order drainageways where overburden is thin and ground water recharge is presumed to be large. Data was obtained from S.C.S. county soil reports.

Introduction

The karst of Iowa is undescribed, except for a few scattered references in cave reports and mention in a statistical study of caves (Curl, 1966). The potential area for karst development in Iowa is large. The eastern half of the state is within the Eastern Ground Water District, which is underlain by carbonate aquifers at shallow depth (Tuthill *et al.*, 1972).

This study is an attempt, first, to determine the degree of karst development in Iowa and, second, to evaluate the general geologic controls on doline distribution. Dolines, rather than caves, are used as a measure of karst, because of their ease of recognition and the resulting potential for observing the entire population. The study is divided into two parts. The first part, presented herewith, compares doline density and distribution and the differences in density attributable to time. Therefore, this study concentrates on the number and distribution of dolines. It is not concerned with the size, shape, or origin of the dolines, though brief mention is made of these characteristics. The second part will be concerned with two townships within one country and will attempt to evaluate the detailed environmental controls on doline density, size, and shape (Palmquist, in preparation).

Study Area

The karst areas in three counties which lie in different geomorphic regions of Iowa were selected for study (Fig. 1). The areas were glaciated and have surface materials of different ages, such that the temporal aspect of karst development can be investigated. The areas, however, are underlain by different carbonate units and have undergone different amounts of dissection. All of the counties have similar climates, though they are in different microclimatic regions (Karsten and Tuttle, 1970). The karst in the three counties may be either mantled or covered karst, because of the blanket of Pleistocene drift and loess; it is best classified as fluviokarst (Sweeting, 1973, p. 259), because of its combination of karst and fluvial features. The dolines are, therefore, either subsidence or collapse dolines, in the terminology of Jennings (1971, p. 125-26). While the surface depressions must be younger than the Pleistocene-age materials, the bedrock karst features may be either younger or older than the overburden. The geochemical speculations of Moorehouse (1966) and the geomorphic reasoning in Hedges (1967) suggest that the caves are Pleistocene in age and the karst is mantled, that is, it is pre-drift in age.

Allamakee County is situated in the "Driftless Area" of northeast Iowa (Fig. 1). The county is underlain by early Paleozoic

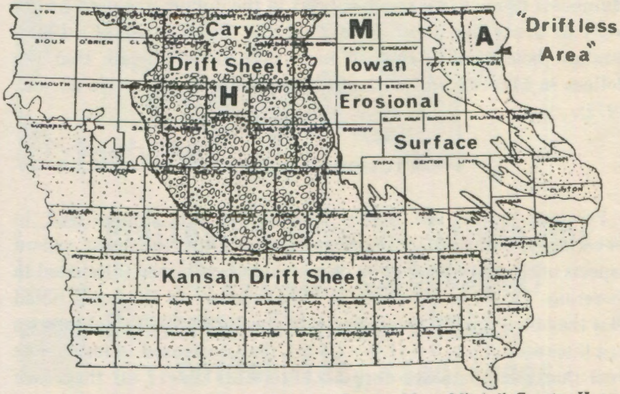

Fig. 1. Index map of Iowa. A = Allamakee County, M = Mitchell County, H = Humboldt County.

carbonates, sandstones, and shales (Hershey, 1969; Calvin, 1894) which have a homoclinal dip to the southwest (Trowbridge, 1934; Calvin 1894). The county is characterized by a rolling topography having about 400 ft (122 m) of relief along the major rivers. Soil data indicate that over 62% of the county has greater than 9% slopes (Runge *et al.*, 1970). Remnants of two peneplains may be preserved as broad, accordant summits (Trowbridge, 1921). Even though this region is called the "Driftless Area", patches of drift are found on the upland. The drift is considered to be of Nebraskan age and older than the dissection of the younger peneplain (Trowbridge, 1966). The county is blanketed with loess having an average thickness of 16 ft (4.8 m) on the upland and an age between 24,000 and 16,000 years BP (Ruhe and Scholtes, 1956).

Mitchell County is situated in northeast Iowa, on the Iowan Erosion Surface (Fig. 1). The county is underlain by carbonates of Devonian age (Hershey, 1969; Calvin, 1904), which dip to the southwest. The bedrock surface is eroded into a mature landscape, with about 200 ft (61 m) of relief (Hale, 1948). Bedrock is buried by Kansan drift varying in thickness from 0 to over 300 ft (0 to 91 m) (Calvin, 1904). The drift surface is characterized by broad, shallow valleys, which have a relief of about 130 ft (40 m). This surface, known as the Iowan Erosion Surface, was developed by accelerated slope erosion, 18,000 to 29,000 years ago (Ruhe *et al.* 1968). Soil data indicate that slightly over 2% of the county is in slopes greater than 9% (Runge *et al.*, 1970). Grant (1972) noted that dolines were developed in the upper part of the Coralville member of the Cedar Valley Formation and occur in a belt parallel to the Cedar River. Voy *et al.* (1972) noted that most of the dolines are formed in less than 25 ft (7.6 m) of overburden and that most are not actively growing. Those that are enlarging are not doing so by cavern collapse, but by the washing of fines into cracks in the limestone.

* Dept. Earth Science, Iowa State University, Ames, Iowa.
† Dept. Agronomy, Iowa State University, Ames, Iowa.

Humboldt County is in north-central Iowa, in the center of the Cary Drift Sheet (Fig. 1). The county overlies bedrock composed of Cretaceous sands and shales and Mississippian carbonates and shales (Hershey, 1969). The bedrock surface is eroded into a mature landscape, with approximately 200 ft (61 m) of relief (Palmquist and Bible, 1974). Bedrock is buried by 0 to 150 ft (0-46 m) of drift (Palmquist and Bible, 1974; MacBride, 1898), the upper portion of which is late Wisconsinan in age (Ruhe, 1969). The surface is characterized by a hummocky morainal topography and incomplete drainage (Kay and Graham, 1943; MacBride, 1898), which forms part of the Humboldt End Moraine (Ruhe, 1969). Soil data indicate that 2% of the county has slopes greater than 9% (Runge *et al.*, 1970). Dolines occur on the floors of some of the larger depressions in the moraine (Richlen *et al.*, 1961).

The karst in the three counties is actively developing. In Allamakee County, collapse dolines containing till outcrops and subsidence dolines with shallow holes were observed in the field. In Humboldt County, the swallow holes in the dolines are used as dry wells to drain the water from field tile (Richlen *et al.*, 1961). Swallow holes and change in form over time indicate that the dolines in Mitchell and adjacent Floyd counties are active (Grant, 1972).

Previous Studies

Previous studies on doline distribution are summarized in Sweeting (1973) and in Jennings (1971). Only the most salient aspects of these studies will be presented here. Clayton (discussed in Sweeting, 1973), in his study of "Shakeholes" in England, noted that they increased in frequency with increasing drift thickness up to a thickness of 6 to 8 ft (1.8—2.4 m), then decreased in number as drift thickness increases until at 30 to 40 ft (9—12 m) they were rare. He also noted that they occurred in groups along drainageways. A study by Ford (discussed in Sweeting, 1973), in the Mendips of England, found that 80.5% of the solution dolines occurred aligned along valley floors, whereas 13% occurred on valley walls aligned in low order tributary valleys and 6.5% occurred on uplands as isolated features. The frequency of dolines on the valley floors appeared to be inversely proportional to the gradient of the floor and ceased at slopes greater than 225 ft/mile (26.5 m/km). Most of the dolines occurred near the contact with the underlying shales. Ford (1964) classified the dolines into populations—mother and daughter, based upon presumed age relationship. Drake and Ford (1972) used nearest neighbor analysis to confirm Ford's (1964) earlier observation that two generations of dolines exist in the Mendips of England. They conclude that, for each randomly distributed "mother" doline, four adjacent "daughter" dolines exist to form a cluster.

Most of the studies dealing with the spatial analysis and morphometry of dolines are summarized by Williams (1972). A study of the spatial analysis of dolines by McConnell and Horn (1972) dealt with the probabilities of different distributions. They point out that, in an array of frequency distributions, if the mean equals its variance, it can be considered to be a random distribution. If the variance is greater or lesser than the mean, the array is more clustered than random or more regular than random, respectively. A more regular array is thought to result from a mix of random and spatial competition processes, which result in a net decrease in points, whereas more clustered than random arrays may result from quasi-random processes such as colonization or diffusion, although multiple random processes could yield clustered arrays. Random arrays result from single stochastic processes. They reject the hypotheses that (1) doline distributions could be the result of a single random process and (2) the development of one doline promotes the development of an adjacent doline. Rather, they accept the hypotheses that (1) the development of one doline decreases the potential for the development of an adjacent doline and (2) doline distributions are the result of two random processes: cavern collapse and solutional development. McConnell and Horn used a quadrate of one square mile, which they later reduced to 0.0247 square mile. They considered the appropriate level of significance for rejection of the null hypothesis as 0.20, in contrast to the more commonly employed levels of 0.05 or 0.01. The higher levels of significance, in their opinion, increase the likelihood of a type II error.

Methods

The distribution of dolines within each county was inferred from soil maps printed at scales ranging from two to four inches per mile. The depressions marked on the soil maps were assumed to represent dolines. On each county soil map, depression densities were contoured and that portion of the county containing depressions was outlined. The average doline density for the karst area within the county was determined by dividing the total number of dolines by the total number of sections (1 section = 1 sq. mi. = 2.50 km^2) containing dolines. A subsample was drawn from each county and the number of depressions per section was categorized by soil type. The distribution of depressions was compared to the available topographic map of the bedrock surface, the overburden isopach map, and to the geologic map of the county. The primary assumption, that depression symbols on the soil maps represent dolines, is considered valid for Allamakee and Mitchell Counties but subject to discretionary use in Humboldt County, as will be discussed later.

Model

The regression of average doline density within the karst area of a county against age of the surface till yields the rate of doline initiation through time. Not considered in the growth model is the rate of doline enlargement through time. The ommission of enlargement data could introduce an error, in that the rate of doline initiation could decrease after a critical doline density is achieved, and further karst development might be limited to doline enlargement.

The possibility of doline enlargement becoming more important than doline initiation as karst development progresses can be evaluated by the use of different growth models. The four models considered are: 1) linear model, wherein doline density (D) is assumed to increase at a steady rate with increasing time (T) ($D = a + bT$); 2) log age model, wherein doline density is assumed to increase at a decreasing rate with increasing time ($D = a + b \log T$); 3) exponential model, wherein doline density is assumed to increase at a slightly decreasing rate with increasing time ($D = a + T^b$); and 4) log density model, wherein doline density is assumed to increase at an increasingly rapid rate with increasing time ($\log D = a + bT$).

Another complication to the comparison of doline densities is inherent in the origin of subsidence dolines. In the subsidence doline, the rate of development is controlled both by the rate of carbonate solution and by the rate of removal of the insoluble, clastic overburden. Differences in doline density among the counties could, therefore, be the result of differences in the strength, permeability, and thickness of the overburden, as well as differences in solubility of the carbonate and duration of solution.

If subsidence dolines do not develop until the till thickness is less than a critical value, the development of subsidence dolines need not begin immediately after the deposition of the till. In Mitchell County, at least 15 to 20 ft (4.5—6.1 m) of till were removed during the interval of intense erosion from 29,000 to 18,300 years B.P., which produced the Iowa Erosion Surface (Ruhe, 1969, p. 98). It is, thus, possible that initiation of subsidence dolines did not occur until this time. A similar problem may exist for Allamakee County, but data are not available with which to evaluate it. To evaluate this factor, models of doline development for Mitchell County included both till age and erosion surface age.

To properly evaluate the influence of overburden upon doline initiation, detailed data on overburden thickness and permeability are required for selected areas of different doline densities and size. Such data are not available. The test employed herein is, at best, suggestive.

Results

Doline Distribution

The dolines in Allamakee County are confined to the southwestern quarter of the county (Fig. 2). There are 5117 depressions shown on the Alamakee County soil map (Scholtes *et al.* 1958). Densities of over 125 depressions per square mile ($48/km^2$) are present, although the average density of depressions within the karst area is 45 per square mile ($17.4/km^2$). Comparison of the depression distribution to the geologic map of the county (Fig. 3) indicates that 85 percent of the dolines occur in the outcrop area of the Middle Ordovician age, Platteville-Decorah-Galena formations (Table I), which are predominantly carbonates. The appearance of dolines in the underlying, St. Peter sandstone is the result of generalized boundaries on the geologic map, though some of the dolines could be subjacent, collapse dolines developed in the lower Ordovician, Prairie du Chien dolomites. The presence of dolines within the overlying, Upper Ordovician, Maquoketa shale could reflect either generalized map boundaries or the development of dolines in the basal, Elgin limestone member, which is 190 ft (58 m) thick (Steinhilber *et al.*, 1961). Most of the dolines within the Platteville-Galena outcrop area (Fig. 3) are probably developed in the 252 ft (77 m) thick, Galena dolomite. The 145 ft (44 m) of underlying Platteville-Decorah formations consist of 66 ft (20 m) of shale and 79 ft (24 m) of thin-bedded limestone and dolomites which yield little water to wells (Steinhilber *et al.*, 1961). The Galena dolomite, on the other hand, yields moderately large quantities of water to wells and large quantities of water to basal springs (Steinhilber *et al.*, 1961).

The majority (99%) of the dolines in Allamakee County occur on the uplands, with only one percent occurring on walls and floodplains of the major valleys (Table II). The ratio of "doline percent to soil extent", which is a measure of relative density, indicates that dolines are over represented in both the Fayette soil, and the Downs-Tama soils, but essentially absent in the other soils which together cover 54 percent of the county.

A field traverse down a shallow drainageway leading to a large, compound doline indicates that doline size varies inversely with distance from the major doline (Fig. 4). This relationship suggests that, first, dolines are concurrently enlarging and being initiated

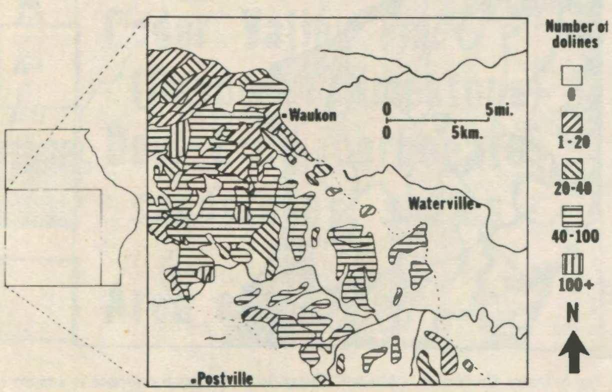

Fig. 2. Doline distribution in Allamakee County, Iowa. Densities are expressed in number of dolines per square mile.

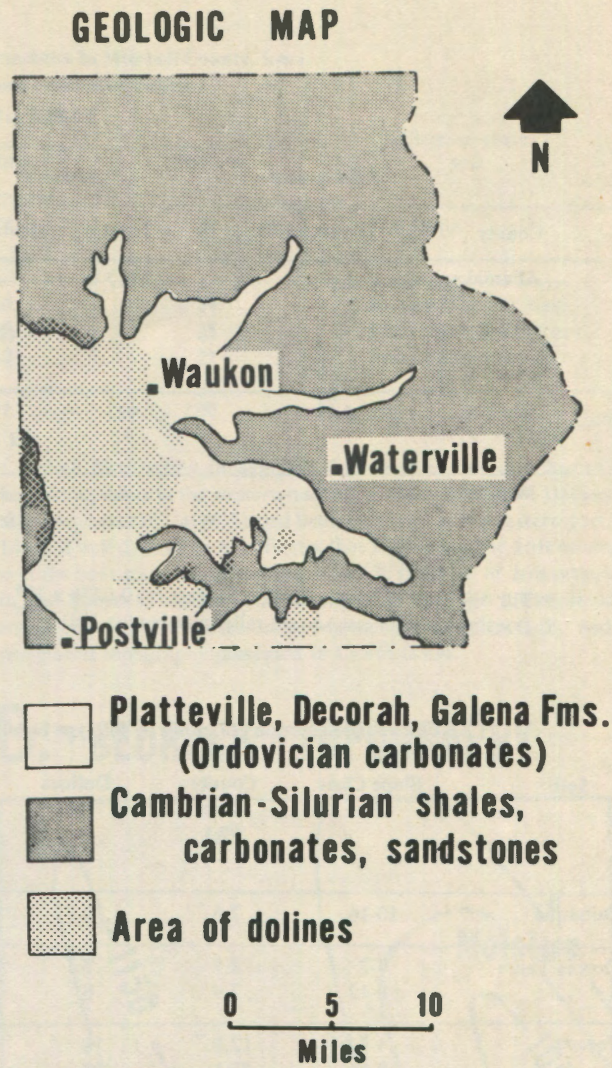

Fig. 3. Geologic map of Allamakee County, Iowa.

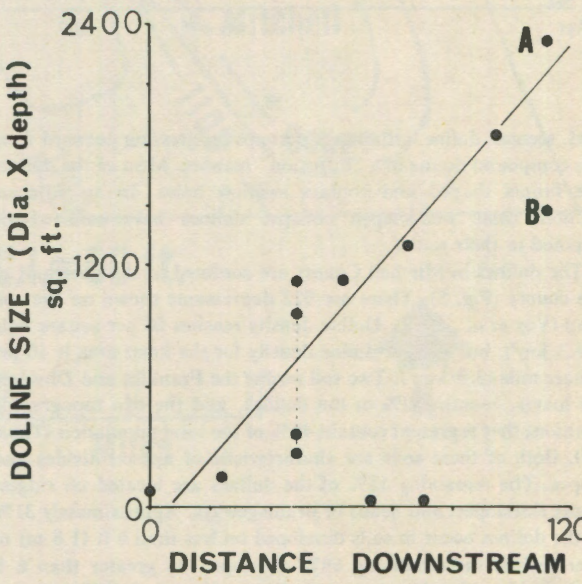

Fig. 4. Variations in doline size "downstream" along a shallow drainageway leading to a large compound doline, Allamakee County, Iowa. Broken lines represent dolines along small drainages tributary to traverse drainage. Compound dolines designated A'-B'.

TABLE I: County data.

County	Estimated till age (Years x 10^6)	Dolines: Total	Dolines: Mean density†	Formation	Number	Percent
Allamakee	1.8	5117	45	Maquoketa	307	6
				Galena	4349	85
			(17.3)	St. Peter	409	8
				Prairie du Chien	51	1
Mitchell	0.9*	922	10	Shell Rock	18	2
			(3.9)	Cedar Valley	904	98
Humboldt	0.013	83	4.6	Hampton-	83	100
			(1.8)	Gilmore City		

* Till Age: age of Iowan Erosion Surface = 0.0236 x 10^6
† Density: number per square mile (number/sq km)

TABLE II: Distribution of dolines by soil type for Allamakee County, Iowa (Soil data from Scholtes *et al.*, 1958).

Soil	Slope Class %	County extent (%)	Dolines (%)	Ratio (Doline % / Soil extent %)	Loess thickness (feet)	Topographic position	Vegetation
Dubuque	10-16	7.6	0.7	0.1	1-3	Ridges/slopes	Forest
Downs-Tama	4-7	4.7	7	1.5	10+	Ridge crest	Forest-prairie
	8-12	2.9	8	2.7	10+	Slope	Forest-prairie
Fayette	4-7	12.9	14	1.1	3-10	Ridge crest	Forest
	8-14	25.5	70	2.7	3-10	Slopes-floor	Forest
Other	—	46.4	0.7	0.0	—	Floodplains, Rocky slopes	—

and, second, doline initiation is perhaps progressing outward from the compound doline in a "diffusion" manner. Most of the dolines are funnel shaped and contain swallow holes. In an adjacent section, some well-shaped collapse dolines have oxidized till exposed in their walls.

The dolines in Mitchell County are confined to the west half of the county (Fig. 5). There are 922 depressions shown on the soil map (Voy *et al.*, 1972). Doline density reaches 50 per square mile (19.3/km^2), but average doline density for the karst area is 10 per square mile (3.9/km^2). Two soil series, the Franklin and Dinsdale silt loams, contain 30% of the dolines, and the two topographic positions they represent contain 48% of the total population (Table III). Both of these soils are characteristic of upland divides and slopes. The remaining 52% of the dolines are located on ridges, upper sideslopes, and heads of drainageways. Approximately 31% of the dolines occur in soils developed on less than 6 ft (1.8 m) of overburden; the remaining 69% developed on greater than 6 ft (1.8 m) of overburden, but, according to Voy *et al.* (1972), in less than 25 ft (7.6 m) of overburden. Data on the extent of each soil in the county are not available and doline percent cannot be compared to soil extent.

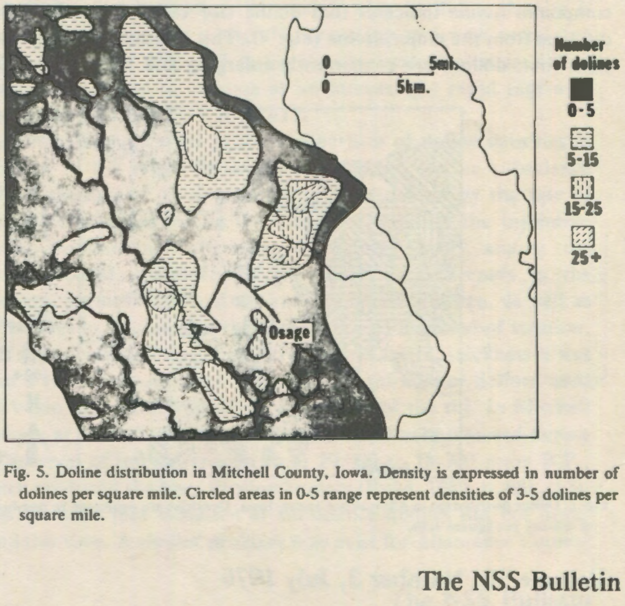

Fig. 5. Doline distribution in Mitchell County, Iowa. Density is expressed in number of dolines per square mile. Circled areas in 0-5 range represent densities of 3-5 dolines per square mile.

TABLE III: Distribution of dolines by soil characteristics in Mitchell County, Iowa.
Sample of 24 randomly selected sections containing dolines.

Overburden thickness	Topographic position	Percent slope	Dolines		Characteristic soil
			Number	Percent	
Less than 6 ft. (1.8 m)	Sideslope	2 - 9	74	30	Waucoma
Greater than 6 ft (1.8 m)	Drainageways	0 - 3	60	24	Klinger silty clay loam
	Sideslopes	2 - 9	65	25	Dinsdale silty clay loam
	Divides	1 - 3	52	21	Franklin silt loam
	TOTAL		251	100	

The geologic map (Fig. 6A) indicates that most of the county is underlain by the carbonates of the Devonian age, Cedar Valley formation. The majority (98%) of the dolines occur in the Cedar Valley formation (Table I). In adjacent counties, the Cedar Valley is over 200 ft (61 m) thick and consists of fine-grained dolomites, with some lithographic limestones (Hershey *et al.*, 1970). According to Grant (1972), the dolines occur in the Coralville member of the Cedar Valley.

Doline distribution in Mitchell County coincides with buried, bedrock uplands in the western half of the county (Fig. 6B). Dolines are absent over the large, buried bedrock valleys in the eastern part. Limited well data (N = 6) indicates that 15 ft (4.6 m) of drift occurs over the bedrock uplands and that over 125 ft (38 m) of drift occurs in the bedrock valleys. This distribution of drift thickness is consistent with the observations on outcrops, well depth, and pre-glacial topography presented in Calvin (1904).

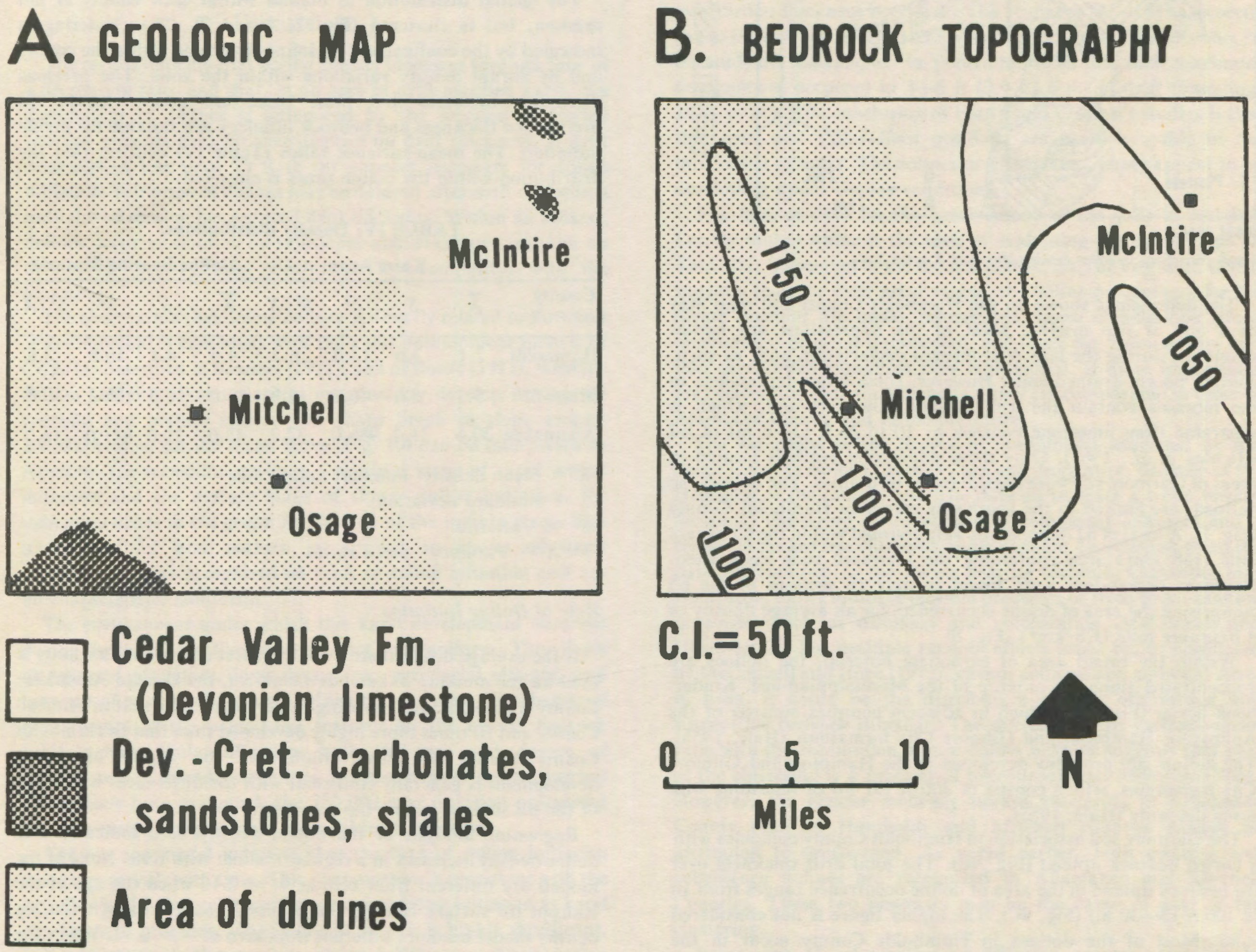

Fig. 6. Doline distribution compared to geology (A) and bedrock topography (B), Mitchell County, Iowa (base maps from Hershey, 1969; Hale, 1948).

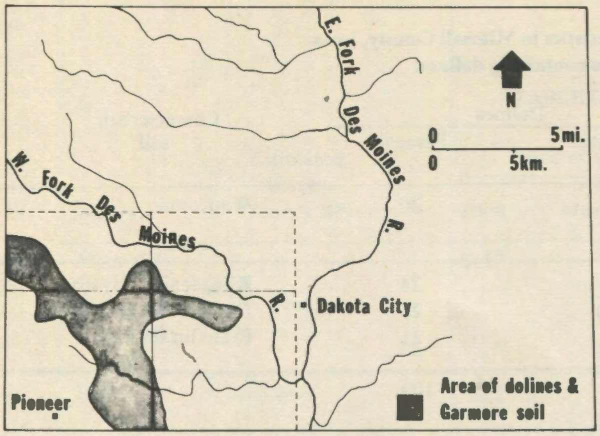

Fig. 7. Distribution of dolines and Garmore soil in Humboldt County, Iowa.

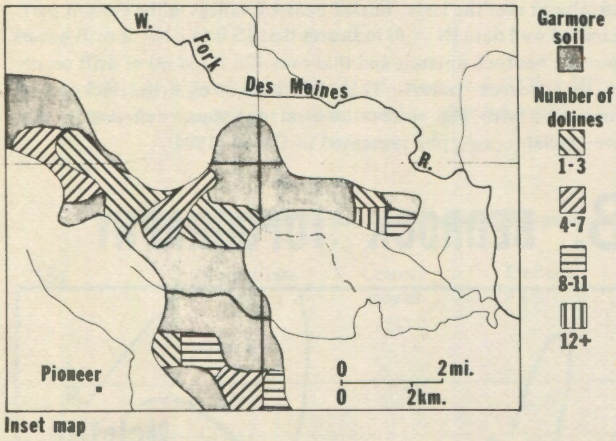

Fig. 8. Variations in doline density, Humboldt County, Iowa.

The depressions shown on the Humboldt County soil map may have one of two origins. Most of the depressions are kettles developed during the late Wisconsinan deglaciation and, as such, contain poorly drained soils. However, some large depressions in the moraine contain the well drained, Garmore soil, which is separated from limestone bedrock by 10 to 20 ft (3—6 m) of till (Richlen *et al.*, 1961). Only those depression symbols located in areas of Garmore soil were considered to be dolines. The dolines, so defined, are limited to the southwest corner of Humboldt County (Fig. 7). A total of 83 depressions were counted within the 65 square mile (168 km^2) area containing the Garmore soils association. However, depressions were found in only 18 sections of the 65 comprising the area of doline occurrence, for an average density of 4.6/square mile (1.8/km^2) (Fig. 8).

Within the broad area of carbonate outcrop, the dolines are concentrated along the outcrop of the Mississippian-age, Kinderhook Series (Fig. 9A), which in adjacent counties consists of the Aplington, Hampton, and Gilmore City formations (Hale, 1955). The dolines are probably developed in the Hampton and Gilmore City formations, which consist of 300 ft (91 m) of limestone with some dolomite (Hale, 1955).

The Garmore soil association in Humboldt County coincides with a buried bedrock upland (Fig. 9B). The total drift thickness over the bedrock upland in the area of doline occurrence ranges from 10 to 100 ft (3—30 m) (Fig. 9C). The higher figure is not considered valid. Most of the dolines in Humboldt County occur in the southern and eastern part of the Garmore distribution, where the drift is less than 25 ft (8 m) thick. Those dolines occurring in the central portion of the Garmore distribution follow a westward re-entrant of thin drift (Fig. 9C).

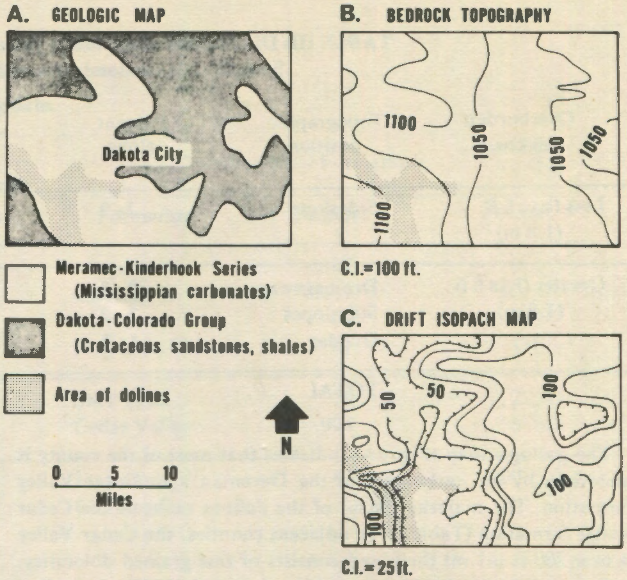

Fig. 9. Doline distribution compared to geology (A), bedrock topography (B), and drift thickness (C), Humboldt County, Iowa (base maps from Hershey, 1969; Palmquist and Bible, 1974).

The spatial distribution of dolines within each county is not random, but is clustered (Fig. 2, 4 and 7). The clustering is indicated by the confinement of dolines to a zone within the county and by abrupt density variations within the zone. The previous discussion indicates that these zones are areas wherein the overburden thickness and bedrock lithology are suitable for doline initiation. The mean-variance ratios (Table IV) confirm that the distribution within the doline zones is clustered.

TABLE IV: Density distributions.

County	Karst Area				Doline Bearing Sections			
	$\bar{X}$	S	S^2	S^2/X	$\bar{X}$	S	S^2	S^2/X
Humboldt	2.1	3.6	13.0	6.2	4.6	4.1	16.8	3.6
Mitchell	5.2	6.8	46.2	8.9	10.0	10.3	106.1	10.6
Allamakee	30.6	31.4	986.0	32.2	45.0	29.0	841.0	18.7

X = Mean density: number/square mile
S = Standard deviation
S^2 = Variance
S^2/X = Variance - mean ratio.

Rate of Doline Initiation

If the average doline density for the karst area within a county is used for the measure of karst development, the karst in Allamakee County is about 5 times more highly developed than that in Mitchell County and 10 times more highly developed than that in Humboldt County (Table I). This variation in the degree of doline development is generally consistent with the differences in the age of the till surface among the counties (Table I).

Regression analysis of the doline density data indicates that doline density increases in a regular manner with time. None of the models are different from zero at $\alpha = 0.10$ when the age of the Kansan till surface is used for Mitchell county, though the log density model becomes different from zero at $\alpha = 0.12$. When the age of the Iowan Erosion Surface is used for Mitchell County, the log age model (Equat. 1) becomes highly significant ($\alpha = 0.01$).

$$\text{Doline Density} = 18.8 \log \text{Age} - 72.3 \; (R^2 = 0.9998) \qquad 1.$$

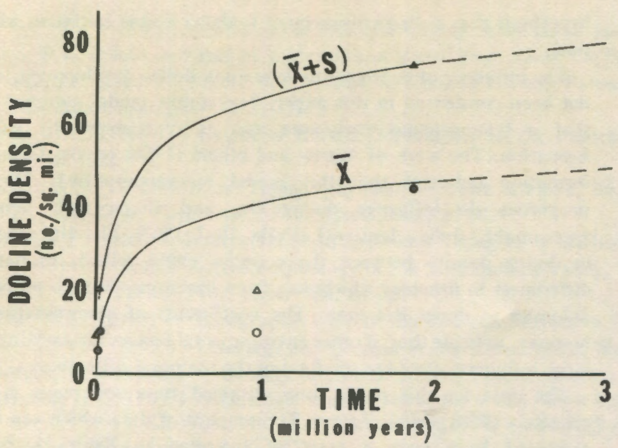

Fig. 10. Change in doline density with time. Closed circles indicate mean densities ($\bar{X}$); closed triangles indicate densities one standard deviation higher ($\bar{X} + S$). Open circle and triangle indicate Kansan age for Mitchell County.

The significance of the log age model (Equat. 1) when the age of the Iowan Erosion Surface is used, can be interpreted to indicate that doline initiation did not begin until the Kansan drift was thinned by erosion to less than a critical thickness. The shape of the log age curve (Fig. 10) indicates that the rate of doline initiation decreases abruptly after 250,000 years.

Discussion

The results indicate that doline density increases through time at a decreasing rate, and that, in an area of drift-mantled karst, the distribution of dolines is influenced by the location of buried bedrock uplands through their control on drift thickness. A drift thickness of less than 25 ft (8 m) appears necessary for doline initiation. Initiation of dolines may be delayed after drift deposition until fluvial erosion has reduced drift thickness. Within an upland, dolines occur more often on sideslopes and drainageways than on divides or crests. These conclusions are compatible with the literature.

The observation that doline density is directly related to duration of development is consistent with the karst development models of Cvijic (as described in Sanders (1921)) and of Beede (1911), wherein dolines and other karst features progressively develop during the youthful and mature stages. The sharp break in slope around 250,000 evident for the mean curve (Fig. 10) can be interpreted to represent the transition between a youthful stage of rapid doline initiation and the mature stage of slower doline initiation. No indication exists in the graph for the end of the mature stage. The terms youthful and mature as applied to stages of karst development were recognized on rates of doline initiation and not on characteristic landforms.

The environment under which this karst development occurred during the last two million years has not been uniform. The effects of repeated glaciation, loess deposition, periglacial climate, and intervals of warmer climate are difficult to evaluate. The degree of development of the paleo-karst features on the buried bedrock surface most likely influences both the rate and degree of development of subsidence dolines. Because none of these historical effects have been evaluated, the 250,000 year duration postulated for the youthful stage should be considered tentative.

The log age model indicates that the rate of doline initiation decreases rapidly with time. This phenomenon is consistent with the hypothesis of doline enlargement replacing doline initiation as karst development proceeds. The intercept, which is highly significant, indicates that, on the average, about 6,000 years elapse before doline initiation begins within a drifted-mantled area. The 6000-year interval between the time of formation of the land surface and the beginning of the first dolines may represent the time required for local thinning of the overburden and/or the initial flushing of clastics from the solutional cavities. Because mean density values are used in the regressions, the regressions may underestimate the initiation of dolines in favorable environments.

Doline development appears to be limited to the bedrock uplands in each county, though for different reasons. In Humboldt and Mitchell counties, their development is limited to bedrock uplands, because of the thicker drift occurring over bedrock valleys. In these two counties, dolines appear to occur in areas having less than 25 ft (7.8 m) of overburden. In Allamakee County, their limitation to the uplands is based more on lithology. Along the valley walls of the deeper, major valleys, the St. Peter sandstone crops out, and, in the tributary valleys, the Platteville-Decorah formations with their mixed shale-carbonate lithologies crop out. It is only in the low order tributaries and on the upland crests that the Galena formation, which is doline-prone, occurs.

The distribution of dolines within the uplands in Allamakee County appears to be controlled by the local topography and thickness of overburden, as suggested by the over representation of the Fayette and Downs-Tama soils in the 8-14% slope classes (Table II). This slope class is characteristic of low order tributaries, in which it occurs on both the sideslopes and valley floor. Floodplain soils are not developed in these valleys. Even though these soils tend to be developed on thinner overburden (Scholtes *et al.*, 1968), the contrast in thickness between the Downs-Tama and Fayette soils suggests that the valley position rather than the overburden thickness is critical. The tendency for dolines to occur in the Fayette soil is consistent with the finding of Clayton, who, in his study of "shakeholes" in Yorkshire, noted that their maximum development occurred in 3—6 ft (0.9—1.8 m) of drift (Sweeting, 1973, p. 60). The conclusion of Ford's and Clayton's studies is that thin drift and valley floor positions are most favorable to the infiltration of water. The dolines in Allamakee County appear to be developing under similar conditions.

The lack of data on the areal extent of the soils in Mitchell County makes difficult the task of evaluating the percentages in Table III. Inspection of the soil association map in Voy *et al.* (1972) suggests that, for their areal extent, the doline percentages for the divide position are under represented, those of sideslope (thick overburden) properly represented, and those for drainageways and thin overburden sideslopes over represented. If these impressions are correct, the data for Mitchell County are comparable to those of Alamakee County and of the literature.

The variations in the degree of randomness indicated by the mean-variance ratios (Table IV) are only suggestive, because of the large size of the sample cell. The data do indicate a trend toward increased clustering as duration of development increases and a tendency toward less clustered, more random distribution when the data are confined to the doline-bearing sections. Clustering may result from either multiple random processes or from quasi-random processes, such as diffusion and colonization (McConnell and Horn, 1972). The multiple random effects could be, as postulated by McConnell and Horn (1972), cavern collapse and solution. Also, in Iowa, it could be the distribution of thin overburden and susceptible bedrock lithology. The decrease in the variance-mean ratio with the elimination of non-doline sections suggests that part of the clustering is a product of overburden and lithic variations. However, the greater tendency toward clustering in Allamakee County than in Humboldt and Mitchell counties cannot be attributed to overburden. As noted earlier, both collapse and subsidence dolines are recognized in Allamakee and Mitchell Counties. These two processes may be the cause of part of the clustering.

Some type of colonization or diffusion process may also be operative in mantled karst areas. Overburden thinning occurs along drainageways in both Mitchell and Allamakee Counties. Doline initiation occurs in these drainageways. Stream spacing is not random, but regular, with the spacing being a function of the regional slope, soil permeability and strength, and climate (Hack, 1960, 1965). Thus, the distribution of drainage-thinned overburden should be relatively uniform, as should the resulting doline distribution, unless the degree of doline development within the drainageways is variable. The traverse along a shallow drainageway (Fig. 4) indicates that the size of the dolines increases toward the largest, compound doline. This data could be interpreted to indicate colonization or diffusion from the central doline. Perhaps, contrary to the conclusion of McConnell and Horn (1972) and consistent with that of Drake and Ford (1972), the initiation of one subsidence doline does favor the development of other, adjacent dolines.

The development of one subsidence doline could increase the capacity of the karst drainage system to accommodate and remove insoluble, clastic overburden. This increase would allow the development of an adjacent doline. An increase in ground water recharge through one doline would both promote the enlargement of the solution cavities and allow for the more complete flushing of the clastic overburden out of the system. This enlargement and flushing of the cavities would allow more overburden to sift in along joints and cause the initiation of additional subsidence dolines. Thus would occur a diffusion of new subsidence dolines away from the initial doline in the drainageway.

The diffusion process just hypothesized appears to be inconsistent with the log age regression model. The diffusion mechanism should promote an increase in the rate of doline initiation with time rather than the observed decrease in rate indicated by the regression model (Fig. 10). However, the diffusion mechanism is postulated to occur in the favorable environments to promote the observed clustering. The regressions might not therefore recognize this phenomenon because of the use of mean density values. To investigate this possibility, for each county the density values for one standard deviation greater than the mean ($\bar{x} + s$) were regressed against the age. The use of ($\bar{x} + s$) should approximate the greater densities developed in the more favorable environments. The use of high densities produced two regression models different from zero at $\alpha = 0.10$. Using a Kansan till age for Mitchell County, the log density model (Equat. 2) became different from zero at $\alpha = 0.07$. Substitution of the age of the Iowan Erosion Surface, made the log age model (Equat. 3) significantly different from zero at $\alpha = 0.04$.

$$\text{Log Density} = 0.90 + (5.2 \times 10^{-7}) \text{ Age } (R^2 = 0.987) \qquad 2.$$

$$\text{Density} = 2.97 \text{ Log Age} - 111.7 \, (R^2 = 0.996) \qquad 3.$$

The high density model produces a growth curve similar to the mean density curve (Fig. 10), in that the zero density intercept is around 5000 years and an abrupt decrease in the rates of doline initiation occurs around 250,000 years. However, the slopes of the high density and mean density curves diverge such that, even though a decrease in the rate of doline initiation occurs around 250,000 years, the difference in density between the high density and lower density sections continue to increase, that is, the clustering effect persists. In effect, the difference between the two curves represents the standard deviation. Thus, the curves illustrate the increasing spread of the density distribution as more dolines are initiated in some sections than in others. These curves are thus compatible with the diffusion hypothesis as well as with the hypothesis that doline enlargement replaces doline initiation with time.

The influence of bedrock lithology upon doline development has not been considered in this paper. The simple model assumed is that a homogeneous carbonate unit is represented by each formation. The work of Agnew and others (1956) on the Galena formation indicates that this model is oversimplified. These properties do influence doline size and shape and, hence (presumably) doline density (LaValle, 1967, 1968). The differences in doline density between the counties could reflect, in part, differences in lithology which vary from limestone through porous dolomite to dense dolomite. The coefficients of determination, however, indicate that, if other environmental factors are exerting a large influence, they are reinforcing the temporal differences.

The data for this paper were obtained from soil maps in a relatively short period of time. The amount of data which can be extracted from such a map is illustrated in Table II. Not considered, because of uniformity of these soils, is data on permeability, clay content, acidity, and drainage class, all of which are potentially useful in karst studies.

Conclusions

Doline densities in Northeast Iowa vary directly with time, as expressed by a log age regression model. Doline initiation begins around 5,000—6000 years after development of the land surface and proceeds at a high rate until around 250,000 years after surface development, at which time the initiation rate decreases. The density distributions are clustered, that is, greater densities develop in some areas than in others. The basic control on doline distribution and density in this area of mantled karst appears to be the thickness of overburden and, hence, the distribution of buried bedrock upland. The most favorable conditions for doline development are a thickness of overburden less than 25 ft (8 m) and those found in low order valleys. The rate of doline initiation in these favorable areas appears to be greater than elsewhere such that the clustering effect increases with time. This clustering may, in part, result from a diffusion effect, as the initiation of one doline promotes the flushing of the clastics from the solution cavities and thus the initiation of adjacent dolines.

This study did not consider all of the environmental influences both past and present, so that the results must be considered tentative. The study does indicate the potential usefulness of soil data in karst-related studies and of the value of dated surfaces in rate studies. It is hoped that this study will promote further investigations utilizing firmly dated land surfaces. It is realized the use of only 3 dated surfaces makes the interpretations based upon the growth curves highly speculative.

Acknowledgements

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Selected Abstracts from Recent Meetings

NSS Convention, Angels Camp, California, 27 June 1975

North American Climate During the Period 200,000 Years BP to Present, as Inferred from $^{18}O/^{16}O$ Variations in U/Th-Dated Speleothems

Russell S. Harmon* and Peter Thompson†

Variations in the $^{18}O/^{16}O$ ratios of speleothems formed under conditions of isotopic equilibrium record secular variations in the regional surface climate (temperature) above the cave. Equilibrium deposits dated by the U/Th method have been obtained from caves in Mexico, Bermuda, Kentucky, West Virginia, Iowa, and Alberta.

Axial profiles of twnety stalagmite and flowstone specimens exhibit cyclical patterns of $\delta^{18}O$ variation with minima interpreted as temperature maxima. Samples deposited over the same period of time within a single cave or region and between regions show sympathetic ^{18}O-time profiles. Periods of relative warmth are found across North America at 180,000, 145,000, 120,000, 100,000, and 60,000 years B.P., with less intense warm periods at 165,000, 82,000, 40,000, 22,000 and 10,000 years B.P. Readvances of the late Wisconsin ice sheet in the North American Midwest correlate well with the cold peaks of the Iowa speleothem record. The observed periods of warm climate correspond in time with the high seastands determined by Y/Th and U/Pa dating of raised coral terraces in Barbados and New Guinea, but show little agreement with the marine ^{18}O foraminiferal record.

* Dept. of Geology, Michigan State Univ., East Lansing, Mich. 48824.

† Dept. of Physics, Univ. of Alberta, Edmonton, Alta., Canada.

Mineralogy of Seven Great Basin Caves

Bruce Rogers§

The mineralogy of seven Nevada and California Great Basin caves has been investigated. Carbonates of calcium, magnesium, and iron, calcium and sodium sulfates, iron oxide, halite, and quartz have been found:

Crystal Ball Cave, Ut.: The usual calcite speleothems, as well as a nearly total covering of nail-head spar encrusted with gypsum; gypsum needles in an anauxite clay matrix between nailhead spar crystals; acicular aragonite crystals.

Goshute Cave, Nev.: Most calcite speleothem forms, including moonmilk; acicular to anthoditic aragonite crystals; massive gypsum corralloids.

Indian Burial Cave, Nev.: Calcite speleothems, including moonmilk, folia, and coralloids; acicular aragonite.

Lower Shoshone Cave, Calif.: Calcite corralloids and cave blisters; acicular aragonite crystal groups.

Snake Creek Cave, Nev.: Most usual calcite speleothems, including moonmilk, rhombic overgrowths on helictites, and pink and yellow coralloids; aragonite as small acicular crystals and up to 15 in. anthodites and stalactites; goethite as layers and micro-vugs in flowstone; quartz as layers with calcite and goethite flowstone.

Snowcap Cave, Calif.: Calcite flowstone and stalactites; acicular aragonite crystals; ankerite, dolomite and quartz crusts and flowstone layers.

§ U.S. Geol. Survey, Menlo Park, Calif. 94025 *and* San Jose State Univ., San Jose, Calif. 95125.

The Last, Best West

D.C. Ford**

The new South Nahanni River National Park of Canada extends for 180 miles, west to east, across the Mackenzie Mountains, the northern continuation of the Rockies. In most of Canada, landscapes are dominated by effects of the last glaciation, but, in Nahanni, the situation is more complex. In the West, are rugged mountains shaped by recent alpine glaciers from the West. In the East, front ranges were penetrated by the last Laurentide (lowland) continental ice sheet. In between, some ranges show evidence of older glaciations; some appear never to have been glaciated. These contain Canada's greatest river canyons and the most accentuated karst terrain known in the western hemisphere—"North Karst." It is a labyrinth ranging in scale from meters to kilometers and containing cenotes, fluctuating lakes, poljes, and karst towers. Caves tend to be short, being blocked by silt or ice. The few longer ones are well decorated and display striking microclimatic zonation.

** Dept. Geography, McMaster University, Hamilton, Ontario, Canada.

Mud Stalagmites and the Conulite: Discussion

Stewart B. Peck*

ABSTRACT

The term "conulite" was proposed by Thayer to name a hollow calcite speleothem from Morris Cave, Vermont. This paper enlarges the definition of the term to include any drip-drilled pit in sediments with walls which have been secondarily impregnated and, perhaps, lined by a mineral. The paper discusses some of the variation that might be expected in conulites and describes other conulites found in caves in Iowa and Florida.

Introduction

In his discussion of conulites and their mode of formation, Thayer (1967) defined conulites as "simple drip-drilled mud pits ... lined with calcite. Subsequent erosion then removed the mud isolating the lining in the form of a conulite." The definition should be broadened to include a wider range of sedimentary features. Thus, conulites may form in pits drilled in sand or gravel, as well as in mud and in any mixture of these clastic materials that does not collapse from the effect of water standing in the pit. Pits drilled in rock and drip basins which are wider than they are deep would expand the definition too greatly. Secondary mineral impregnation of the walls of the pit may be of calcite, but it is possible that aragonite, gypsum, or other materials could impregnate and cement the particles of the pit wall. The thickness of the conulite wall may be parchment thin, such as that in the Morris Cave structures, or it may grade into much thicker layers of secondarily impregnated material. An interior crystalline lining of varying thickness may be present. The conulite need not be exposed by later erosion, but only by removal of the surrounding unconsolidated sediments would the mineral impregnation of the lining of the pit usually be evident.

The depth and width of the original pit, before mineralization, will vary, depending upon the nature of the falling water drops and the nature of the cave floor. Water dropping from a higher ceiling has more drilling potential than water falling from a lower ceiling. The area of impact of the water drops is likely to be greater for water falling from a higher ceiling. This may result in a wider and more shallow pit. The rate of dripping controls the amount of time required to drill a hole. The composition of the clastic material will control the shape of the hole. Compact, fine clays will resist drilling more than sandy clays.

The above are purely physical factors. It is important to note that the pit is not created through the chemical process of solution. With these possible variations in mind, the term conulite can be seen to be applicable to any drip-drilled pit in sediments which has been secondarily impregnated and, perhaps, lined with a mineral coating.

Wm. B. White (pers. comm.) has pointed out that work by Kunsky and others in Czechoslovakia reveals a continuum from unlined drip pits in sediment to stalagmites. First comes the drip pit. Next come drip pits that have become calcite-lined, which may develop into nests of cave pearls. The stalagmite is erected on the conulite or nest of pearls. If one should uproot a stalagmite, one might discover at its base a conulite type of structure buried in the sediment underneath, sometimes including an encapsulated and cemented nest of pearls. The transition is in part chemical, related to the degree of supersaturation of the dripping solution. No one has tried to work it out.

The Weber's Cave Conulite

Weber's Cave is located 4 miles southeast of Dubuque, Dubuque County, Iowa, in Middle Ordovician Galena dolomite. It is typical of the caves reported on by Bretz (1938) and by Brown and Whitlow (1960). The cave was discovered by a shaft sunk by miners searching for galena, or lead ore. Intermittent, small-scale mining continued in the cave from about 1860 until the late 1890's. In the western half of the cave is a raise shaft, capped at the surface, near which is a widening in the cave passage used as a dumping area for dirt which was dug from other passages. The conulite was found in this dump. It was found broken into two pieces, representing the bottom and the middle portion. The top has never been found. The two sections were coated on the outside with laminated clay and sand.

Figures 1 and 5 show the exterior surface of the conulite after it was cleaned of clay. The horizontal bands are the sediment layers of the original cave floor. The bottom is rounded; particles of various sizes have been cemented into it. The height is 15 cm. Since the top has not been found, the original total height is not known. The average diameter is 3.5 cm.

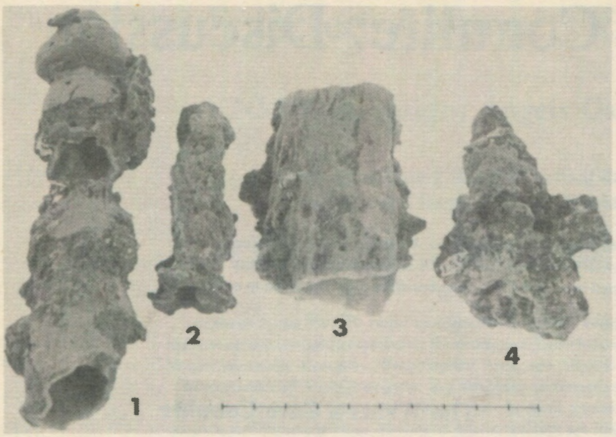
The conulite is composed (externally) of silt and sand particles cemented together, forming a cylindrical sheath 2 to 5 mm in thickness. Under this, lies an interior lining of scalenohedral calcite crystals. This calcite layer is uniformly 1 mm thick and is the light colored interior layer visible in Figure 1.

The conulite center is hollow and irregularly tubular, rather than gently tapering as in the Morris Cave specimens. The center is hollow for the entire 15 cm length. Since the place of formation of the conulite is not known, we do not know the height of the cave ceiling above it. Most likely, the conulite was formed close to the area in which it was found. In this section of the cave, the ceiling averages 2 m high and is 4 m at the highest. Thayer did not report the height of the ceiling above the Morris Cave conulites.

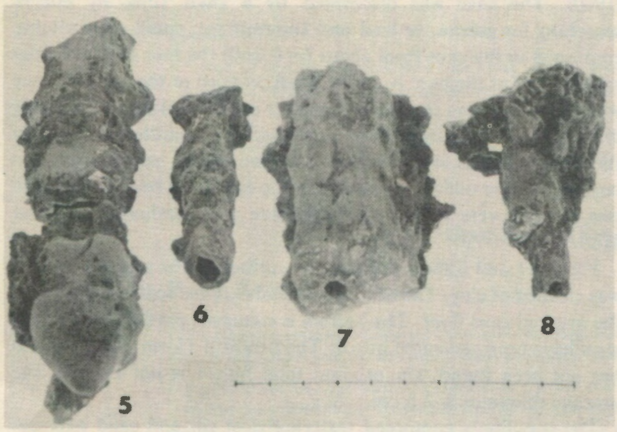
Thayer observed that 11 of the 23 observed drip-drilled pits were at least partially calcite lined. That all were not lined, he attributed to a small number of ions in the dripping water. He attributed the lack of ions to a relatively low solubility product of the dripping water, produced by the less-soluble dolomite contained in the limestone through which the water percolated. In Weber's Cave, the mining disturbances prevent us from taking a census of lined and unlined pits, but the lower solubility of dolomite is not responsible for the variation because Weber's Cave is formed entirely in dolomite and its conulite has a substantial calcite lining. Additionally, the cave possesses a profusion of both massive and delicate calcite speleothems.

In discussing the fact that no aragonite was found in the Morris Cave conulite linings, Thayer misinterpreted the findings of Moore (1956) by stating that "only calcite speleothems are found north of the 45°F isotherm." Actually, Moore observed that only calcite speleothems are now being formed north of the 60°F isotherm, and that aragonite and calcite speleothems are now being deposited

* Department of Biology, Carleton University, Ottawa, Ontario K1S 5B6, Canada.



Figures 1-4. Upper ends of conulites, showing external surfaces with small cemented stones and the central canals. 1. Weber's Cave. 2. Hunters Cave. 3 and 4. Hebron Ranch Cave. Scale line is marked in centimeters.



Figures 5-8. Lower ends of the conulites illustrated in figs. 1-4. 5. Weber's Cave. 6. Hunters Cave. 7 and 8. Hebron Ranch Cave.

south of this isotherm. Hence, Moore believed a temperature of 60°F or above to be required for aragonite deposition. However, caves between the 60°F isotherm and the 45°F isotherm contain aragonite formed in the past. Therefore, Moore believed these deposits to be indicators of higher cave temperatures in the past. Weber's Cave demonstrates that Moore's paleo-temperature hypothesis is imprecise. Aragonite is actively being precipitated in Weber's Cave, which lies in the vicinity of the 47°F isotherm. Temperatures measured in the cave agree with this isotherm.

The Hunters Cave Conulite

Hunters Cave is located near Andrew, Jackson County, Iowa, in Silurian Niagara dolomite. It is not known in what region of the cave the conulite was found, but it was formed in a silt and dolomite-sand cave floor. The specimen (figs. 2 and 6) is the upper portion of the conulite, 7.5 cm long, and about 2 cm wide. Its total length is unknown, because the bottom is missing. The walls are about 4 mm thick, with the inner 1 mm being a calcite-crystal lining. The central canal averages 7 mm in diameter, but is reduced to 5 mm where it opens to the top of the specimen. The top appears such that it could be mistaken for a drain hole in a thin flowstone sheet. Hunters Cave has not been intentionally studied for the presence of other conulites or drip holes.

The Hebron Ranch Cave Conulites

Hebron Ranch Cave is located near Brooksville, Hernando County, Florida. The conulites (figs. 3, 4, 7, 8), collected by E.J. Marcin, were formed in a clay- and sand-floored cave passage. The more cylindrical specimen (figs. 3, 7) is 7 cm long and varies from 3 to 4 cm wide. The central tube is about 2.5 cm wide at one end and 5 cm at the other. The calcite lining is about 2 mm thick at the wide end and from 7 to 20 mm thick at the narrow end. Extensions on either side of this specimen are casts of calcite-filled mud cracks that developed after (or during) the drilling of the original pit. A second specimen (figs 4, 8) is more nearly conical, with a length of about 7 cm, an upper diameter of about 4 cm, and a lower diameter of 1 cm. The central canal is open only in the lower 3 cm. The canal is filled in its upper portion with sand and rock flakes, cemented by calcite. The calcite lining is 2 mm thick at the lower end and averages 8 mm at the upper end. The tube filling does not seem to have infiltrated and cemented any of the sediments in which these two conulites were formed, except for small rocks attached to their sides.

Conulite and Cave Sediment Conservation

In closing, I would like to note that the above specimens of conulites came to my attention only because they were exposed in cave sediments that were dug up for purposes, and by people, unknown to me. Excavating in search of such structures would be a violation of NSS cave conservation principles and would not be a justification for digging in cave sediments. Any conulites existing in undisturbed cave sediments should be left where they are, just as any other type of speleothem.

All specimens described in this report have been deposited in the Division of Petrology and Vulcanology, Smithsonian Institution, Washington, D.C.

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Observations on the Mating Behavior of the Gray Bat and of the Eastern Pipistrelle in Northwestern Florida

David S. Lee *

ABSTRACT

Field studies of fall mating of *Myotis grisescens* and of *Pipistrellus subflavus* in northwestern Florida are described and discussed. Observations include time, season, location, and behavior of reproductive activity. The complete reproductive sequence was not witnessed for either species, but observed activity appears similar to that reported for other vespertilionid bats.

On four occasions between October 1969 and November 1973, I witnessed various aspects of the mating behavior of the gray bat, *Myotis grisescens*, in Florida Caverns State Park, Jackson County, Florida. Few field observations on copulation of bats and, apparently, no such information on the gray bat, have been published.

On 18 October 1969, Richard Franz, James Stevenson, and I took an inventory of the bats in Old Indian Cave. We noticed several torpid adult *Myotis grisescens*. These bats, which typically roost in groups in holes in the ceiling, were hanging from the walls of the cave. The fur in the area of the neck and upper back of these bats was wet and matted to such an extent that the skin was partially exposed. Examination, during which the bats remained torpid, showed that all were females.

Several mating pairs were observed through this same section of the cave. The males were mounted in the position of coitus *a posteriori* and, unlike the females, were active and usually departed as soon as our lights shone upon them. On two occasions, we were able to capture a pair and confirm the sexes of the individuals. In each case, the fur on the nape of the female was wet, having apparently been chewed by the male. Wimsatt (1945) reported that, when mating, male *Myotis lucifugus* "... grasped with his teeth the hair of the female at the base of her skull and pulled her head far back, usually at a right angle to her body." Although complete sexual contact was not observed in *M. grisescens*, the mounting aspect appears similar to that described by Wimsatt. The fur around the vaginal opening was also moist on most of the individuals I examined. One pair remained together for at least 15 min; on the final check, the bats were no longer together but were hanging side by side. Several pairs (sexes not confirmed) were seen flying through the cave, one individual following closely behind the other. Audible vocalizations were emitted by one or both. These observations were made between 1230 and 1400 (CST); the air temperature in the section of the cave used by the bats was 18°C.

The next morning, Gordon Persohn and I entered the same cave to make additional observations (0800-1015). Ten mating pairs were counted. Ten to fifteen additional, single, bats were noted that were not in typical roosting sites. Of the nine bats examined in this second group, only one was male. All of the females were torpid and had wet areas on their backs. One pair was observed in which the male's wings were flexed to form a canopy over both animals. Seconds later, the presumed male (sex not confirmed) flew away, leaving a wet-naped, sluggish female behind.

For the most part, the copulating pairs and recently-mated females occurred singly, but we observed one aggregation of five pairs. All activity took place in one area that was also used as a roost. Most individuals were four to ten feet from the floor of the cave, on walls undercut at approximately a 45° angle.

I made similar, though less detailed, observations in Old Indian Cave and in two other caves in Florida Caverns State Park on 16 October 1971, 19 October 1972, and 20 November 1973. These other caves are not normally inhabited by gray bats and, perhaps, are used only during the mating season.

Several pairs of eastern pipistrelles, *Pipistrellus subflavus*, were also found mating in these caves during this study period. Notes taken on this species indicate that they use the same section of the cave as do the gray bats, but that most pairs were on ceilings and several were in low passages less than two feet in height. Unlike the *Myotis*, pipistrelle males as well as females were sluggish and not easily disturbed. One male was observed with its forearms tucked under the head of the female.

Apparently, Old Indian Cave was a major hibernaculum for both species until the late 1950's and early 1960's. By the late 1960's, this was no longer a principal hibernating site (Lee and Tuttle, 1970). Tuttle (1974) commented on the migratory patterns of the Florida population of *M. grisescens* and searched without success for a second wintering site. Based on this and on his band returns, he believed that most of this population now winters in northern Alabama. Thus, it appears that at least some of the observed mating activity may have taken place just prior to migration and hibernation. Season and locality of the *M. grisescens* mating is of particular interest, since this peripheral population in northwestern Florida is greatly depleted and additional observations from this area may be difficult to obtain.

Guthrie (1933) noted that, based on the occurrence of sperm in the uterus, copulation in Missouri populations of *M. grisescens* occurred in late fall, with some females containing sperm as early as October.

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* North Carolina State Museum of Natural History, 101 Halifax St., Box 27647, Raleigh, North Carolina 27611.

INFORMATION FOR CONTRIBUTORS TO *THE NSS BULLETIN*

Papers discussing any aspect of speleology are considered for publication in ***The NSS Bulletin***. We particularly welcome articles describing important caves and cave areas, on the history of caves and of speleology, on problems and techniques of cave conservation, and critical reviews of current literature, in addition to papers on the more traditional subjects of cave geology, geography, anthropology, fauna, and ecology. The material presented must be original and of lasting interest. Authors should demonstrate the significance of their work to speleological theory and should elucidate the historical antecedents of their interpretations by reference to appropriate literature. Presentations consisting of raw data, only, will not be accepted.

A narrative style of writing is preferred. Fine prose is terse yet free from lacunae, sparkles without dazzling, and achieves splendor without ostentation. Data and interpretations blend effortlessly along a logical continuum so that the reader, having read, neither knows nor cares how many pages he may have turned while following the author's exposition.

As written language must communicate through time as well as across space, neologisms should be introduced only if needed to express new concepts or to record new percepts. Standard usage, therefore, is required of all authors. For general style, refer to papers in this *Bulletin* and to the following handbooks: "Suggestions to Authors" (U. S. Geological Survey), "Style Manual for Biological Journals" (American Institute of Biological Sciences, Washington, D. C.), and "A Manual of Style" (The University of Chicago Press).

Articles on earth sciences (including pseudokarst), life sciences, conservation, social science (including history), and exploration should be sent directly to the appropriate specialist on the Board of Editors (see masthead); articles not clearly falling into any of those categories may be sent to the Managing Editor. Potential contributors, especially those not professional scientists or writers, are invited to consult with the editors for guidance or aid in the presentation of their material.

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CONTENTS

PAPERS TO APPEAR IN LATER ISSUES

OGLE CAVE SYMPOSIUM

MINERALOGY OF OGLE CAVE — Carol A. Hill

MAP — John Corcoran III and J.M. Hardy

KARST HYDROLOGY IN THE UPPER ELK RIVER BASIN, WEST VIRGINIA — Douglas M. Medville

INDIANA AND KENTUCKY KARST SYMPOSIUM

GEOMORPHOLOGY AND HYDROLOGY OF THE INDIANA AND KENTUCKY KARST — A.N. Palmer and M.C. Moore

THE MITCHELL PLAIN OF SOUTHERN INDIANA — Margaret V. Palmer

INVESTIGATIONS OF SINKHOLE STRATIGRAPHY AND HYDROGEOLOGY, SOUTH-CENTRAL INDIANA — Robert D. Hall

KARST AND CAVE DISTRIBUTION IN NORTH-CENTRAL KENTUCKY — Angelo I. George

A REVIEW OF THE HYDROGEOLOGY OF THE CENTRAL KENTUCKY KARST — John W. Hess

BURNSVILLE COVE SYMPOSIUM

HYDROGEOLOGY OF THE DRAINAGE SYSTEM, BURNSVILLE COVE — Nevin Davis and Jack Hess

GEOCHEMISTRY OF THE CAVE WATERS — Russ Harmon and Ike Nicholson

HISTORY AND EXPLORATION — Fred Wefer and Ike Nicholson

PRELIMINARY REPORT ON THE CAVE FAUNA OF BURNSVILLE COVE — John Holsinger

POSSIBLE EXHUMED FOSSIL CAVERNS IN THE MADISON GROUP (MISSISSIPPIAN) OF THE NORTHERN ROCKY MOUNTAINS — Newell Campbell

CHARACTERIZATION OF KARST SOILS BY NEAR INFRARED SPECTROSCOPY — William B. White

ALTERNATIVES TO THE LETTER-WRITING APPROACH IN CONSERVATION — Gene Hargrove

A CAVE POPULATION STATISTICAL STUDY — John A. French

RELATIONSHIP BETWEEN THE HYDROLOGY OF BLOWING CAVE AND COWPASTURE RIVER — Philip C. Lucas

INVERTEBRATE FAUNAS OF CAVES TO BE INUNDATED BY THE MERAMEC PARK LAKE IN EASTERN MISSOURI — John L. Craig

SURVEYING ERRORS AND MULTIPLE LOOP CLOSURES — William S. Hawes

CAVE COÖRDINATE CALCULATIONS WITH PROGRAMMABLE CALCULATORS — Bill Douty and Alan Armstrong