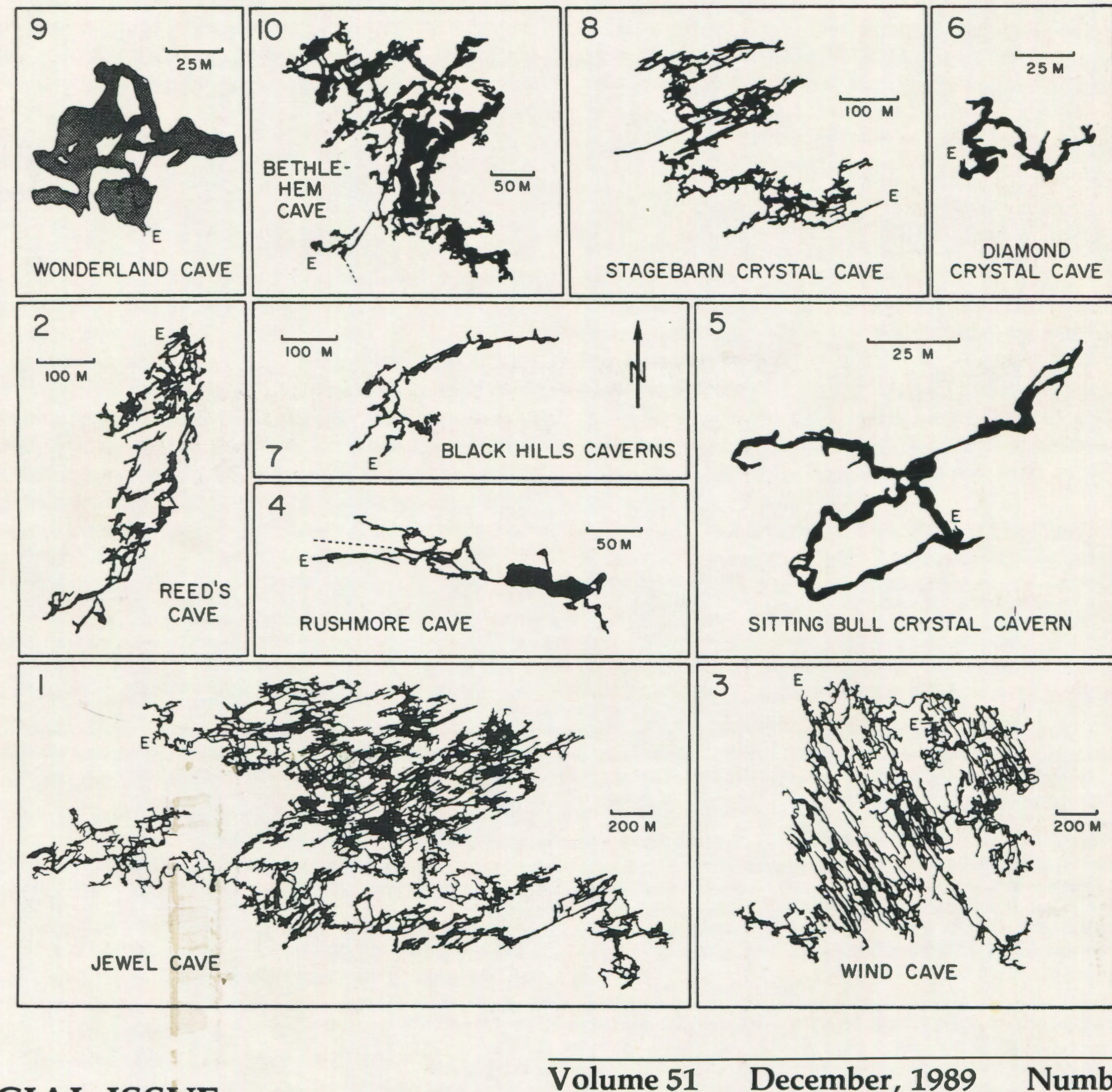
THE NSS BULLETIN

Journal of Caves and Karst Studies



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Cover: See Palmer and Palmer, page 72.

INTRODUCTION

In May of 1984 I arrived at Wind Cave National Park to begin work as a seasonal park ranger. This was not my first trip to Wind Cave. In 1975 I took my first trip off trail into that ultimate maze of passages. For 12 hours I wandered around totally lost trying to keep up with those who were leading the trip. That day I was happy to see the surface. I have, to this day, never felt so lost in a cave.

The Rocky Mountain Regional in 1977 was my first caving venture into Jewel Cave. With care and love, Herb and Jan Conn guided the group I was with through the "Hub Loop." Again I was amazed by the size and complexity of the caves of the Black Hills.

With these thoughts in mind, I walked into the Visitors' Center at Wind Cave and introduced myself to the ranger at the desk. I was quickly whisked into the office and deposited on the threshold of Assistant Chief Naturalist Kay Rohde's office. On that day who would have guessed at all the changes, studies, and adventures that would follow in the next five years.

To the average caver the most notable change was the explosive growth in the explored and surveyed sizes of both Wind and Jewel Caves. Jewel Cave, fourth longest cave in the world, went from approximately 116 km in length to 124 km. In the same period, Wind Cave went from ninth longest cave in the world at 62 km to seventh longest cave at 84 km.

But more noteworthy and significant changes were happening behind the scenes at both Wind and Jewel Caves. In the intervening years between 1984 and 1988, the National Park Service actually got serious about managing the world class cave resources under its jurisdiction which Wind and Jewel Caves represent.

The impetus behind this change was Kay Rohde. In 1984 Kay finished the first of what would be many drafts of a Cave Management Plan for Wind Cave. It is actually difficult to imagine that Wind Cave, the first cave anywhere in the world to be designated as a National Park (3 January 1903), had no designated plan for its management. Once that initial hurdle was crossed there existed a game plan for organizing activity not only at Wind Cave National Park but at nearby Jewel Cave National Monument.

With Kay Rohde's recently completed cave management plan in hand, new park Superintendent Ernest Ortega literally jumped into cave management feet first. Within weeks of his arrival the cavers on the staffs of both Wind and Jewel Caves had him and the other chiefs of the divisions in the park down in the cave to see what the resource was all about. Superintendent Ortega was the first park superintendent to actually travel into the remote areas of Wind Cave.

With a sympathetic ear in the superintendent's office, a plan specifying a direction, and Kay Rohde's knowledge of where one should go with managing a cave, the first federally funded cave research project in Wind Cave was begun in 1985. That project concerned the hydrology of Wind and Jewel Caves and was awarded to Calvin Alexander at the University of Minnesota.

Calvin Alexander brought his graduate student Marsha Davis onto the scene and together with all the other people such as Art and Peg Palmer, Tom Miller, Derek Ford, Jim Martin, and a host of others who were doing research in Black Hills caves there was a lot of activity with regard to the caves in the Hills.

Then on 10 June 1987 John McLane came up to Wind Cave to take a short trip down to the lakes to get a water sample. I was along on that trip to lead John through the maze to Calcite Lake and to make sure that no one would get lost.

Like all such trips, there was lots of crawling, climbing, and other caving stuff going on including the camaraderie that always seems to flow from caving trips. One thing lead to another and John made a comment that it was a shame that the National Parks all seemed to have lots of "really neat stuff" in their files that no one seems to know about. I agreed with him, but John's idea stuck and it seemed that all that "really neat stuff" really should be out there for all interested cavers and cave scientists to look at. Seeing that the NSS Convention was to be in the Black Hills, I wrote to everyone I knew who was doing or had done some research in Black Hills caves. That symposium and this volume that you are now reading is the result of that effort.

Good cave (researching),
Jim Pisarowicz

Palmer, Arthur N. and Palmer, Margaret V.—Geologic History of the Black Hills Caves, South Dakota: *NSS Bulletin* **51**: 72-99.

GEOLOGIC HISTORY OF THE BLACK HILLS CAVES, SOUTH DAKOTA

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The origin of caves in the Black Hills is intimately linked to the diagenetic and erosional history of the Pahasapa Limestone. Beds of gypsum and anhydrite were deposited contemporaneously with the limestone, particularly in the middle dolomitic beds. Mobilization, hydration, recrystallization, reduction, and solution of the sulfates caused fracturing and brecciation of the surrounding rock in irregular zones, some nearly vertical and others sub-concordant with the bedding. Most of the remaining sulfates were replaced by calcite and quartz, producing competent calcite veins surrounded by crumbly quartz-rich bedrock. A late Mississippian karst surface developed on the Pahasapa, and caves formed in the mixing zone between fresh water and underlying saline water. The caves consisted mainly of irregular vaulted rooms, although some were vertical fissures that connected with the overlying karst surface. Surface depressions, as well as most of the caves, were filled with clay and quartz sand early in the Pennsylvanian Period.

The present caves were formed during the Tertiary Period, probably by artesian groundwater, but there is also evidence for cave origin by rising thermal water. Solution may have been enhanced by mixing with low-CO_2 water infiltrating under closed vadose conditions through the sandstone caprock. The Tertiary caves intersect both the ancient sulfate zones and the paleokarst. Many of the upper levels are exhumed and enlarged paleo-caves, and most lower-level passages follow former sulfate zones. Much of the present cave pattern is therefore inherited from Mississippian features. The network outline of the caves is closely related to the stresses within the uplifting Black Hills during the late Cretaceous and early Tertiary Laramide orogeny. Yet the basic outlines of the caves were determined long before, during the Mississippian Period. Apparently many of the present fracture zones were formed by minor tectonic unrest at that time and were merely reactivated during the Laramide.

INTRODUCTION

Caves in the Black Hills of South Dakota have perhaps the most complex evolutionary history of any caves in the world. This paper concentrates on the two largest, Wind Cave and Jewel Cave, but the topics discussed apply as well to nearly all other caves of the Black Hills. Jewel Cave, currently fourth longest in the world, contains 131 km of surveyed passages, and Wind Cave contains 85 km. These are multi-level network caves in the Pahasapa Limestone (Madison Limestone equivalent) of Mississippian age. Not only do the caves have extraordinarily intricate patterns, but they have experienced a bewildering variety of geologic events, all of whose effects can be seen superimposed on one another in the caves. Elsewhere in the world, most large caves owe their complexity merely to successive passage development at lower elevations as the water table drops.

The geologic history of the caves includes five major stages, each of which had a significant influence on the present cave character:

1. Brecciation and secondary calcite mineralization of the Pahasapa Limestone during the middle Mississippian Period by migration, hydration, solution, reduction, replacement, and recrystallization of sulfates within the bedrock soon after it was deposited.
2. Late Mississippian exposure of the Pahasapa to circulating fresh water, resulting in an extensive karst surface with underlying caves.
3. Burial of the Mississippian karst and caves by sedimentary rocks of Pennsylvanian through Cretaceous age, which preserved a zone of paleokarst at and near the Mississippian/Pennsylvanian boundary, followed by minor calcite, hematite, and quartz mineralization during deep burial.
4. Uplift of the Black Hills during the late Cretaceous and early Tertiary Periods, which exposed the Pahasapa to the circulation of solutionally aggressive water and resulted in the major phase of cave development.
5. Periodic draining and reflooding of the caves during the late Tertiary and Quaternary Periods, including a long and complex history of subaerial weathering. Thick carbonate sediment accumulated, boxwork fins were exposed, and speleothems were deposited both below and above the water table.

The resulting caves contain features that span more than 300 million years. Although the overall geologic history is clear, several major questions remain. Surprisingly, it is stage 4, the main phase of solutional cave origin, that creates the most controversy. Where did the water come from that formed the caves? What was its chemistry? When was the major cave-forming event? These and other questions are considered at length in this paper.

PREVIOUS WORK

In the mid-1890s, shortly after Wind Cave was discovered, geologist Luella Owen visited the cave fresh from a trip to Yellowstone National Park. Her observations (Owen, 1898) were deeply colored by her recent experience, as she unhesitatingly interpreted the cave as an extinct geyser vent. Offering a more traditional viewpoint, Darton (1918) attributed the Black Hills caves to solution by artesian groundwater in the Pahasapa Limestone. The first thorough study of the Black Hills caves was that of Tullis and Gries (1938), who described the basic geologic setting and mineralogy of the caves and strengthened the argument for an artesian origin. During the mid-1900s several bachelor's theses by students at the South Dakota School of Mines involved the mapping and geologic interpretation of short segments of Wind Cave. The work of Deal (1962, 1968) in Jewel Cave for his master's thesis did much to unravel the geologic complexities of the cave. He recognized most of the cave deposits and was able to establish a clear sequence of events. He found no evidence for a hydrothermal origin. However, White and Deike (1962), on the basis of crystalline hematite and quartz found in Wind Cave, suggested that mineralization temperatures were several hundred degrees Celsius, thus reviving the view that the cave was formed (or at least modified) by hot water. Howard (1964) presented a conceptual view of the hydrologic behavior of the various parts of the Pahasapa aquifer and attributed the known parts of the present caves to artesian flow. Loskot (1973) made a petrographic study of the paleofills in Wind Cave, and Roth (1977) examined dolomitization of the Pahasapa Limestone in Jewel Cave.

Bakalowicz et al. (1987), including the present authors, gave isotopic, geomorphic, petrologic, and geochemical evidence in favor of a thermal origin for Wind and Jewel Caves. Millen and Dickey (1987), using similar isotopic data, consider the present waters in Wind Cave to be part of the regional artesian system. In a study of Wind Cave, Miller (1988) proposed that Wind Cave formed by artesian groundwater flow, but that the deposition of calcite wall crusts took place under thermal conditions. Evidently the thermal debate is far from closed.

Our own study began in 1978. At the request of the National Park Service and Wind Cave/Jewel Cave Natural History Association, three interpretive booklets have resulted from that work (Palmer, 1981, 1984, 1988), which show the evolution of our thinking as field evidence accumulates. The first attributed cave origin to fluctuations in artesian groundwater flow, the second placed greater emphasis on the role of thermal water, and the third gave nearly equal weight to thermal and artesian origins. Since then our interpretation has evolved back in favor of an artesian cave origin.

Geologic mapping and petrographic analysis have progressed enough now that the ideas in this paper are substantiated by many different lines of evidence. Still, this is only a progress report. Specific topics, mainly the petrography and geochemistry, will be covered in more detail in future publications.

FIELD AND LABORATORY PROCEDURES

The first step in the project was to map the stratigraphic column in surface canyons near Wind and Jewel Caves, as well as in the caves themselves. Geologic mapping of the main cave passages is still in progress. Samples were obtained from each bed, with a vertical interval of no more than a meter, from which petrographic thin sections were prepared. Staining was done for calcite, dolomite, gypsum, and ferroan content. Mineral identification was verified with X-ray diffraction. Scanning electron microscopy and EDX analyses of certain samples were performed at the University of Illinois by Rick Olson. Isotopic analyses and radiometric dating of calcite samples were done at McMaster University by Derek Ford.

Most of the geologic interpretation is based on observations with light-transmission microscopes at magnifications up to 630X. This simple tool is essential in determining the relationships among the many stages of cave evolution. Knowledge of the composition, interrelationship, and genetic sequence of minerals has made it possible to reconstruct the geochemical evolution of the water that was in contact with the minerals throughout the geologic history of the region.

GEOLOGIC SETTING

Stratigraphy

The major caves of the Black Hills are located in the Pahasapa Limestone of early Mississippian age, which varies in thickness from 100 m at Wind Cave to 140 m at Jewel Cave and more than 190 m in parts of the northern Black Hills (Figs. 1 and 2). The Pahasapa dates from the Kinderhookian, Osagian, and early Meramecian Stages (350–335 million years ago) and is correlative with the Madison Limestone of Wyoming and Montana, the Leadville Limestone of Colorado, and the Redwall Limestone of Arizona.

The Pahasapa is underlain by 10–20 m of shaly, dolomitic Englewood Limestone (Devonian-Mississippian) and 10–120 m of Deadwood Sandstone (Cambrian). The Deadwood lies uncomformably on Precambrian metamorphic and igneous rocks. Overlying the Pahasapa is the Pennsylvanian-Permian Minnelusa Formation (100–250 m thick), mainly sandstone and shale with interbeds of limestone and dolomite, whose basal beds fill paleokarst features in the upper Pahasapa. The Minnelusa forms a moderately permeable cap that transmits groundwater recharge to, and discharge from, the underlying Pahasapa. In the plains surrounding the Black Hills the Minnelusa is overlain by more than 2 km of mainly detrital sedimentary rocks that range in age from Permian to Tertiary, but erosion has removed them from the areas directly above the caves.

The Pahasapa Limestone consists of three distinct units of nearly equal thickness (Fig. 3). The lower and middle units correlate with the Lodgepole Member of the Madison Limestone farther west, and the upper unit correlates with the Mission Canyon Member. The lowest unit consists of massive, prominently fractured dolomite with fossil-moldic porosity. The middle unit consists of prominently bedded dolomitic limestones and dolomites, with beds varying from 20 cm to 3 m thick. This unit is capped by 3–5 meters of thin beds and lenses of chert interbedded with the carbonates. At Jewel Cave this chert is underlain by lesser zones of bedded chert spanning a total range of about 30 m. References to "the" bedded chert in this paper apply to the major zone at the top of the middle carbonate unit. The upper unit consists of massive or thick-bedded fossiliferous limestone containing local breccias and sparse interbeds of dolomite and chert. The upper contact of the limestone is a highly irregular karst surface of late Mississippian age. Paleo-caves and fissures are abundant in the upper unit, less common in the middle unit, and virtually absent from the lowest. Most of them are filled with the earliest detrital Pennsylvanian sediment. The present caves are concentrated at the boundary between the middle and upper units. Wind Cave occupies mainly the upper two-thirds of the formation. Except for one narrow passage, the cave does not reach more than 15 m into the massive basal dolomite. Jewel Cave occupies only the upper half of the formation, with the exception of a single fissure that extends about 20 m lower. Other known caves in the Black Hills span less than a third of the local Pahasapa thickness.

Geologic Structure

The Black Hills were formed by a block-like domal uplift during the Laramide orogeny about 60–70 million years ago (Fig. 1). They are contemporaneous with the Rocky Mountains and share many characteristics with them. The Black Hills contain a central core of Precambrian igneous and metamorphic rocks, and the eroded edges of the younger sedimentary rocks form concentric ridges and valleys around the central area. The uplift is asymmetrical, dipping steeply to the east and gently to the west. Dips on the eastern flank average 20–30° over large areas, with local dips that are even steeper. On the south, west, and north flanks the average dip is only about 5°.

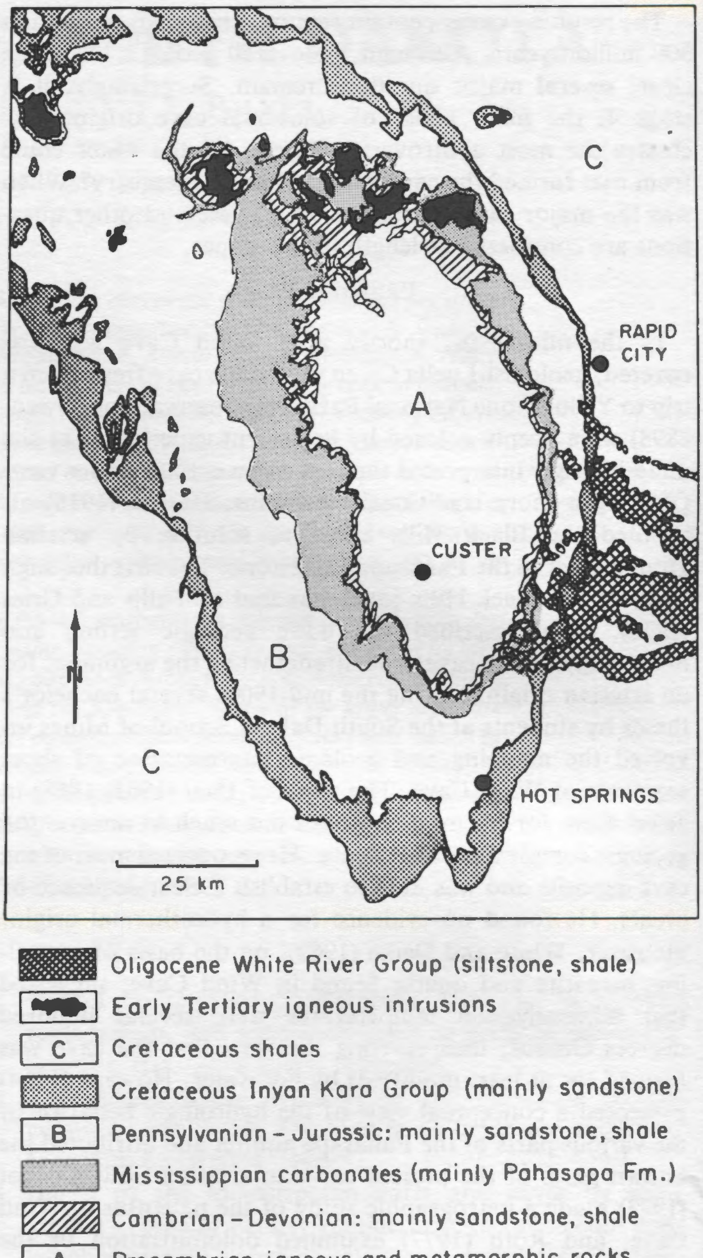

Figure 1. **Geologic map of the Black Hills.**

Joints and faults are very prominent in the massive, competent Pahasapa. Major joints and most faults are perpendicular to the bedding, but the largest faults are mainly low-angle thrust faults. Many of the major faults in the Black Hills date from the Precambrian, as they do not extend upward into the Paleozoic cover (see Fig. 4 and Lisenbee,

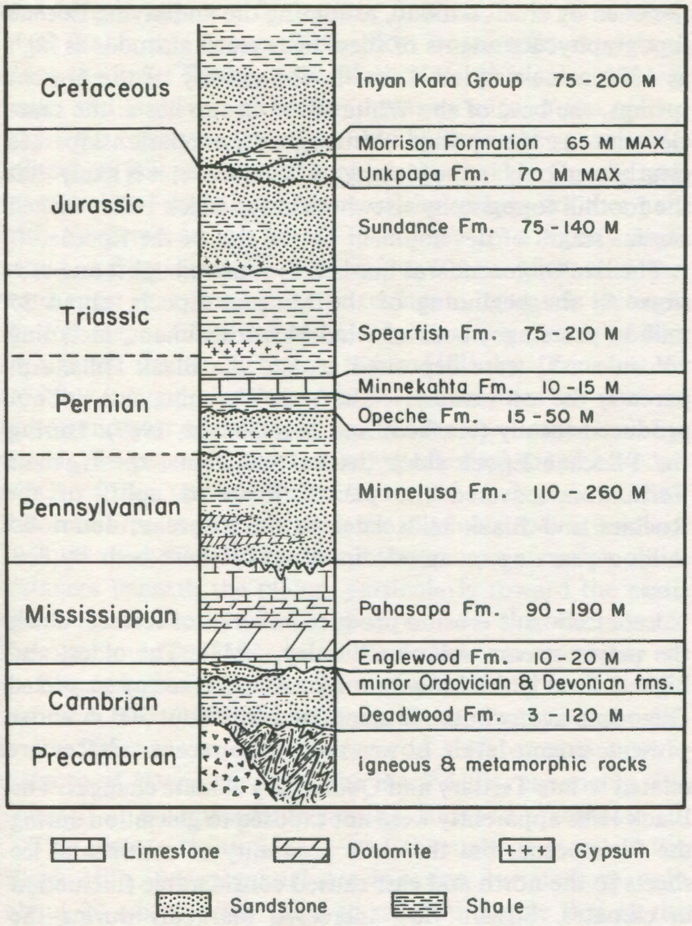

Figure 2. **Paleozoic and Mesozoic strata exposed in the Black Hills.**

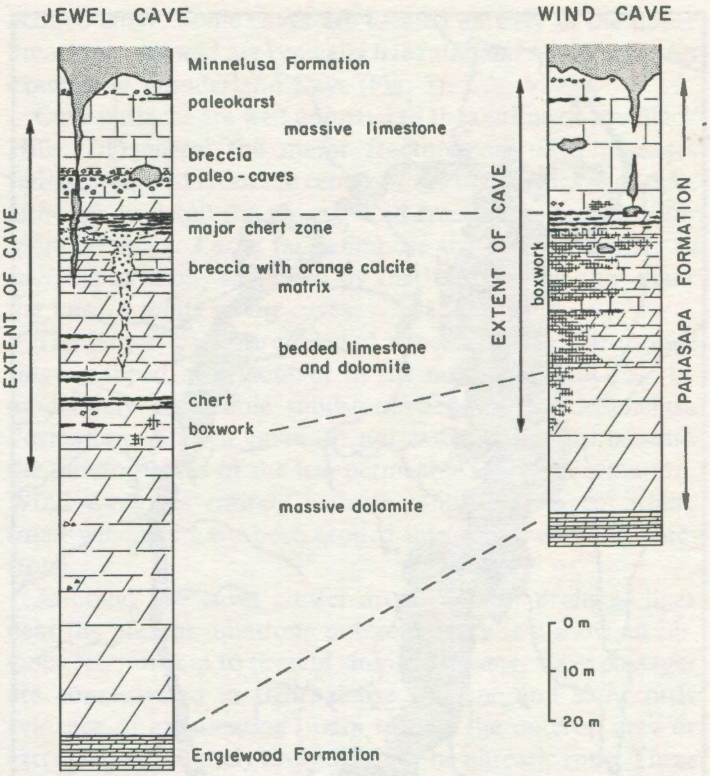

Figure 3. **Stratigraphy of the Pahasapa Formation. The vertical ranges of the caves exclude a few minor passages that extend as much as 20 m below the general limits.**

1985). Those that cross the boundary are probably Precambrian faults rejuvenated during later tectonic events. Major cave passages in Wind, Reed's, and Jewel Caves are parallel to the trends of nearby major faults, even though the faults near Wind Cave are confined to the Precambrian (Fig. 4). Evidently many of the fracture trends in the Pahasapa are inherited from Precambrian structures, or at least influenced by them.

Folding along the southeastern flank of the Black Hills has brought the Pahasapa Limestone relatively close to the surface 5-25 km downdip from its nearest outcrop. Anticlines with axes trending north-northeast are the favored sites for springs draining the limestone, where the limestone is overlain by only about 200 m of detrital sedimentary rocks. Although these rocks are considerably less permeable than the Pahasapa Limestone, local fractures and brecciated zones in the detrial rocks provide favorable routes for upward flow from the limestone.

Geomorphic Setting

Erosion of the Black Hills uplift, mainly during the Paleocene and Eocene Epochs, exposed the Precambrian core of igneous and metamorphic rocks in the area of highest peaks and caused the eroded edges of the sedimentary rocks to retreat radially outward. Differential resistance of the sedimentary rocks has produced valleys and ridges around the outer edges of the uplift. Surface drainage is centripetal from the center of the hills, with the east/west drainage divide located approximately along the western outcrop belt of the Pahasapa, where the limestone forms a broad, gently sloping plateau.

After the Eocene, the climate changed gradually from humid and sub-tropical to semi-arid and temperate (Harksen and MacDonald, 1969; Retallack, 1986). It is important to note that at the end of the Eocene the landscape in the cave areas was rather similar to what we see today. The Oligocene White River Group, consisting of river and wind deposits, covered the truncated edges of all the Paleozoic and Mesozoic strata in many places, preserving the Eocene landscape beneath it (Fig. 1). Fossils and buried soils indicate a warm but increasingly dry climate throughout the Oligocene. The White River beds are being

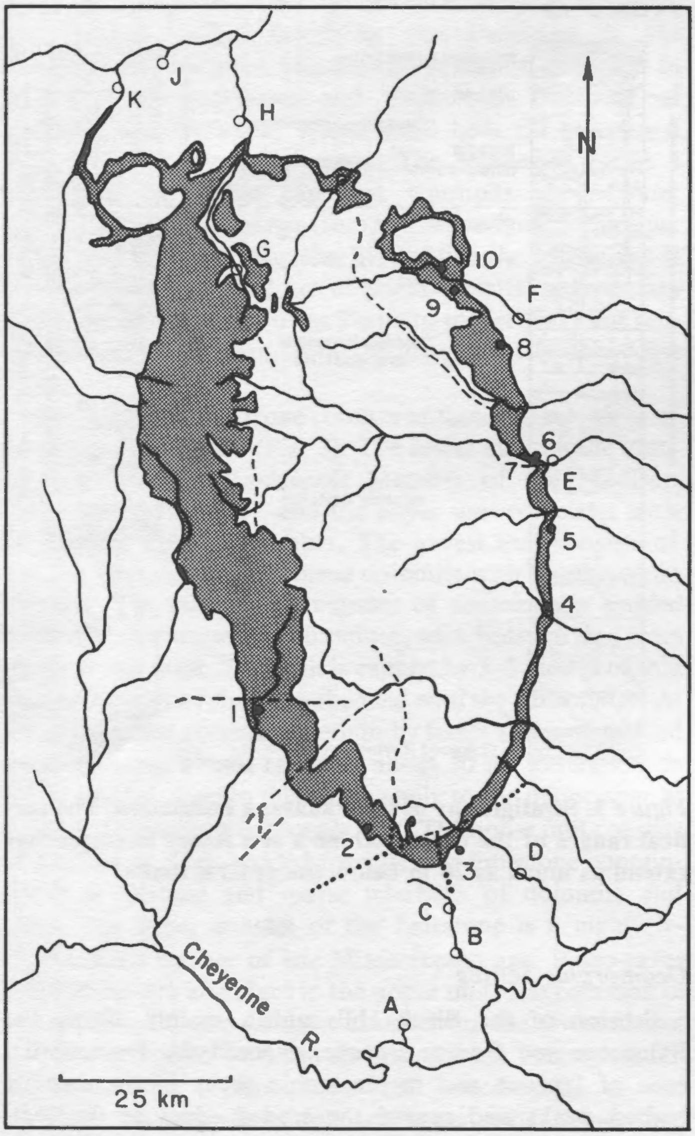

Figure 4. **Location of major caves, springs, and surface streams in the Black Hills. The outcrop area of the Pahasapa Formation is shaded. Major faults are shown as dashed lines where exposed at the surface and as dotted lines where covered by Paleozoic rocks (after Lisenbee, 1985). 1= Jewel Cave, 2 = Reed's Cave, 3 = Wind Cave, 4 = Rushmore Cave, 5 = Sitting Bull Crystal Cavern, 6 = Diamond Crystal Cave (Nameless Cave), 7 = Black Hills Caverns (Wildcat Cave), 8 = Stagebarn Crystal Cave, 9 = Wonderland Cave, 10 = Bethlehem Cave (Crystal Cave). A = Cascade Spring, B = Fall River springs, C = Hot Brook springs, D = Buffalo Gap Spring, E = Cleghorn Spring, F = Elk Creek Spring, G, H = Spearfish Creek springs, J = Crow Creek springs, K = Sand Creek springs. All listed springs are fed by water from the Pahasapa Formation (Rahn and Gries, 1973).**

removed by erosion today, exhuming the underlying Eocene topography. Remnants of these beds reach altitudes as high as 1800 m near Wind Cave. In the vicinity of the present springs, the base of the White River Group lies at the same elevation as the springs. Although these relationships are clearly seen only in the vicinity of Wind Cave, it is likely that the foothill topography elsewhere in the Black Hills reached similar stages of development by the end of the Eocene.

The late Oligocene was marked by renewed uplift and erosion. At the beginning of the Miocene Epoch, about 25 million years ago, beds of wind-blown sediment, including volcanic ash, were deposited around the Black Hills. Apparently the ash was carried in from Wyoming and was not produced locally (Harksen and MacDonald, 1969). During the Pliocene Epoch thick detrital sediments, the Ogallala Formation, covered the plains. Renewed uplift of the Rockies and Black Hills later in the Pliocene, about 4.5 million years ago, caused dissection of these beds by erosion.

Late Cenozoic erosion produced a series of terraces along the major stream valleys (Plumley, 1948). The oldest and highest level is the Mountain Meadow surface of suspected Oligocene age, whose remnants now lie about 300 m above present stream level. Lower and more recent terraces are related to late Tertiary and Quaternary climate changes. The Black Hills apparently were not exposed to glaciation during the Pleistocene, but the close proximity of continental ice sheets to the north and east caused considerable fluctuation in climates. Stream flow increased markedly during the glacial maxima, but during interglacial times the climate was semi-arid.

Interpretation of terrace levels has been scant in recent decades in the prevailing attitude that erosion surfaces are a figment of overworked imaginations. Yet terraces continue to prove useful in determining hydrologic conditions during the late Cenozoic. Dating of terrace deposits along the Fall River in Hot Springs shows that the river may have eroded downward 60 m in the past 26,000 years (Laury, 1980). This rapid erosion rate may help account for the concentration of spring outlets in the area, but many remnants of paleosprings at this location pre-date the dissection. The Mammoth Site in Hot Springs is one of these. The relationship of terraces to the Black Hills caves is still uncertain, because water-table fluctuations in the Pahasapa are controlled largely by the rate of groundwater recharge and by changes in the permeability of the overlying rocks through which the water must rise to reach the springs.

Hydrochemistry of the Pahasapa Aquifer

Today the Pahasapa forms one of the most extensive artesian aquifers in North America. Most of its groundwater recharge apparently comes from the Black Hills, its only outcrop area in the region, although much recharge also comes from adjacent formations (Darton, 1918; Swenson, 1968; Rahn and Gries, 1973). Most water exits at springs along the lower flanks of the Black Hills, particularly along faults and anticlines at the southeastern edge of the uplift, where the limestone is capped only by the rather permeable Minnelusa Formation and, in places, by the thin Opeche and Minnekahta Formations. Many of the springs are thermal, with temperatures of 15–30° C, owing to a locally high geothermal gradient, as well as the 500-m depth that the water must penetrate below the surface in its travels through the Pahasapa. Chemical and isotopic measurements of hot springs of the southeastern Black Hills by Davis et al. (1988b) suggest that their high temperature is due to radiogenic heat or enhanced local heat flow, rather than by chemically generated heat or a hypothetical magma body. Much of the upward flow, both past and present, has followed breccia pipes through the insoluble rocks (Bowles, 1968; Gott et al., 1974). Some water continues for great distances beneath the plains, particularly toward the east, where it eventually flows upward into the Cretaceous Dakota Sandstone, because of diminishing permeability of the Mississippian rocks (Swenson, 1968). Busby et al. (1983) and Back et al. (1983) show from well data that groundwater in the Pahasapa aquifer increases in sulfate content with distance of flow away from the Black Hills, enhancing the solution of dolomite while causing calcite to precipitate.

The climate today is semi-arid throughout most of the Black Hills. Mean-annual precipitation varies from only about 30 cm at Wind Cave to roughly double that in the highest elevations. Rainfall is often flashy, however, and violent floods are common.

CHARACTER OF THE CAVES

Nearly all caves in the Black Hills share the same pattern and geologic history, although only Wind and Jewel Caves clearly exhibit every feature described in this paper. The location and pattern of major caves are shown in Figures 4 and 5. Detailed descriptions are given by Conn and Conn (1981) and by Palmer (1981, 1984, 1988).

With few exceptions, the caves are complex multi-level network mazes arranged in several stratigraphically controlled "levels" that coincide more or less with favorable strata in the Pahasapa. The lowest passages are fracture-controlled fissures in the massive dolomite, the middle passages are mainly wide tubes in the prominently bedded limestones and dolomites, and the uppermost are wide vaulted rooms in the massive limestone. Passages are highly varied in cross section, with large changes in cross-sectional area over short distances. The lower levels are highly interconnective, but junctions are abrupt and ungraded, with sharp jogs in ceiling and floor level (Fig. 6). Upper levels consist of isolated rooms or irregular galleries usually connected to the lower levels by narrow vertical holes in the bedded chert. Some caves are located entirely in the upper limestone unit and are typically irregular and non-linear. An example is Wonderland Cave (Fig. 5).

Cave patterns are well adjusted to the outline of the Black Hills. In general the major fracture-controlled passages radiate outward from the center of the uplift (Figs. 4 and 5), although in detail the passage and fracture patterns are far more complex. Large faults impose the greatest barrier to cave exploration, as shown by the linear trends that mark the known limits of the caves.

The largest caves are located beneath rather flat-topped ridges capped at present or in the not-too-distant past by moderately permeable sandstone beds of the Minnelusa Formation. Known caves do not extend downdip beneath the outcrop areas of the less permeable Opeche Formation. Wind Cave lies entirely beneath sandstone, except where small windows have been eroded into the underlying limestone.

Although the caves cluster around major drainage lines near the present limestone outcrops, very few show an obvious relationship to present sinking streams. Cave passages are concentrated in tight, dense patterns and show little evidence of culminating updip toward the outcrop area or extending indefinitely downdip into the phreatic zone. These limits are surely biased by explorational barriers. In most caves elsewhere in the world it is possible to extrapolate easily beyond the truncated ends of passages to former groundwater inputs and outlets. However, such relationships in the Black Hills caves are cryptic at best.

The present caves intersect many of the late Mississippian paleokarst features, whose fill of detrital sediment is easily recognized by its bright red or yellow color. This sediment spills into the more recent passages, exhuming some of the paleo-caves (Fig. 7). Most of the vaulted upper-level rooms have narrow irregular tubes rising from their ceilings (Fig. 8). It is possible to follow many of them upward to the base of the Minnelusa at the paleokarst surface (Fig. 9). Many of these rooms and rising tubes are partly filled by Pennsylvanian sediment.

Vadose drips are common in places, but running water is very scant in the caves. Almost every one of the few known cave streams appears to occupy preexisting passages rather than form caves by itself. The cave-forming water was slow moving, as suggested by the lack of continuity in cross-sectional area and by the almost total lack of flow markings. In a few passages the solutional walls appear to contain local scallops with wave lengths that range between 20 cm and a meter, but their direction and interpretation are unclear. Many passages in the middle and lower levels of Wind Cave contain ripple marks in the indurated surfaces of carbonate weathering debris, which generally indicate flow in the down-dip direction, but they merely represent late-stage fluctuations in water level after the caves formed. Gravel and sand have been carried into the caves during floods, but

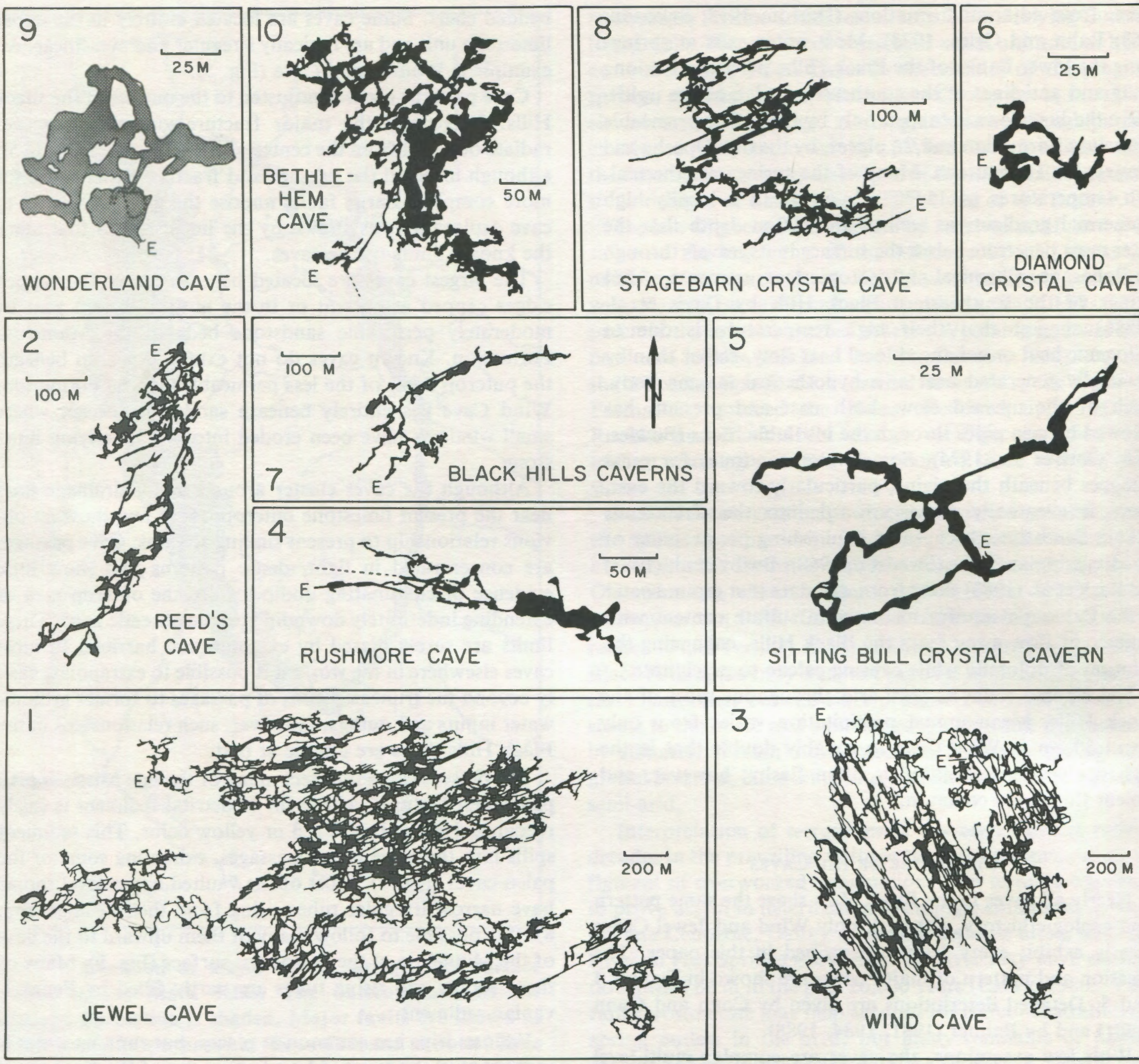

***Figure 5.* Maps of major Black Hills caves. Locations are shown by identifying number on Figure 4. E = natural entrance. True north is upward on all maps. For detailed maps and sources, see Schilberg and Springhetti (1988).**

the only extensive stratified sediments are in paleo-caves older than the present caves.

Speleothems are abundant but strongly dominated by small evaporative forms such as aragonite needles, popcorn, and moonmilk. Most of the caves contain thick wall crusts of calcite, or weathered remnants thereof, which post-date the solutional cave development. Cave walls and ceilings that are free of calcite crust are deeply weathered by subaerial processes, and the resulting carbonate sand covers many of the floors. Boxwork is abundant in several caves, particularly in passages that extend through the middle unit of the Pahasapa and in which subaerial weathering of the bedrock has been intense.

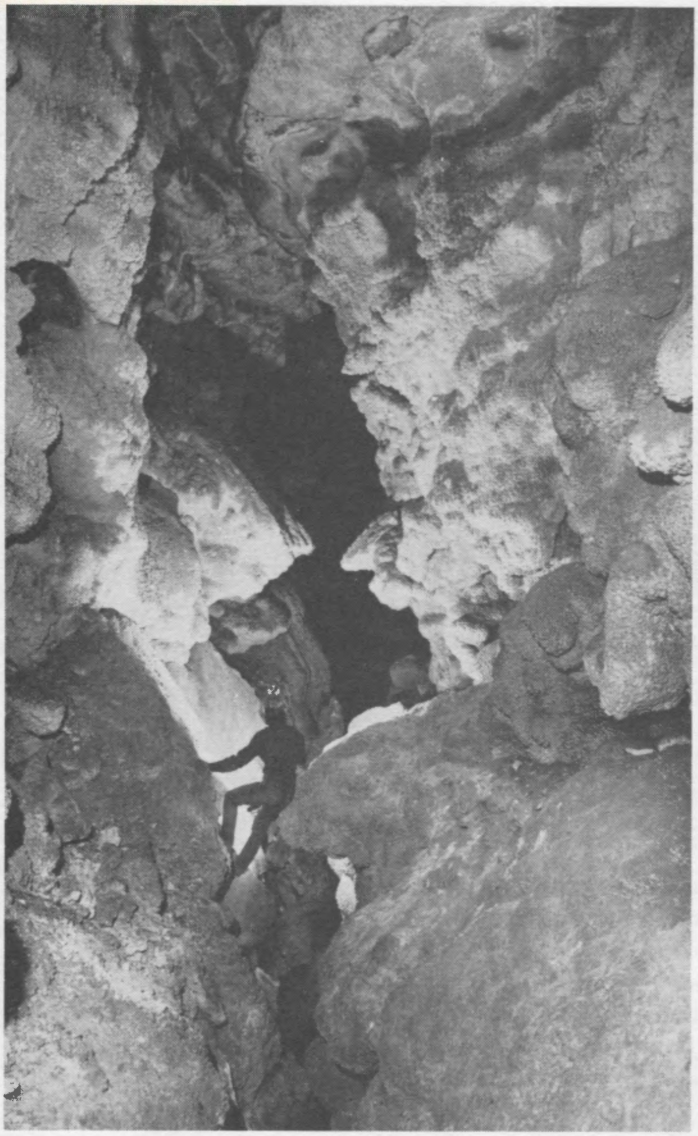

Figure 6. **A typical fissure passage in Jewel Cave.**

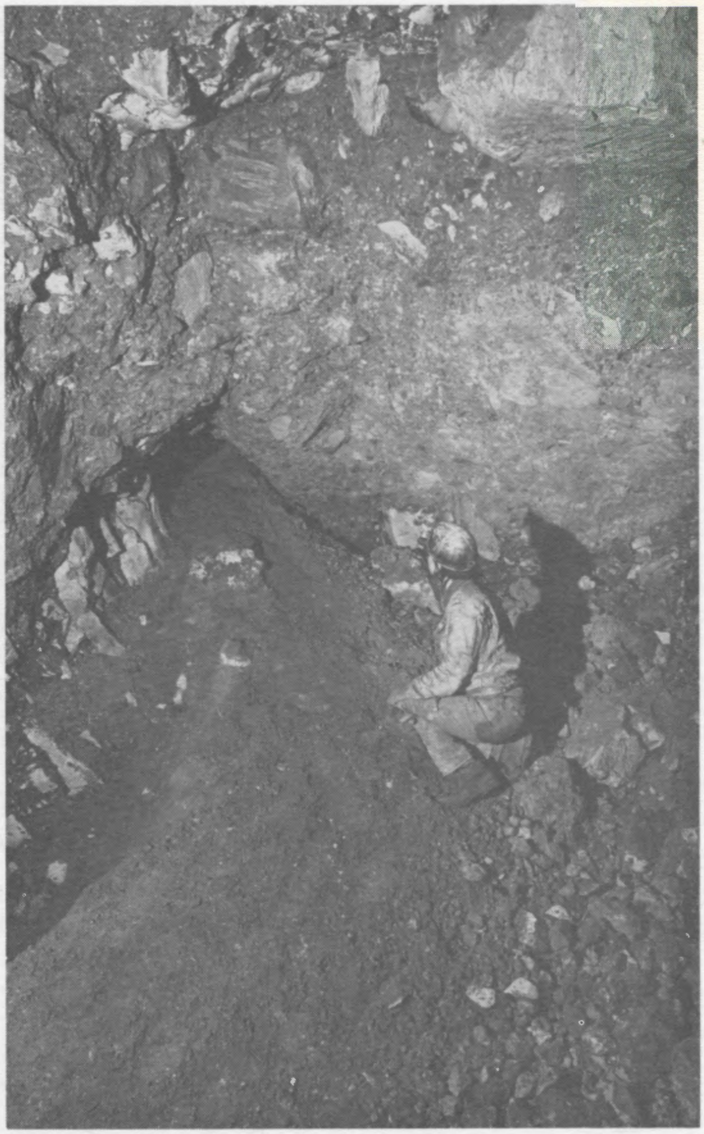

Figure 7. **Pennsylvanian paleofill subsiding from a Mississippian cave intersected by the later stages of Wind Cave.**

GEOLOGIC HISTORY OF THE CAVES

The following interpretation is based on geologic mapping and petrographic analysis. All major events are represented to varying degrees in nearly all the Black Hills caves. Although many of the events described here pre-date the cave origin, they all had a marked effect on the caves. Controversial topics regarding cave origin during the Tertiary Period are discussed at greater length in a later section.

Stage 1: Early Diagenesis and Brecciation of the Limestone

Early chemical and physical changes (diagenesis) in the Pahasapa Limestone soon after it was deposited have greatly affected the character of caves in the Black Hills. This is the first detailed study of early diagenesis in the Pahasapa Limestone. The previous lack of information on this subject is one reason for the rather slow progress in interpreting the Black Hills caves. Our main conclusion is that the early history of the Pahasapa was dominated by processes related to extensive sulfates (gypsum and anhydrite) intermixed with the carbonate bedrock. The sulfates have been almost entirely removed from the exposed areas of Pahasapa, so their interpretation must be determined from indirect evidence.

Role of Sulfates

Former evaporites have had a remarkable effect on wall

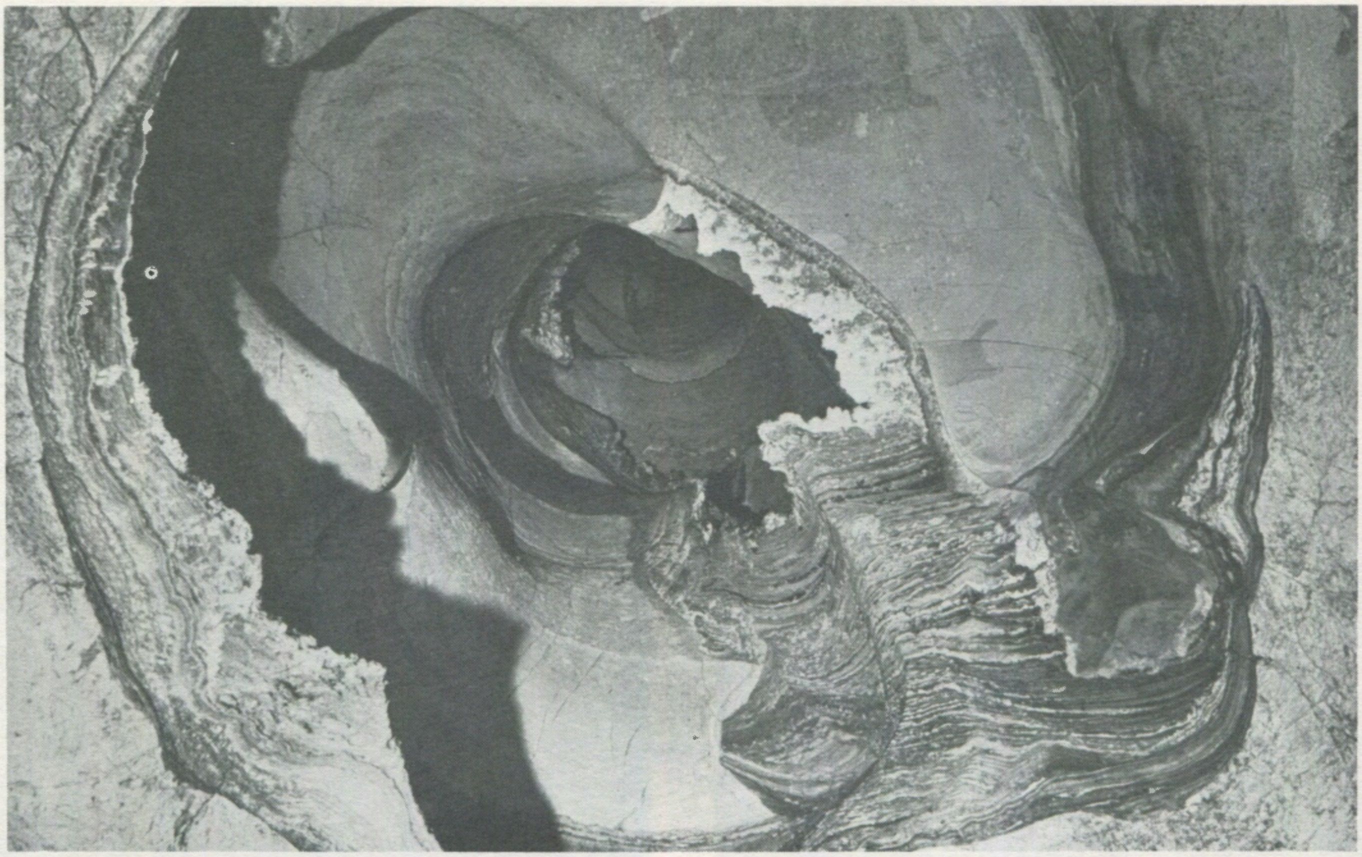

Figure 8. **View straight upward into a solutional dome in the ceiling of an upper-level room in Jewel Cave. The dome diameter is about 2 m. The walls are partly lined with remnant layers of carbonate sediment from subaerial weathering of bedrock.**

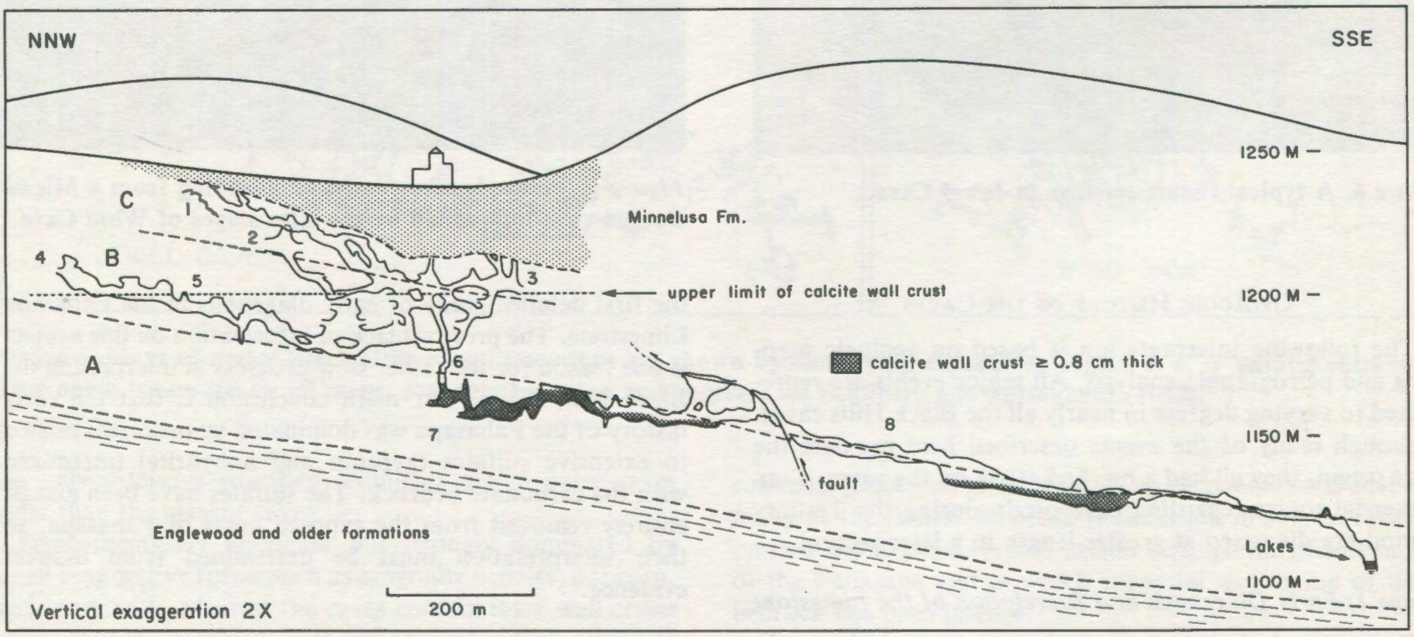

Figure 9. **Geologic profile through part of Wind Cave, showing the relationship of passages to the strata. A = Massive dolomite unit, B = bedded dolomite and limestone unit, C = massive limestone unit. 1 = Club Room, 2 = Fairgrounds, 3 = Garden of Eden, 4 = Pearly Gates, 5 = Candlelight Tour route, 6 = Boxwork Chimney, 7 = Calcite Jungle, 8 = route to the Lakes.**

textures in the caves. Their modification of the carbonate rocks in contact with them is shown by diagnostic mineral associations, fracture patterns, and preservation of relict textures. Many beds are extensively fractured and brecciated, with vein fillings and breccia matrix of iron-oxide-rich calcite. Breccia is most extensive in Jewel Cave, whereas closely spaced fractures with paper-thin calcite veins are more common in Wind Cave. Today, in zones of deep weathering, both the breccia matrix and the veins project into the caves as boxwork fins. In addition, a variety of mineralogical changes took place in the bedrock that are represented in the present caves by resistant rinds of calcite and granular quartz discordant to the cave walls.

This view is compatible with that of Sando (1974), who interpreted breccias in the Madison Group of Montana and Wyoming as collapse features in former sulfate zones. Most other workers consider such breccias to result from solution of carbonate bedrock (e.g., Sangster, 1988). In mining areas they are usually regarded as hydrothermal (e.g., Ohle, 1985) because their matrix contains sulfide ores.

Much of the Pahasapa was deposited in a shallow-water environment exposed periodically to high evaporation. The dolomite in the lower and middle units is a product of this environment. A large amount of gypsum and anhydrite also accumulated as beds, nodules, and irregular bodies within the Pahasapa. Some was also disseminated as pore filling within the dolomite. These sulfates have been removed almost entirely from the cave areas by solution and replacement, but they remain in correlative beds deep beneath the surface around the Black Hills.

Before the Pahasapa was buried beneath Pennsylvanian sediment, it experienced a change from salt-water to freshwater conditions as the area gradually and intermittently emerged above sea level. This change undermined the stability of the gypsum and anhydrite.

Under these conditions, sulfates disrupt the surrounding rocks considerably (Fig. 10). Hydration of anhydrite to gypsum causes a volume increase as much as 30%, which can distort and fracture adjacent rocks. The reverse reaction may cause subsidence and fracturing. Depending on the character of the ambient fluid, gypsum can replace limestone, or vice versa, again usually with a change in volume. Stoichiometric replacement of gypsum by calcite results in a two-fold volume reduction, which creates vuggy porosity. Crystalization of gypsum in the limestone bedrock results in mineral-filled fractures that intersect to form angular blocks or are anastomotic, curved, or scimitar-shaped. Differential pressures cause plastic flow of sulfates, which fractures and distorts the surrounding rock. Recrystallized sulfates can fill fractures and wedge their walls apart.

Selective solution of dolomite from carbonate bedrock is

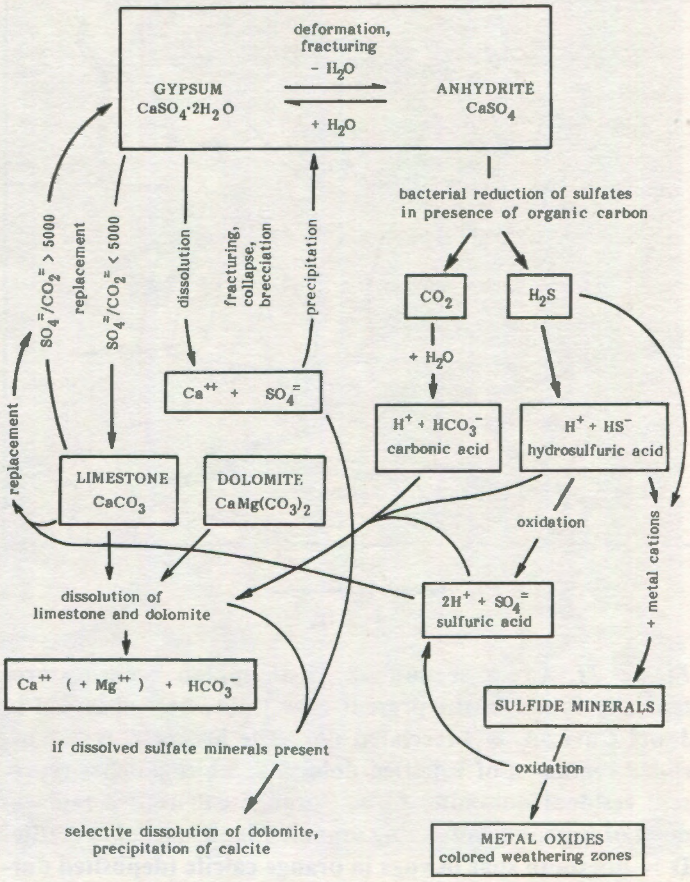

Figure 10. **Effects of sulfates on carbonate rocks.**

common in the presence of sulfates, as is the direct replacement of dolomite by calcite (dedolomitization). The solubility of both limestone and dolomite is reduced by the presence of dissolved gypsum or anhydrite by the common-ion effect, in which calcium ions are contributed by each rock type. However, the solubility of dolomite decreases much less than calcite because the magnesium ions from the dolomite readily combine with the sulfate ions to produce uncharged ion pairs ($MgSO_4^o$). Under these conditions, calcite tends to precipitate long before dolomite reaches saturation, restricting the calcium concentration and allowing still more dolomite to dissolve. Dedolomitization (direct replacement of dolomite by calcite) is very common in the presence of sulfates.

Reduction of sulfates in the presence of organic material produces carbon dioxide and hydrogen sulfide. When in solution, both of these gases form weak acids (carbonic and hydrosulfuric) that can dissolve carbonate rocks. Oxidation of H_2S can produce sulfuric acid, a very powerful solvent of carbonate rocks that also replaces limestone with gypsum if the sulfate concentration is sufficient. Reaction of H_2S with dissolved metals produces sulfide minerals, occasionally of

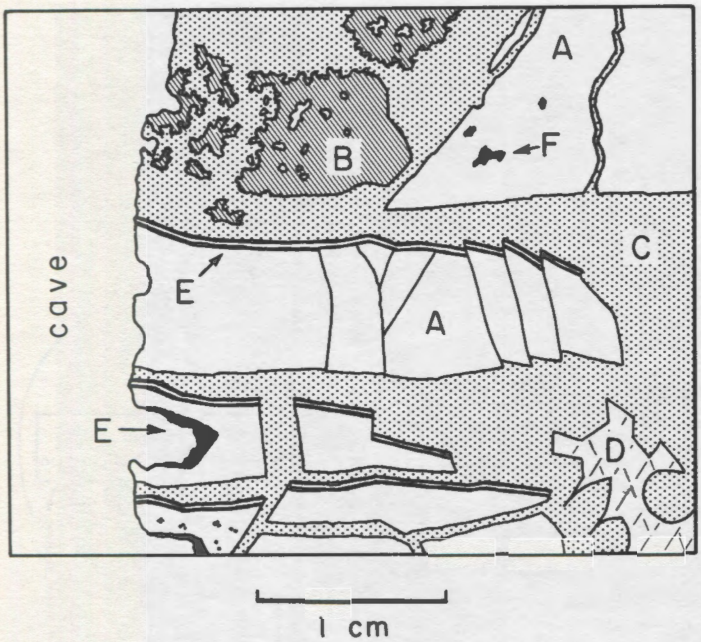

Figure 11. **Cross section of Mississippian rock textures typically found in the present cave walls, best observed in Jewel Cave. A = brecciated dolomite bedrock. B = corroded remnants of calcified dolomite; white patches represent residual dolomite. C = "orange calcite," a replacement of gypsum, containing organic filaments and hematite. D = dogtooth spar in vugs in orange calcite (deposited during post-Pennsylvanian burial). E = scattered bodies and fracture linings of hematite, mainly after pyrite. F = pores due to solution of gypsum inclusions.**

ore grade. Later oxidation of these minerals (particularly iron sulfides) in a more aerated environment produces multi-colored weathering products.

All of these processes occurred within the Pahasapa soon after it was deposited, and their effects dominate the wall character of the caves today. Probably no other caves in the world show such a profound influence by early diagenetic events.

Petrographic Analysis of Mississippian Features

Sulfates were most abundant in the middle Pahasapa unit and in the lower beds of the upper unit, with far smaller amounts in the lower unit. Breccia and fracturing are most prominent in these beds (Figs. 11 and 12). Below the major chert zone, the breccia consists of angular clasts (fragments) that are highly disrupted in places, grading laterally and vertically into zones of less disruptive fracturing where the clasts are still in place and have barely moved relative to each other (called "crackle breccia" in the mining literature). Most clasts have dropped and rotated from their original positions, but some show unmistakable evidence of

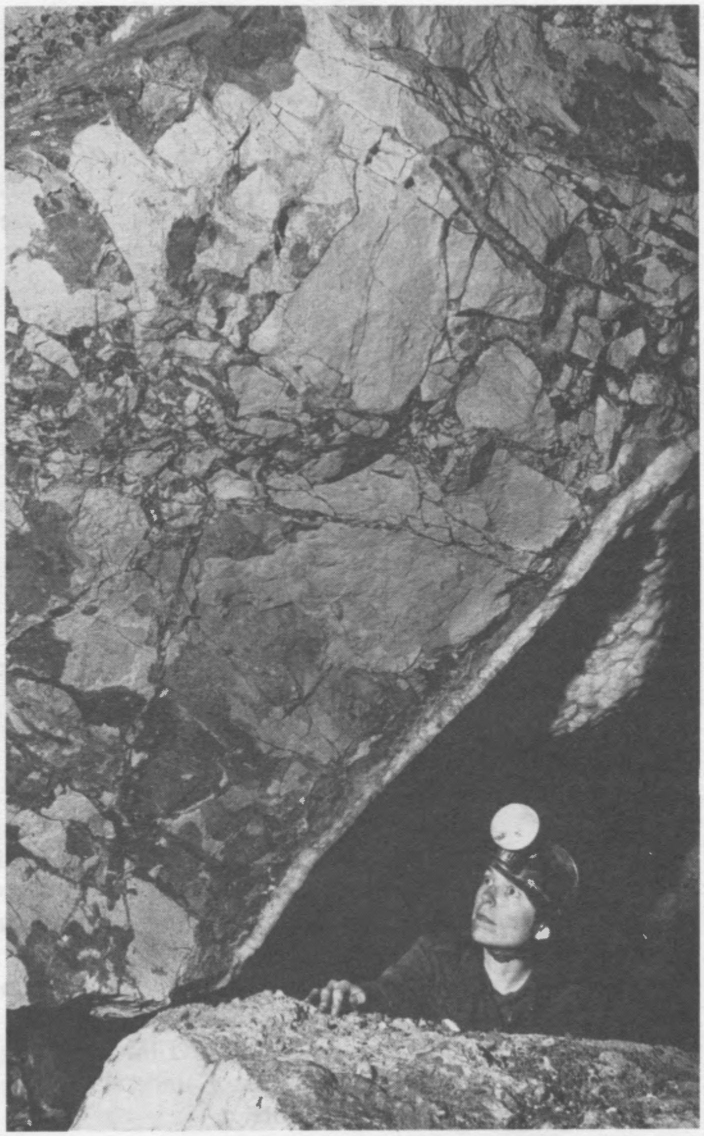

Figure 12. **Sulfate-related breccia in Jewel Cave.**

having been pried upward from undisturbed bedrock. Although previous workers have interpreted these features as cave breakdown related to the Mississippian paleokarst, they apparently resulted from the solution and recrystallization of sulfates considerably earlier than the karst event. Most unusual is a ubiquitous orange-brown calcite (henceforth referred to as the "orange calcite") forming the matrix between breccia clasts and filling the narrower fractures. Its color is caused by abundant particles of hematite (and possibly other iron oxides) averaging 0.05–5 mm in diameter. Most of the iron oxide particles are enclosed in a tangle of microbial filaments, possibly of the bacterium *Leptothrix* (see Hackett, 1987), which average one micron in diameter and up to one millimeter long, and which were encrusted with ferric hydroxide during growth (Fig. 13). Both

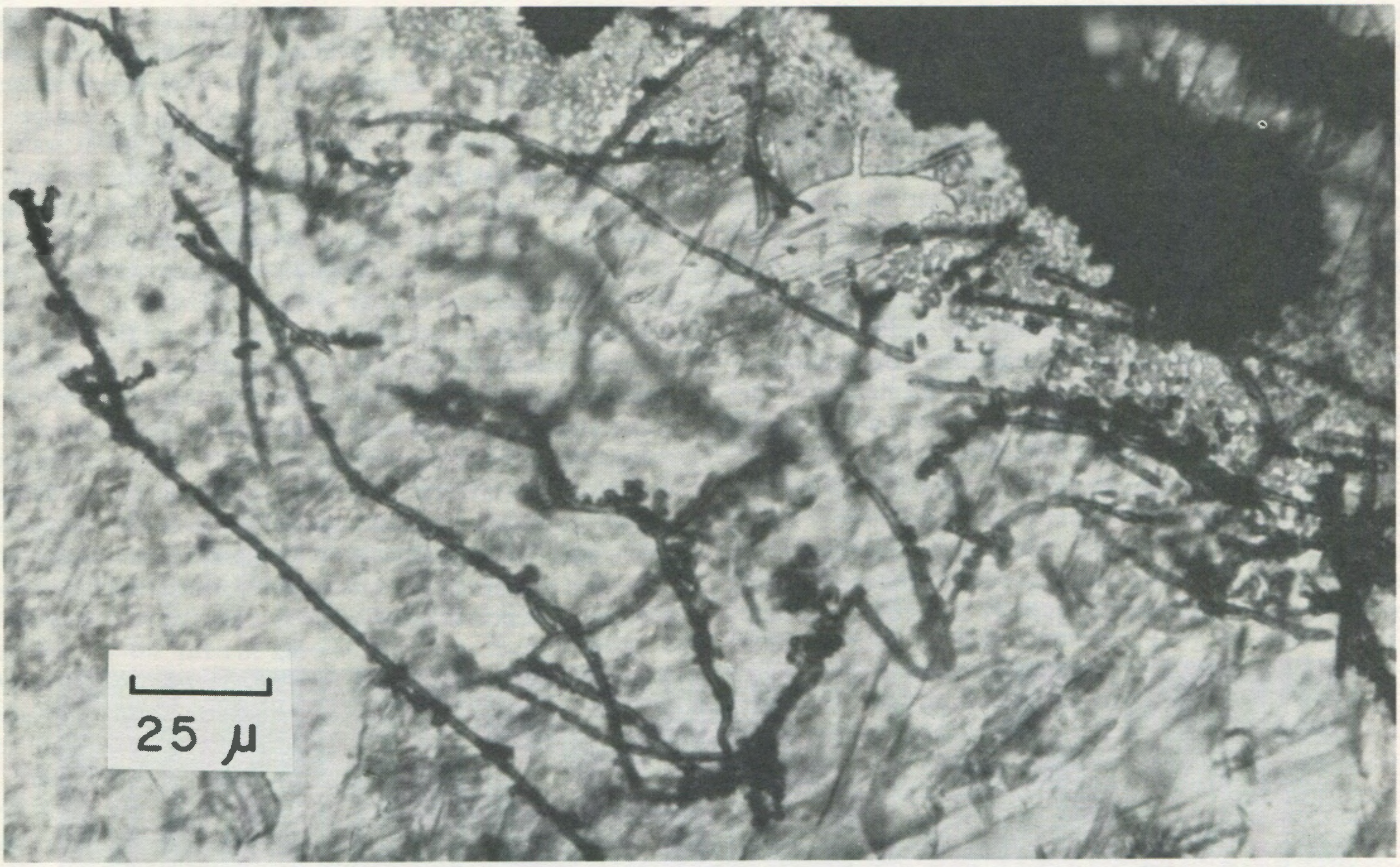

Figure 13. **Photomicrograph of bacterial(?) filaments in the orange calcite matrix between breccia fragments. The large opaque bodies are hematite.**

the oxide particles and the filaments are solidly encased by the calcite. The orange calcite is most apparent in the breccia matrix of freshly exposed bedrock in Jewel Cave and the translucent boxwork veins of Wind Cave.

As sea level began to drop near the end of the Mississippian Period, fresh groundwater gradually displaced the original saline water in which the Pahasapa was deposited. Apparently this change was also accompanied by a change from reducing to oxidizing conditions, because most of the iron oxides are pseudomorphic after pyrite. The orange calcite represents a replacement of original sulfates in the low-sulfate fresh water. This replacement is shown by numerous diagnostic textures within the calcite veins and nearby bedrock: wedging and repeated anastomotic fracturing of carbonate clasts, doubly terminated quartz and calcite crystals, selective solution of dolomite in preference to calcite, alteration of dolomite to calcite, and weathering of dolomite to sand, all of which are associated with sulfate-rich conditions.

In a few places, notably in Wind Cave, the orange calcite forms zones of fine-textured boxwork in wide lens-shaped bodies encased in a thick outer calcite shell. These are the replaced remnants of isolated pods of sulfate within the limestone. In general they are irregularly concordant with the bedding and form zones as much as 5 m thick and hundreds of meters in lateral extent. In places the orange bodies cut across the normal limestone or dolomite strata. Passages in these zones are highly irregular in plan and profile, with crumbly boxwork-lined surfaces.

The thick calcite walls of former sulfate zones are resistant and coarsely granular, with thin laminations parallel to the bedding. They form parts of the walls of many passages in Wind Cave, and although they appear to belong to the solutional cave surface they are actually the outer shells of Mississippian features that have been intersected by the present cave. Smaller bodies of orange calcite form hollow balls that look like hornet nests. Some of the bodies contain thick fins of calcite sub-parallel to the bedrock bedding, spaced about a centimeter apart and lined with orange spar crystals that extend into the intervening voids. This is a form of "zebra rock" derived from interlaminated gypsum and carbonate in algal zones.

In the upper Pahasapa unit at Jewel Cave, 3–10 m above the chert, is an extensive bedded breccia, the lower 5 m of which is red-orange from the presence of iron oxide and contains nodular blobs of white calcite. This zone is overlain disconformably by a more chaotic breccia of light-gray limestone with a matrix of clay and detrital carbonate. The clasts are slightly rounded and have sutured interpenetrating boundaries, indicating that brecciation took place before the limestone was completely indurated. These breccias correlate with evaporite beds still present in the Williston Basin north of the Black Hills (Sando, 1974). All major lofts in Jewel Cave are located in these breccias, and the greatest passage widths are at the contact between the light gray and orange breccia beds. Their domed ceilings extend upward through two or three red-weathering beds of limestone containing rhombic calcitic pseudomorphs after dolomite ("dedolomite"). The brilliant red is a memento of the former iron-rich dolomite. Conversion of dolomite to calcite is very common in sulfate-rich zones (Fig. 10).

Another unusual Mississippian feature inherited by the present caves, particularly Wind Cave, is porous but resistant rinds of granular friable quartz averaging one centimeter thick, concentrated within and below the major chert zone. They form hollow meringue-like shells overlying very porous and crumbly bedrock or forming a continuous lining around chert nodules. Individual quartz crystals average about 15 microns in diameter. At first they appear to be weathering rinds on the present cave walls, but closer inspection shows that they are discordant to the present walls, forming rounded pillar-shaped or nodular bodies typically about 50 cm in diameter (Fig. 14). Some emerge discordantly from more solid surrounding bedrock or line both sides of narrow fractures, and others are partly free-hanging from the walls. They apparently were zones rich in organics and/or sulfates in which the outer part of the original material was replaced by silica. In places their porous interiors contain fossil burrows. Some of the bodies consist entirely of large burrows replaced by quartz. Below the chert the quartz is disseminated within crumbly altered zones surrounding boxwork veins and does not form a resistant rind.

The age of the breccia, orange calcite matrix and boxwork veins, and friable quartz is unquestionably middle-to-late Mississippian, because they are sharply truncated by the late Mississippian paleokarst. Calcite veins in the paleokarst fill

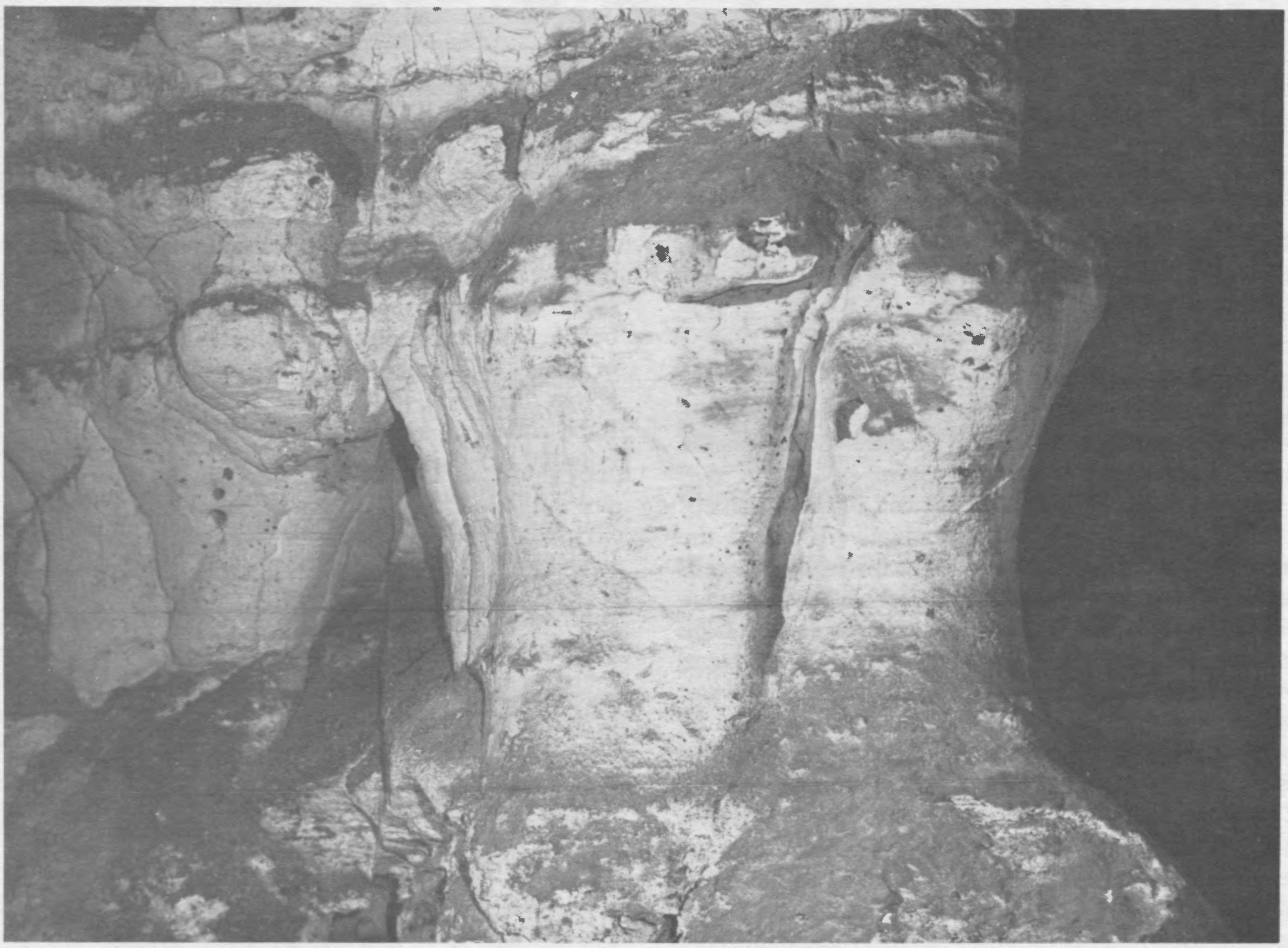

***Figure 14.* Rinds of finely crystalline quartz of Mississippian age, once bordering sulfate bodies, now form partly hollow pillars intersected by the later stages of Wind Cave. Height of pillars is about one meter.**

are white, rather than orange and iron-oxide rich, and contain no organic filaments. The white veins apparently formed during the deep burial of Stage 3.

Stage 2: Mississippian Karst

The top of the Pahasapa Limestone is a karst surface with up to 20 m of local relief, preserved as a paleokarst by a cover of basal Pennsylvanian clay and sand. Former surface karst features include broad sinkholes formed by solution and collapse, as well as narrow fissures that extend several tens of meters into the limestone. Paleo-caves are numerous, many of which were completely filled with Pennsylvanian sediment. Paleo-cave remnants visible in surface exposures or intersected by later caves are mainly linear fissures or wide arched rooms. Some have been partly or completely exhumed where they were intersected by canyons or later caves. Others seem to have been filled only partly, or not at all, by the Pennsylvanian sediment. The spatial density of paleo-cave passages appears to have been approximately the same as that of the upper levels of the present caves but not as great as the present cave density in sub-chert levels. Where paleo-caves extend below the major chert zone, as is common in Jewel Cave, they preferentially follow preexisting breccia zones.

Veins of orange spar do not project as boxwork into the paleo-caves, as they do in the present caves. Instead, they were dissolved concordantly with the surrounding bedrock walls, or even dissolved preferentially. Evidently the chemical environment of the paleokarst was quite different from that in which the present caves formed.

The recently deposited Pahasapa Limestone was just emerging above sea level when the late Mississippian karst formed. Saline groundwater was slowly being displaced by fresh water, and the caves may have formed in the mixing zone between the two. Their origin and geologic setting was rather similar to those in the Bahamas, Bermuda, and the Yucatan today (see Mylroie, 1988; Palmer et al., 1977; Back et al., 1984).

It is possible that many or all of the upper-level rooms above the chert (including the "lofts" of Jewel Cave) were once paleo-caves. Their pattern of isolated vaulted rooms is markedly different from the network of fissures and tubes beneath the major chert, and their trends are not the same. Instead, their morphology is similar to that of the Mississippian caves still filled with Pennsylvanian sediment seen in many exposures of the Pahasapa and correlative strata, such as the Madison Group in the walls of the Bighorn Canyon of Wyoming-Montana. Their elevations appear to have been hydrologically controlled, because the cavernous horizon cuts across the strata. For example, in Wind Cave their floors are located within or at the top of the major chert zone, but in Jewel Cave they average 10 m above the chert. Such discordance would be expected of cave origin at a fresh-water/salt-water contact.

Stage 3: Post-Mississippian Burial and Related Events

The Mississippian caves and karst features were filled with a variety of sediments of late Mississippian and early Pennsylvanian age and by weathered bedrock fragments. The earliest paleo-cave and sinkhole fill in the Wind Cave area is laminated pink and white sand, composed mainly of calcite and quartz and containing hematite-lined calcite pseudomorphs after gypsum. The laminations are interrupted in places by white cumulus calcite bodies that are probably inherited gypsum textures. The laminated beds appear to be local weathering debris that accumulated in arid conditions. In the Jewel Cave area the earliest sediment consists of thin beds of yellow kaolinite (clay) or black manganese dioxide, suggesting a wetter climate than at Wind Cave. Much of the manganese dioxide in the paleofill of Jewel Cave has since been dispersed throughout other passages, reaching a thickness of more than a meter in some lower levels.

All of the early deposits are of Mississippian age, contemporaneous with karst development. They are overlain by thick red clay and sand interspersed with angular fragments of limestone and sandstone and sparse rounded cobbles of quartzite. Most of this material represents basal Minnelusa beds derived from material eroded from nearby land areas to the south. However, the competent sandstone blocks show that fully lithified sandstone must have overlain the limestone before the paleokarst developed. These beds are entirely absent from the Black Hills as the result of Mississippian erosion, but subsurface correlatives occur in adjacent basins.

By the end of the Cretaceous Period up to 2 km of mainly detrital sediment had been deposited on top of the Pahasapa. During this depositional phase of nearly 240 million years the Pahasapa lay buried well beneath the surface at high temperatures and with very little groundwater circulation. Several minor crystalline deposits date from this stage. They occupy vugs and paleokarst features that were not entirely filled with sediment, and they are intersected by the Tertiary cave passages. The earliest deposit was a single layer of white or clear calcite (dogtooth spar), which lined the walls of voids and formed overgrowths on exposed orange calcite of Mississippian age. Similar spar fills vugs in all formations above the Pahasapa up through at least the Permian, and in the cave it forms at least some of the white calcite veins in the paleofill and bedrock.

Crystalline hematite covers the dogtooth spar in a few places, particularly in the vicinity of faults. In Wind Cave some of the underlying calcite has been dissolved away, leaving molds of dogtooth spar in the hematite caps. Hematite also forms thin stalactitic features 1–2 mm in diameter that hang from the roofs of isolated pockets. Some hematite contains abundant organic filaments. Microorganisms are known to thrive thousands of meters beneath the land surface, so their presence does not necessarily imply shallow conditions.

Crystalline quartz was the last pre-cave deposit. It is also concentrated along fault zones. It lines fractures and (rarely) dogtooth spar surfaces in vugs, and it has silicified nearby paleofill deposits. The hematite "stalactites" and spar caps are coated with quartz, with crystals from half a millimeter to one centimeter long. Rare composite quartz/hematite stalactites up to 25 cm long occur in Wind Cave. In Jewel Cave, in the vicinity of faults, tiny quartz crystals line what appear to be fossil burrows in the bedded chert. These are the "scintillites" reported by Deal (1964). Some show branching and internal structure characteristic of roots, but recrystallization has destroyed most of the diagnostic features that would allow exact identification. Crystalline quartz and hematite normally indicate high temperatures of deposition (White and Deike, 1962), which supports an origin at considerable depth beneath the surface. Similar minerals and textures are associated with lead-zinc-gold-silver ores in the northern Black Hills and probably represent the same event. Igneous intrusions near the ore district are of Eocene age (Shapiro and Gries, 1970), but the age of the ores is uncertain.

Stage 4: Uplift of the Black Hills and Origin of the Caves

Uplift during the Laramide orogeny (late Cretaceous—early Tertiary) created the Black Hills as they presently appear, although minor intermittent uplift continued throughout the Tertiary Period. Erosion of the sedimentary rocks overlying the main uplift exposed the underlying Precambrian igneous and metamorphic rocks and caused the eroded edges of the sedimentary rocks to retreat outward in concentric outcrop patterns. The Pahasapa, exposed at the surface for the first time in almost 250 million years, provided the path for much groundwater movement, which resulted in the major phase of development for the caves seen today in the Black Hills. The exact flow routes and chemical nature of the water are still debated, as is the exact timing of speleogenesis. These topics are discussed in a separate section later in this paper.

Tertiary cave passages formed preferentially along porous zones created by the Mississippian breccias, bedrock alteration, and paleokarst. As the walls retreated, resistant Mississippian rinds of orange calcite and quartz around former sulfate bodies stood out in relief. Intersected paleokarst features were partly or completely exhumed of their Pennsylvanian fill, spilling out cones of red-brown sediment and, where phreatic conditions prevailed, producing a dust of sediment on upward-facing surfaces over large areas. The white dogtooth spar of Stage 3 is visible today only where exposed by breakdown or by subsidence of paleo-fill, or as truncated edges of sheets protected by sediment on upward-facing surfaces of paleo-caves. Most of the paleo-voids were small pockets, but some were rooms ranging up to several meters in diameter. Their age is shown by linings of dogtooth spar overlain discordantly by the later calcite wall crust.

Enlargement of the caves may have involved more than simple solution of carbonate bedrock, but it is difficult to distinguish the resulting features from those that formed during diagenesis in Stage 1 and subaerial weathering in Stage 5. Breakdown blocks in the lower levels of Wind Cave, where the walls have been protected from recent weathering by a calcite crust, show an increase in porosity from the bedrock toward the caves. About one meter from the cave walls the bedrock contains scattered needle-shaped pores that are molds of former evaporite crystals. Closer to the cave, centimeter-sized patches of rhombic porosity have been created by the solution of dolomite crystals. The pores are outlined by crystalline quartz, apparently a replacement of gypsum, as shown by the presence of gypsum pseudomorphs. Surrounding the rhombic porosity zones are regions of corroded dolomite crystals. Adjacent to the caves the porosity zones are larger and contain crystals of calcite pseudomorphic after gypsum, with relics of corroded dolomite rhombs. The water that formed the caves was, at least in its later stages, particularly aggressive toward dolomite, suggesting sulfate-rich conditions. Sulfates were still abundant in the bedrock at that time, and it is probable that some of the porosity bordering the caves was produced by solution of interstitial gypsum crystals.

Passage shapes are the best indicators of what was removed during the solutional stage. Both limestone and dolomite seem to have been dissolved at comparable rates. Dolomite in boxwork zones has retreated farther than the limestones, but the intervening calcitic boxwork fins protrude roughly the same distance as adjacent limestone beds. Possibly the additional retreat of dolomite took place in subaerial conditions. Besides the chert beds, the greatest resistance to Tertiary solution was provided by the granular quartz and massive calcite rinds of Mississippian age. Some walls in Wind Cave consist of these rinds over several tens of square meters.

Although caves are common in the dolomite beds of the middle unit of the Pahasapa, no known caves extend into the lower dolomitic unit except in a few places where boxwork veins and other orange calcite bodies are present (e.g., the route to the water-table lakes in Wind Cave, shown in Fig. 9). Therefore, almost all the caves are limited to former sulfate zones below the bedded chert and to the paleokarst-riddled limestone unit above. The location and pattern of Wind Cave have been determined almost exclusively by Mississippian features. This is less true of Jewel Cave, where much solution has occurred along Laramide-age faults. However, the faults also tend to follow Mississippian breccia and fracture trends, which in turn appear to have been inherited from the patterns of major faults of Precambrian age.

Stage 5: Post-Solutional Events

Post-solutional events are almost as complex as those that pre-date the caves. They have produced a variety of phreatic and vadose deposits, as well as subaerial weathering (Fig. 15). Many different features are superimposed because the water table fluctuated considerably during its overall gradual descent since the late Tertiary. The caves lie completely above the water table today, with the exception of Wind Cave, which has several deep water-table lakes in its southeastern (downdip) end.

Phreatic Deposits

Cave development ceased sometime in the Tertiary Period and was replaced by a lengthy episode of calcite wall coating in slow-moving or stagnant water. As described later, the calcite crust appears to have followed a period of subaerial weathering. Such regional fluctuations in the water table were probably caused by late Tertiary and Quaternary climate changes after the main sources of aggressive recharge had shifted to new areas. For example, the three largest caves, Wind, Jewel, and Reed's, are located along major Tertiary valleys whose Quaternary drainage has been pirated elsewhere (see Palmer, 1981). The caves are now located in backwater areas of the aquifer, rather than along lines of aggressive flow. The crust was deposited in nearly all the Black Hills Caves and in many of them it has survived later weathering to remain a prominent feature (Fig. 16). It is most spectacular in Sitting Bull Crystal Cave, where it reaches a thickness of 30–50 cm, with individual crystals projecting as much as 40 cm. Jewel Cave has a wall crust that averages 15 cm throughout all but the upper levels. The crust was once present in the upper levels as well but most has been removed by condensation water. Where the crust remains in the upper levels it is about 10 cm thick. In Wind Cave the crust reaches 15 cm in one isolated place near the southern end, but in general it varies from zero in the upper and middle levels to 2 or 3 cm in the lowest levels (Fig. 9). The crust thickness decreases abruptly downward in places where carbonate sediment once filled the passages and has since been removed by subsidence, erosion, or solution. Many of the lowest passages in Wind Cave have only a one-millimeter crust that accumulated only after the carbonate sediment was removed.

The wall crusts in both Wind and Jewel Caves contain more than 20 distinct layers, although radiometric dating (Bakalowicz et al., 1987) shows that the Wind crust is younger. Remnants of aragonite needles between layers indicate periodic subaerial exposure. Two main crust stages are evident: an inner phase, about one-third the total thickness, which also penetrates several centimeters into the

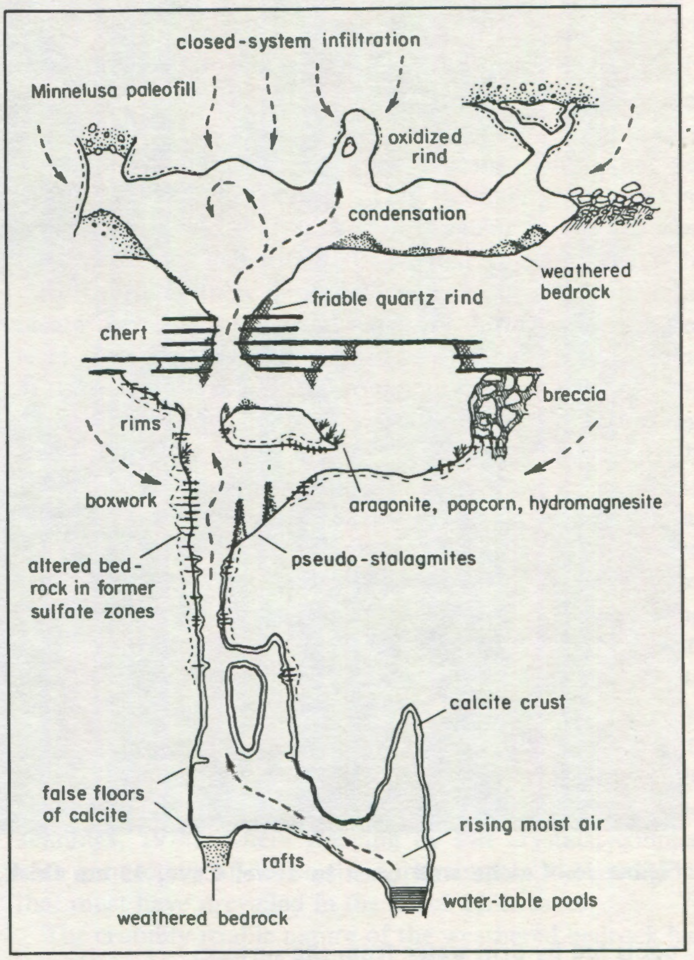

Figure 15. **Stage 5 weathering processes and depositional features.**

weathered bedrock surface as a pore-filling cement, and an outer one consisting of palisade-like crystals. The inner crust appears to have replaced a layer of gypsum in the weathered bedrock surface. Parallel slivers of carbonate wall rock now "float" in this layer. Anastomotic fractures parallel to the rock surface, as well as scimitar-shaped cracks extending into the rock, are now filled with the first crust. These features are common in many dry caves where gypsum, leached from the adjacent bedrock or produced by oxidation of iron sulfide, crystallizes in cave walls.

It seems likely that the calcite crusts were deposited only within the upper few tens of meters of the phreatic zone, where the effect of degassing of CO_2 from the water surface was greatest. Differences in the duration and rate of degassing probably account for lateral variations in crust thickness.

Wall crusts do not necessarily have a direct relationship to cave origin. For example, in Jewel Cave, calcite crusts up to several centimeters thick are presently forming in vadose

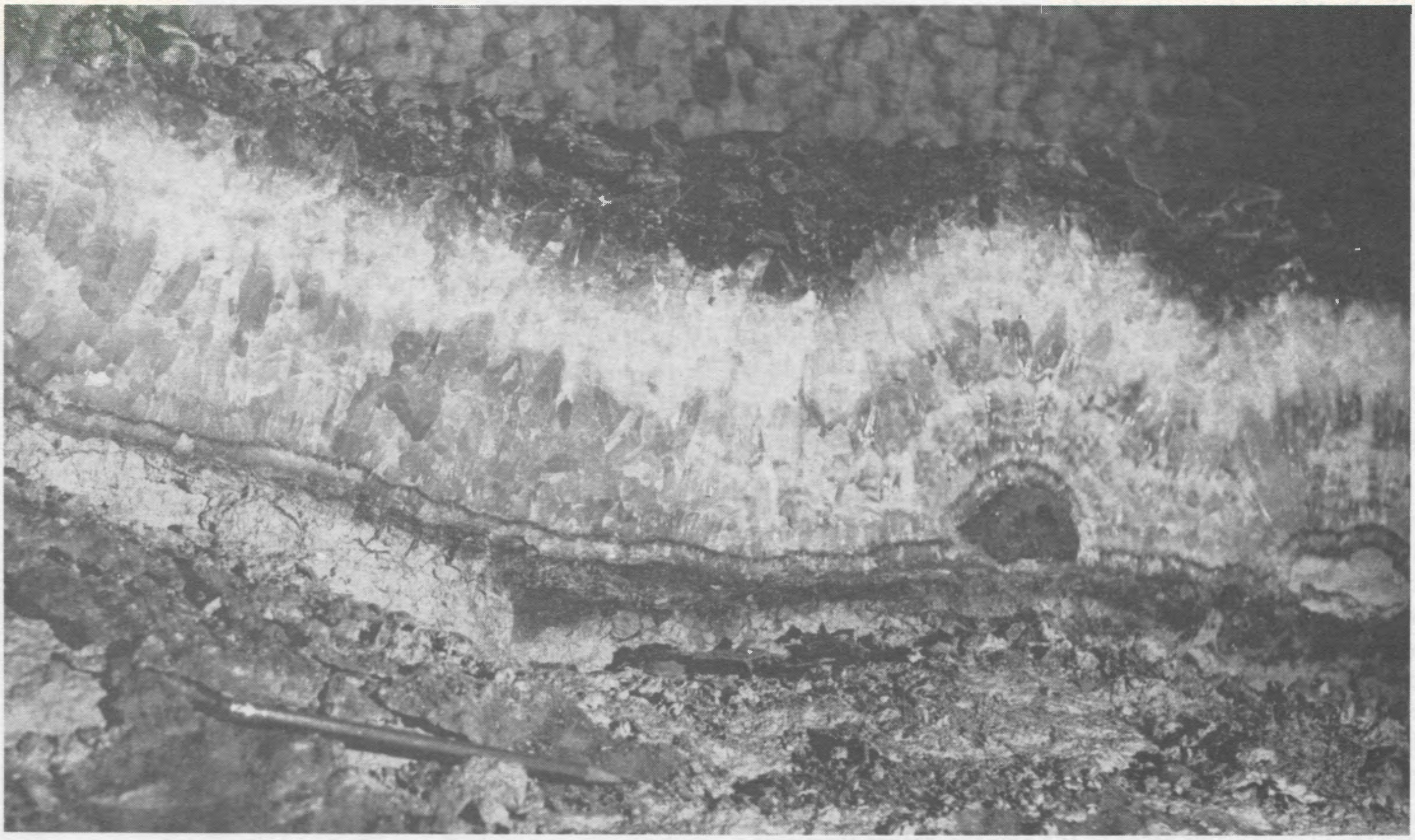

Figure 16. **Calcite wall crust in Jewel Cave, 15 cm thick, broken away naturally from a weathered dolomite surface.**

pools fed by drip water from the surface.

Calcite helictites up to a meter high are common in the lowest level of Wind Cave and are coated with calcite crust identical in thickness and crystal habit to that of nearby walls. Most, but not all, rise from floors composed of thick carbonate sediment likewise coated with calcite crust. The calcite coating suggests that the helictites formed at or near the water table and experienced the same history of water-table fluctuations as the adjacent cave walls. Although most helictites are of vadose origin, these may have formed in pools. Most of their central canals have irregular cross sections up to several millimeters in diameter, too large to conduct capillary water to the requisite height. Inspection of broken pieces shows that the helictites have been greatly recrystallized and that many of their canals mimic crystal shapes. The larger canals are probably secondary, because in some places they are lined with tiny solidified mounds of weathered debris and in others by organic filaments growing away from the openings. Both of these features show that the present canals were originally occupied, at least partly, by solid material that has since been removed. The origin of the helictites is currently being studied by Donald G. Davis and Edward LaRock of Denver.

Calcite rafts form on water surfaces today, especially where there is a dusting of fine weathered debris to form nucleation sites. When they grow too large to stay afloat by surface tension, they settle to the bottom. Rafts occur as much as 55 m above the present water table in Wind Cave, giving evidence of the progressive decline of the water table in the southern end of the cave. Rafts are forming today in certain vadose pools in Jewel Cave, however, so isolated examples do not necessarily indicate former water tables.

Vadose Mineralization

Types of vadose speleothems depend on the amount of water present, the rate of evaporation, the chemistry of the vadose seepage, and the nature of the local bedock. They are described in general terms by Hill and Forti (1986), and with specific reference to the Black Hills caves by Palmer (1981, 1984, 1988). The following brief summary is focused mainly on the relationship of the speleothems to cave origin and hydrology.

The most common vadose speleothems are aragonite needles and calcite popcorn. Less common are dripstone and flowstone, gypsum, and hydromagnesite moonmilk and balloons. Gypsum leached from the local bedrock is deposited as star-shaped crystal clusters, selenite needles, and thin crusts. Porous pseudo-stalagmites, composed of layered calcitic popcorn, aragonite needles, or calcite pseudomorphs of aragonite, form beneath drip points where water condenses on the ceiling, falls to the floor and evaporates slightly. Because the condensation water is usually still aggressive when it drops, many of these stalagmites contain a central drip hole. Some of these features may have grown subaqueously as raft cones beneath drip points, as shown by nearly identical tip elevations in a few locations, but their central holes, if any, are vadose.

Most of the vadose water entering the caves today, and apparently also in the past, is rather diffuse. It tends to evaporate rather than form discrete drips, so dripstone and flowstone are rare. Furthermore, water that has passed through the sandstone caprock is rather isolated from soil CO_2 and approaches saturation with respect to limestone under essentially closed chemical conditions. Under these conditions very little solution is required to drive the water to saturation with calcite, and the CO_2 content of the water drops to a very low level (Palmer, 1987). When the water enters the caves it absorbs carbon dioxide from the high-CO_2 cave air, rather than degassing. It becomes aggressive and corrodes the bedrock where it enters the cave. This phenomenon also applies where the water infiltrates through porous carbonate rock, as shown by chemical analyses of aggressive drip waters in Reed's Cave.

Aragonite, even though it does not contain magnesium, usually forms only in water having a Mg/Ca ratio greater than 1.5. This ratio is exceeded in most drip waters in Jewel Cave (Alexander and Davis, 1987) and undoubtedly also in the diffuse water that deposits the aragonite speleothems. Magnesium is supplied almost exclusively by solution of dolomite bedrock, but even pure dolomite cannot produce substantially more magnesium than calcium. To achieve the high Mg/Ca ratios necessary for aragonite to crystallize in fresh water, evaporation is necessary to drive the solution to supersaturation so that calcite will precipitate. Dolomite does not readily precipitate, so the Mg/Ca ratio is raised. Incongruent solution of dolomite in the presence of sulfates can also boost the Mg/Ca ratio. The necessary evaporation can take place near the land surface where the water infiltrates into limy beds in the Minnelusa Formation, as well as inside the caves where the water emerges into dry passages. Hydromagnesite can also precipitate where evaporative enrichment of magnesium is sufficiently high.

Aragonite occurs mainly above the calcite wall crust, but rarely in the highest levels. Much of it has recrystallized to calcite. In Jewel Cave it is concentrated in bushy growths that lie in the path of rising air currents. In Wind Cave it is concentrated in sub-horizontal zones just above the highest calcite wall crust, suggesting crystallization above static water levels, with secondary control by stratigraphy and patterns of air movement.

Subaerial Weathering

Most of the bedrock exposed in the caves has been deeply weathered to a multi-colored granular powder that crumbles at the slightest touch. The floors of upper levels are heaped with soft piles of carbonate weathering debris that accumulate beneath domes. This sediment consists of equigranular calcite crystals with irregular corroded faces, mixed with tiny hematite particles. Redder quartz-rich sediment overlying these deposits or interbedded with them indicates periodic input from paleo-fill during rises in water level. These episodes of reflooding also redistributed some of the carbonate sediment throughout the caves. Weathering diminishes downward and is nearly absent in the lowest levels. However, many low-level passages have been partly choked with the carbonate sediment from above, as seen in places where the sediment has dissolved or compacted or has subsided to still lower levels. False floors of calcite in Wind Cave indicate former levels of sediment fill. The false floors are continuous with the phreatic calcite wall crust, showing that most of the deep weathering pre-dates the crust. Isolated pseudomorphs of gypsum crystals in the carbonate sediment suggest that the sediments, and therefore also the weathering, had a subaerial origin. Similar piles of carbonate sediment are common in certain other caves in arid climates, as in the Nullarbor Plain of Australia (Lowry and Jennings, 1974), where wedging by salt crystals, coupled with convection of moist air, approximates the conditions that must have prevailed in the Black Hills caves.

The crumbly friable nature of the weathered bedrock has two different origins. Where warm moist air rises from lower in the caves, water condenses on the cooler rock surfaces in the upper levels. This effect is undoubtedly greatest in caves with ponded water in their lower levels. However, this is not necessary, as shown by condensation in the upper levels of Jewel Cave, whose known parts lie entirely above the water table. Condensation water absorbs CO_2 from the cave atmosphere and becomes highly aggressive, readily dissolving both calcite and dolomite and creating porosity that extends to depths of several millimeters. Pre-cave vugs that are intersected by a weathered surface and exposed to this process show little if any solution of their interiors, because water does not condense on their sheltered concave surfaces. Where condensation takes place on calcite wall crusts, the crystals are deeply etched and the faces retreat at rather uniform rates. Flared rims are formed in narrow passages by condensation corrosion in the direction from which moist air is rising, coupled with deposition of aragonite and calcite on the entrance side, which is exposed to drier air from the surface. The condensed water generally drains downward as capillary and gravitational flow through the bedrock and weathered surface rind, depositing needles and popcorn on lower walls wherever evaporation is sufficient. Calcite and aragonite are deposited mainly by evaporation, rather than by CO_2 degassing, because the CO_2 in the condensation water is already in equilibrium with the cave air.

Weathering of cave walls is also caused by diffuse water infiltrating from the surface and drawn by capillary potential toward relatively dry caves. As explained in the previous section, this water is rather isolated from sources of CO_2, so it approaches solutional equilibrium at very low CO_2 partial pressures and dissolved carbonate content, as well as at high pH. As it nears a vadose cave, this water absorbs CO_2 from the cave air, renewing its aggressiveness and dissolving the wall rock. Far from the cave, the bedrock contains tiny elongate pores left by the solutional removal of gypsum. Within a few tens of centimeters of the cave calcite has been selectively removed. Within a few centimeters of the cave, dolomite, if present, has been dissolved incongruently and replaced by calcite. Such dedolomitization is enhanced by the presence of sulfates, which supports other evidence that there was once a great deal of gypsum in the cave walls that has since been removed by solution. The mean concentration of sulfates in drip water in Wind and Jewel Caves is about 10 mg/1 in Wind Cave and about 80 mg/1 in Jewel Cave (Alexander and Davis, 1987), showing a continued but undoubtedly diminished presence of sulfates in the bedrock. Some gypsum continues to crystallize in and on the walls of the caves.

Most infiltrating water produces a deep porous zone that is self-perpetuating: growth of pores allows greater circulation of CO_2 into the bedrock from the cave air, renewing the aggressiveness of the infiltrating water at increasing distances from the cave. This water can precipitate carbonate minerals only by evaporation, as it cannot degas in the cave air. Like the condensation water, it tends to drain through the weathering rind to lower levels, where some of it evaporates and deposits popcorn, aragonite needles, and hydromagnesite.

In the upper levels both kinds of weathering leave a soft fluffy surface. Most susceptible to this kind of weathering is former dolomite that was later calcified either during the Mississippian or in the present cave environment. The ill-fitting interfaces between grains are corroded preferentially, allowing the rock to disaggregate into sand. Limestone exposed to infiltrating water but not to condensation corrosion has developed a white rind up to several centimeters thick, in which organic carbon compounds that darken the unaltered limestone have been oxidized to CO_2.

The irregular tubes and domes rising from the upper levels show much evidence for enlargement by condensation water. Even totally blind pockets contain a large amount of carbonate sediment perched on ledges and on the underlying floors. Most of those that extend upward to the base of the Minnelusa contain easily perceptible air movement. Some are paleo-features that were once filled partly or completely with basal Minnelusa sediment. Unlikely as it seems, much of the air appears to flow through the pores and fractures in the Minnelusa. Most unusual is Plummer's Pit in Wind Cave, a high complex dome in paleo-fill that extends 10 m above the Pahasapa and ends in breakdown rooms in sandstone and dolomite bedrock of the Minnelusa. Air movement is considerable, even at the very top.

In the middle levels the weathered surfaces have a gritty texture and produce a rain of sandy particles when disturbed. They consist of an open network of pores bridged by Mississippian-age quartz crystals. The pores probably represent solution of interstitial gypsum prior to deposition of the basal calcite crust layer.

Where calcite wall crusts are thick, bedrock weathering has been arrested almost completely. The weathered surface was cemented by the basal layers of calcite crust, although in places the cementation was incomplete enough that exposure of bedrock by breakdown allows weathered material to fall out, leaving large pores directly behind the crust. Where chips of calcite wall crust fell away a long time ago, subaerial weathering has eaten into the exposed bedrock like tooth decay beneath a break in the enamel. In Jewel Cave large sheets of calcite wall crust have spalled off a distinctive gray punky dolomite bed. It appears that the dolomite surface was highly weathered before deposition of the crust and that cementation by the first layer of calcite was not sufficient to bond the layer tightly to the walls (Fig. 16).

Certain iron-rich dolomite beds weather to brightly colored reds and yellows. Evaporite textures and pseudomorphs are common in each, indicating that these beds were deposited in arid conditions. Below the main chert, several very finely crystalline dolomite beds contain swirls of contrasting colors (Liesegang banding) caused by iron oxides differentiated from one another by colloidal migration through the bedrock.

Weathering has considerably altered the petrology of the bedrock around the caves. Miller (1988) made chemical analyses of four wall-rock samples from Wind Cave and was surprised to find that none had a high enough magnesium content to be classed as dolomite and that one appeared to be sandstone with a 90% silica content. As shown in previous paragraphs, dedolomitization and silicification of bedrock in the cave zone are responsible.

Deep weathering and calcite wall crusts are also common in caves formed by rising water or gases where there has been an abundance of CO_2 or H_2S. Given enough time, such weathering is possible in any cave where air movement between entrances at different levels is subdued, regulation of temperatures by inflowing surface water is absent, and thermal convection of warm moist air is enhanced by large vertical relief. Regardless of their origin, caves with entrances only at their upper ends, with great vertical relief, and having lower-level pools (past or present) are especially prone to thermal convection. Infiltrating closed-system water may be the dominant source of weathering in some caves. Weathering in caves of the Guadalupe Mountains, New Mexico, is most prominent in dolomites that once contained interspersed sulfates, regardless of the relief of the caves. This relationship applies to the Black Hills caves as well.

Origin of Boxwork

Boxwork consists of resistant fins of calcite projecting into the caves in an intersecting pattern, surrounding box-shaped voids where the intervening bedrock has weathered or dissolved away (Fig. 17). It occurs in most Black Hills caves but is prominent only in Wind Cave, which contains more boxwork than any other known cave in the world. The vein cores consist of the orange calcite of Mississippian age, now largely recrystallized and fattened by the accretion of white calcite crystals and, in places, by calcite wall crust or evaporative speleothems such as aragonite needles and popcorn. The intervening voids are generally 10–20 cm in both depth and width, but with considerable range from much smaller to much larger. Vein thickness ranges from paper thin to several millimeters, although the massive matrix of weathered breccia in Jewel Cave can form boxwork partitions several centimeters thick. Most veins are either perpendicular or parallel to the bedding, but in places they splay outward in fan shapes. The orientations of vertical boxwork fins rarely match those of the major fractures along which the passages are developed.

The traditional explanation for boxwork is that of simple differential solution, in which the coarser crystals of the calcite veins protrude from the finer grained and more readily soluble bedrock. This explanation is inadequate because calcite veins are common in many carbonate rocks whose caves contain no boxwork at all.

Boxwork occurs almost exclusively in porous dolomites, particularly in the middle unit of the Pahasapa. Where exposed by breakdown or by artificial trail widening, the boxwork veins are seen to diminish in width into the bedrock, typically disappearing entirely a few tens of centimeters inward from the cave walls. Observing that the boxwork veins are concentrated around the caves and are rare elsewhere in the Pahasapa, Palmer (1981) suggested that the veins are related to the cave origin, and that vadose water seeping through the porous dolomite begins to lose CO_2 through pores to the cave atmosphere before reaching the cave, depositing less soluble minerals (calcite in this situation) in

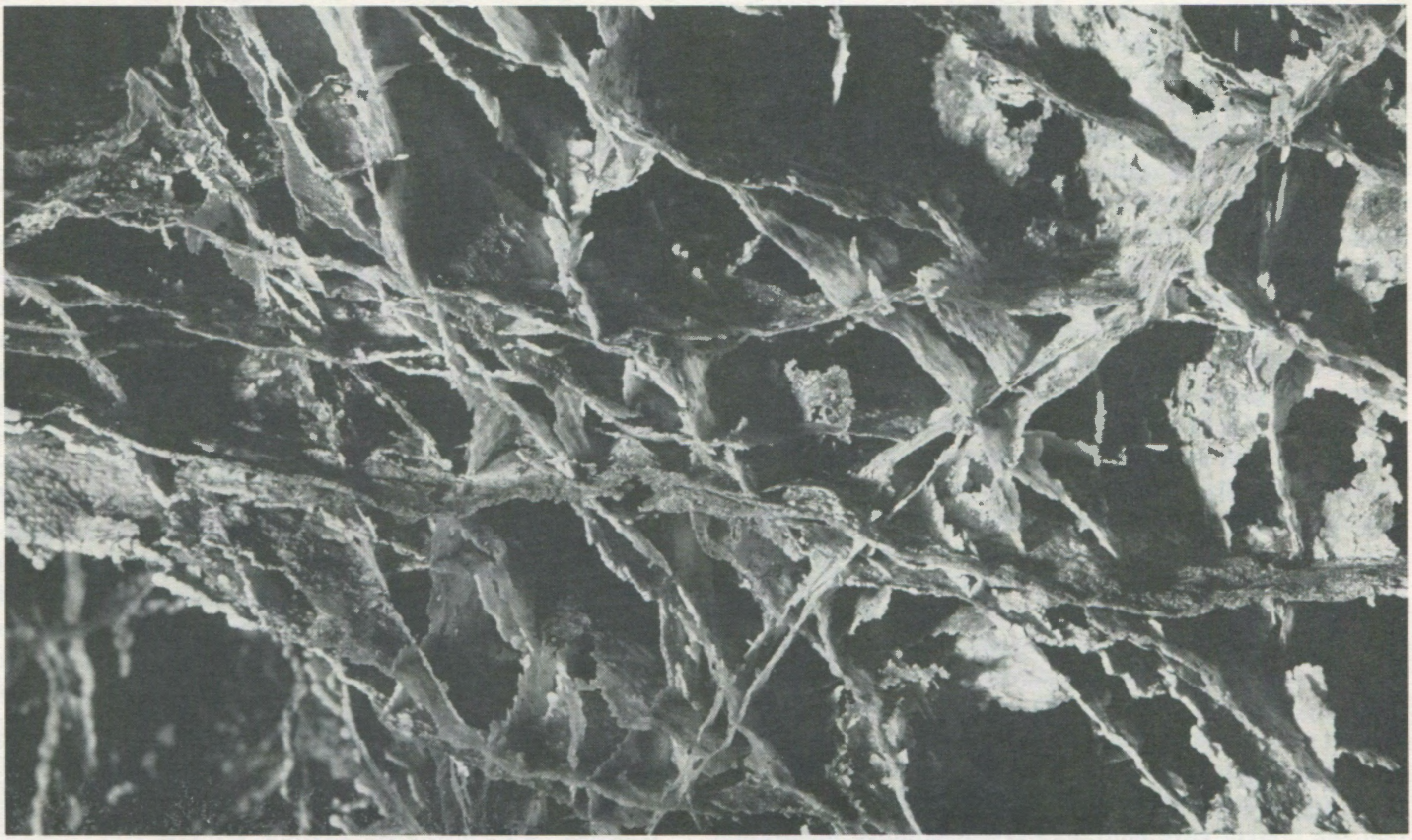

Figure 17. **Boxwork in Wind Cave. Width of photo is about 50 cm.**

fractures and pores. Ford and Bakalowicz (1983) favored a phreatic origin for boxwork, in which incongruent solution of dolomite causes simultaneous deposition of calcite. Millen and Dickey (1987) agreed, citing evidence for incongruent solution deep within the Madison aquifer today (Busby et al., 1983; Back et al., 1983).

Unfortunately for all concerned, the boxwork veins are cut by the Mississippian paleokarst features and therefore pre-date the present caves by hundreds of millions of years. This relationship was overlooked at first because the paleokarst rarely extends downward as far as the boxwork zone. Petrographic evidence shows that the boxwork veins

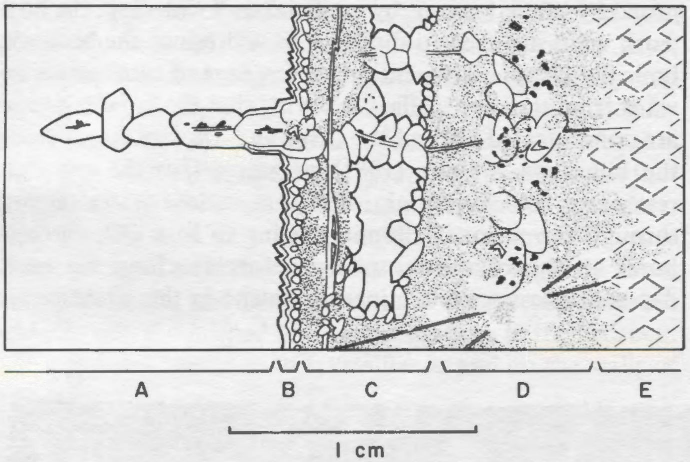

Figure 18. **Cross section through a typical boxwork zone in Wind Cave. A = calcite boxwork vein projecting into cave, enlarged by recrystallization and overgrowth. The original vein is now discontinuous and outlined by remnants of hematite. Most boxwork veins are longer than shown here. B = layered calcite crust on weathered bedrock. This thickness is typical in the middle levels of Wind Cave but can be much greater. C = bedrock altered to friable, finely crystalline quartz, with pores probably formed by solution of primary gypsum and lined by calcite crystals pseudomorphic after gypsum. D = zone in which boxwork veins die out. Calcite crystals partly envelope friable quartz. Former dolomite crystals are outlined by hematite and euhedral quartz casts. E = unaltered dolomite bedrock.**

are of Mississippian age, formed by replacement of sulfates by calcite (Stage 1). Incongruent solution may have played a part, but not in the way envisioned by those supporting simultaneous boxwork-and-cave origin. Boxwork is concentrated around the present caves simply because the caves formed preferentially along zones of preexisting porosity, in particular the brecciated, fractured, and solutional zones exposed to sulfate-related processes. This is further evidence that the present cave patterns were determined mainly by Mississippian events.

Microscopic examination of the boxwork reveals an astonishing complexity (Fig. 18). The original calcite veins filled the earliest Mississippian fractures, many of which were lined even earlier with hematite (see "E" in Figure 11). Dolomite has been almost completely removed from the intervening bedrock, leaving a fragile porous framework of diagenetic quartz interspersed with calcite crystals. The veins and friable quartz occupy former sulfate zones that appear today as deeply weathered or altered bedrock highly colored by oxides. The contact between the red, yellow, or brown granular altered zone and the unaltered bedrock is fairly sharp. In places it is marked by a white centimeter-thick band probably caused by oxidation of carbon in the bedrock during the Mississippian, accompanied by bacterial reduction of sulfates. The removal of dolomite today appears to be by infiltrating closed-system water, as described earlier. Weathering by condensation corrosion alone does not differentiate sufficiently between calcite and dolomite to cause veins to stand out in strong relief. The resistance of the boxwork veins is caused mainly by the crumbly nature of the intervening altered bedrock, rather than by supersaturation of water with respect to calcite. Boxwork is uncommon in the uppermost levels for several reasons. Calcite veins do not contrast in mineralogy with the surrounding limestone, and, more importantly, the severe alteration and porosity of the former Mississippian sulfate zones is absent. Sulfates were originally present, but they were removed by solution before they could be replaced by calcite, as shown by the many bedded breccias with sutured clast contacts and detrital matrix.

Boxwork fins generally extend outward only as far as the limits of Tertiary solution of the caves. In the northern end of Wind Cave, however, many narrow passages contain networks of paper-thin boxwork veins that extend across almost the entire passage width. Weathering reaches depths of about a meter. The sandstone caprock is thinnest in this area, and the walls are moist, supporting the idea that closed-system infiltration is an important ingredient in the weathering that causes boxwork to stand in relief.

CAVE ORIGIN: SOME CONTROVERSIAL TOPICS

Despite so much recent field work, or perhaps because of it, the question of cave origin in the Black Hills is more controversial than ever. Three topics have yet to be resolved: (1) nature of the dissolving power of the cave-forming water; (2) source and flow pattern of the water; and (3) age of the caves. Competing hypotheses on cave origin are shown diagrammatically in Figure 19. Evidence for and against each is given here to help identify the most feasible. Any hypothesis must fit the geologic history of the region and explain the irregular distribution of the caves, their stratigraphic relationships, their network patterns, and the reason why most of them appear to have undergone a nearly identical sequence of events.

Cave-Forming Acid

It is generally assumed that carbonic acid was the solutional agent in forming the Black Hills caves. However, many caves of somewhat similar morphology (e.g., those of the Guadalupe Mountains of New Mexico) were formed by sulfuric acid derived from oxidation of H_2S. Could the Black Hills caves share this origin? Most such caves have thick deposits of gypsum formed either as replacement of limestone in the presence of sulfuric acid (Egemeier, 1981) or in sulfate-rich brines (Queen, 1973), or as primary bedded deposits (Hill, 1987). Sulfur, formed by oxidation of H_2S, and the clay mineral endellite occur in scattered places (Hill, 1987). Both minerals are stable only in low-pH settings. These caves also show evidence of water-table control, where rising gaseous or aqueous H_2S reacts with oxygen. Passages tend to be large with abrupt lateral terminations. Floor fissures are abundant and have been shown in still-active caves to be the inputs for rising water rich in H_2S (Egemeier, 1981). A source of H_2S is present in the basins

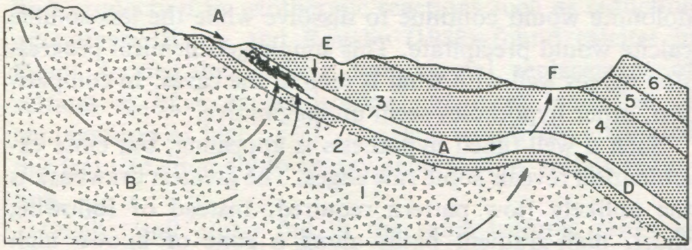

Figure 19. **Conflicting hypotheses for the origin of the Black Hills caves: A = artesian flow with most recharge from the limestone outcrop (many authors, including Tullis and Gries, 1938); B = rising thermal water with dispersed recharge from the higher Black Hills, shifting to C as more efficient down-dip routes become available (Bakalowicz et al., 1987; Ford, in Ford et al., 1988); D = rising basinal brines or gases (after Egemeier, 1981); E = augmentation of cave development by diffuse infiltration through the Minnelusa Formation; F = typical location of present springs. 1 = Precambrian rocks, 2 = Cambrian-Devonian strata, 3 = Mississippian Englewood and Pahasapa Formations, 4 = Pennsylvanian-Permian strata, 5 = Triassic-Jurassic strata, 6 = Cretaceous Inyan Kara Group.**

adjacent to the Black Hills, particularly to the north in the Williston Basin, rather than the Powder River Basin to the west (Paul Clikeman, Alexandria Bay, NY, personal comm., 1987; Andrew Flurkey, Houston, TX, personal comm., 1983).

However, the Black Hills caves contain no known sulfur or endellite, and all gypsum appears to have been derived from primary deposits within the local strata. The largest caves are located along the southern flanks of the Black Hills, where adjacent sedimentary basins are poorly developed. No large caves are yet known undip from the most likely basinal sources. A short distance to the southeast of Wind Cave, the Pahasapa and underlying rocks are entirely absent. Unless future work reveals more positive evidence, solution by H_2S or sulfuric acid is unlikely.

Water Source

The greatest controversy concerns the mode of water flow during cave development. There are several possibilities: (1) artesian flow fed by sinking streams and by dispersed inputs along the limestone outcrop; (2) diffuse recharge through the sandstone caprock; (3) water rising under thermal gradients and fed by remote surface sources; and (4) rising brines and/or gases from adjacent sedimentary basins.

Artesian Flow with Recharge from the Limestone Outcrop Belt

The Pahasapa Limestone surrounding the Black Hills is one of the best-documented artesian aquifers in the country. Were the caves formed by this artesian water, or does it merely follow preexisting caves formed in another way? The main argument for artesian cave origin is that this is the pattern of flow seen today, and there is no evidence that it has changed since the early Tertiary. Groundwater always follows the most efficient combination of steep hydraulic gradient and high permeability. Extensive Mississippian breccia and paleokarst provided high-permeability zones within the middle and upper beds of the Pahasapa, and most of the caves are located in these zones. Known caves are also concentrated around the outcrop areas where aggressive recharge is most expected.

The complex network pattern of the Black Hills caves is commonly attributed to an origin by artesian flow. Why artesian flow should produce a network pattern is never clearly stated, except that slow uniform flow would intuitively seem likely to produce uniform enlargement of all available openings. Palmer (1975) argued against this idea, stating that networks are favored by uniformly distributed recharge or greatly fluctuating flow, both of which are available in the Black Hills.

We originally favored an origin by water fed by sinking streams along the Pahasapa outcrop (Palmer, 1975, 1981). During low flow the head in the main conduits would be lower than in adjacent smaller openings, owing to greater conductivity in the former, and the flow would converge toward the main conduits. During high flow the main conduits would be the first to receive the increased recharge, and relatively unsaturated water would be forced into the surrounding openings. Repeated surges of this type are important in the origin of certain maze caves (Palmer, 1975). Violent floodwaters pour into some of the Black Hills caves today during heavy rain, and fresh organic debris in some low-level passages in Wind Cave attests to fluctuating water levels. Wind Cave and Jewel Cave are located along major Tertiary drainage lines that could have been recharge sources in the past, although when examined in detail these relationships are ambiguous.

Most water today sinks through alluvium in the stream channels, with overflow continuing downstream across the limestone outcrop area, so most sediment is filtered out of the infiltrating water. Sediment that enters through large openings is trapped where the inflowing water first reaches the slow flow in ponded parts of the aquifer. Still, the small volume of detrital sediment, the general lack of vadose passages superimposed on older phreatic ones, and the uncertain correlation with sources of concentrated recharge all argue against cave origin by groundwater fed by recharge from the Pahasapa outcrop.

Basinal Brine

The case against rising hydrogen sulfide has already been discussed, but it is still possible that the Black Hills caves were formed by rising carbonic acid from basinal sources. If so, it would have been necessary for sedimentary compaction to drive the water upward to the elevations at which the caves formed. Excess lithostatic pressure from compaction of the 1500 meters of sedimentary rock overlying the Pahasapa in the surrounding plains could have driven water to sufficiently high elevations, but the water would have become nearly saturated with dissolved limestone before it reached the site of the present caves. Mixing with near-surface water or substantial cooling would have been necessary to cause such dense cave development. No minerals diagnostic of basinal brines, such as saddle dolomite or ferroan calcite, have been found. Finally, as mentioned before, potential basinal sources do not match the location of the largest Black Hills caves.

Rising Thermal Water

An origin by rising thermal water is supported by the morphologic similarity of Black Hills caves to those of known hydrothermal caves elsewhere (Bakalowicz et al., 1987). For example, compare the features of Black Hills caves to the following ones observed in hydrothermal caves in the Atlas Mountains of Morocco (Collignon, 1983): three-dimensional network pattern, no evidence for vadose water, absence of allogenic sediments, bedrock weathering up to 20 cm deep, rounded domes having no fracture control, calcite wall crusts, and condensation corrosion. Domed ceilings, so common in the upper levels of Black Hills caves, are often considered the product of convection cells in thermal water (Rudnicki, 1978; Dubljanskij, 1980). Further evidence for thermal cave origin is the light isotopic ratios of oxygen and carbon in the calcite wall crust and boxwork of Wind and Jewel Caves, which fall into the same range as known hydrothermal calcite (Bakalowicz et al., 1987).

D. C. Ford, in his contribution to Ford et al. (1988), refined the thermal hypothesis by proposing that the cave-forming water infiltrated into the igneous and metamorphic rocks of the central Black Hills, flowed along deep paths and rose to the surface through the limestone. As the physiography changed, so did the outlets and sites of cave development. The caves seen today formed at the low points in the outcrop belt in the vicinity of major river valleys. More favorable outlets eventually developed downdip at the present springs, so cave development is now greatest in the limestone directly underlying them (Fig. 19).

The hypothesis of rising thermal water is geochemically most elegant. As thermal water rises through the limestone and cools, the zone of cave development would be below the water table where degassing of CO_2 is prevented by hydrostatic pressure. As the water table gradually drops past the caves, solution gives way to precipitation of calcite as degassing of CO_2 reduces the solutional capacity of the water. During the transition from one stage to the other, dolomite would continue to dissolve while the less soluble calcite would precipitate. This simple geochemical progression agrees well with what is seen in the caves. In comparison, other hypotheses seem unimaginative.

Such a well-tuned hypothesis is appealing, but close examination reveals some raw edges. The greatest drawback is the unlikely flow pattern required. Instead of following short, high-gradient paths along a zone of known high permeability, the water must follow long routes through nearly impermeable rock before rising into the sedimentary rocks. If stream valleys acted as the former outlets for thermal water, the caves should probably be focused more sharply on these hypothetical outlets. Boxwork and deep wall weathering, often considered evidence for deep cave origin, are now known to be independent of the mode of speleogenesis. Upwelling thermal water could account for the network pattern by uniform recharge through fractures in the underlying strata, but no known caves extend into the lower 20 m of the Pahasapa, except for a few minor passages. The domes and chimneys in upper levels may be former routes for water rising into the Minnelusa, but they could just as easily have been formed by rising artesian water fed from higher sources, as occurs today in the spring areas. There is also evidence that much, if not most, of the dome solution was performed by condensation water.

The distribution of caves is not related to thermal anomalies. Temperatures in Wind Cave rise 3° C in the 150 m from the entrance to the lowest levels, leading to the common impression that the cave is located over a local heat source. Actually this gradient is at the lower end of the typical range of 20–30°/km observed in caves and mines (Turcotte and Schubert, 1982). A true thermal anomaly is located near Hot Springs but has no relationship to known caves. Areas of Tertiary igneous intrusion and mineralization in the northern Black Hills are nearly devoid of caves. High-temperature minerals such as crystalline hematite and quartz pre-date the caves.

The ideal thermal model assumes that the calcite wall crusts were formed by the same water that formed the caves, with no subaerial event in between. Since the crusts are still forming today in Wind Cave in what is now an artesian aquifer, the thermal model requires a radical shift in hydrology in the past few million years without disrupting the depositional environment.

The fact that the oxygen and carbon isotopic ratios of many of the cave calcites fall in the same range as those of hydrothermal calcite does not necessarily guarantee a hydrothermal origin. Modern popcorn and some vadose travertines also fall in the same range (see Bakalowicz et al., 1987). Other factors must be considered, such as changes in isotopic fractionation of water in the past (especially by the active evaporation and condensation of the cave environment), recrystallization of calcite since deposition, and local heat production by exothermic reactions such as reduction of sulfates. Pierre and Rouchy (1988) found calcites in Egypt formed by shallow non-thermal replacement of sulfates to have oxygen-isotope signatures indicative of temperatures as high as 77° C. They tentatively attributed this isotopic anomaly to energy generated by sulfate reduction.

Calcite wall crusts fall in the same isotopic range as travertine in modern hot springs in the southeastern Black Hills, with typical $\delta^{18}O$ values of –14 to –16 $^{o}/oo$ PDB and $\delta^{13}C$ of -3 to -5 $^{o}/oo$ PDB; yet most of the springs are only about 20° C. The cave crusts probably formed at such slightly elevated temperatures, rather than in truly "hot" water, and may represent only the mild heating expected of stagnant water hundreds of meters below the surface at times when the normal cooling effect of rapidly circulating groundwater was absent.

Boxwork veins in Wind Cave also fall in the hydrothermal range, with $\delta^{18}O$ values as high as –18 $^{o}/oo$ PDB, and yet they are known to have formed at shallow depth in normal meteoric groundwater. This unusually light isotopic ratio is curious, as it does not match that of the orange calcite from Jewel Cave (mean $\delta^{18}O$ = -10 $^{o}/oo$ PDB), although their genesis has been shown to be contemporaneous and virtually identical. Possibly the isotopic ratios in the porous boxwork zones shifted in the presence of high-temperature water during deep burial, as the thin veins in porous zones would have had greater contact with circulating water than the massive, less permeable breccia matrix. On the other hand, calcite veins in the breccias above the main bedded chert, which are also contemporaneous with those of the boxwork, have $\delta^{18}O$ values of about -9 $^{o}/oo$ PDB, well above hydrothermal values. These breccias and veins were formed in an oxidizing environment in which sulfates had been completely removed by fresh water before calcite deposition. Within the Mississippian calcites, therefore, oxygen-isotope ratios are more highly negative in the zones of greatest sulfate concentration and in which sulfate reduction seems to have been most abundant.

Cave origin by rising thermal water, as well as by rising basinal brine, does not require a down-dip outlet for water. Therefore it is possible for solution by rising water to predate the artesian flow pattern. This advantage would apply only before the late Eocene, when the present outlets first became available. It is possible that one or both of these deep-seated process contributed to the early Tertiary enlargement of the caves before artesian flow developed.

Recharge through the Overlying Sandstone

Recharge through the sandstone caprock has never been seriously considered in the origin of the Black Hills caves because the caves reach the sandstone contact in only a few places. The carbonate content of the Minnelusa is fairly high, and water infiltrating through it loses much of its aggressiveness before entering the Pahasapa. However, solution of the Pahasapa could still be possible by the renewal of solutional aggressiveness by mixing of waters of differing carbon-dioxide content beneath the sandstone caprock. Many network mazes are suspected to have formed in this way. Mixing would be greatest in the high-permeability strata of the middle and upper Pahasapa, where the descending water would converge with the more abundant flow entering from the limestone outcrop belt. Both of these recharge sources are still active today.

Still, it is uncertain how the many passage levels in Wind and Jewel Caves could have formed by this kind of mixing. It seems likely that the aggressive water would have been limited to a rather narrow vertical range in the uppermost high-permeability zone. On the other hand, in areas of vertically stacked passages the passage size typically diminishes downward, and there is a high degree of interconnectivity between them. Also, most Tertiary caves were merely an enlargement of Mississippian porosity zones, which, because of their interconnectivity, would have allowed mixing over an unusually large vertical range.

Renewed aggressiveness by mixing is most effective where there is a large difference in the CO_2 concentrations of the two water sources. Water infiltrating through the sandstone would quickly encounter virtually closed chemical conditions, as it is isolated from the CO_2 of the soil. Solution of carbonates dispersed throughout the Minnelusa would bring the water close to saturation with respect to calcite and dolomite under closed-system equilibrium very low in both dissolved carbonates and CO_2 (Palmer, 1987). In contrast, water from the limestone outcrop belt is high in both dissolved carbonate and CO_2 (Miller, 1979, 1988).

Mixing and diffusion are slow enough in these conditions that undersaturation would persist for a longer time than if mixing were instantaneous. Thus, solution would be spread over large areas rather than focusing at the exact sites of convergence between the two waters. Small amounts of the low-CO_2 water would enter the larger amounts of high-CO_2 phreatic flow, and mixing would consist mainly of dispersion of both CO_2 and dissolved solids into the relatively pure water entering from above. Although limited space prohibits exact calculations here, it is easy to show that the amount of solution necessary to form the known density of cave passages (approximately 0.1–0.5 million cubic meters/square km) could be produced by a tiny infiltration rate (less than 1 cm/yr) in the millions of years the appropriate conditions persisted in the Black Hills.

Miller (1988) is skeptical of the ability of the Minnelusa to transmit water, citing the nearly total absence of vadose travertine. However, dye injected at point sources of recharge over Jewel Cave shows a large amount of dilution (Alexander and Davis, 1987), indicating much diffuse infiltration. The sandstone also does not offer much resistance to upward flow from the limestone in the discharge areas. Paucity of travertine does not imply an absence of infiltrating water; as explained in the section on bedrock weathering, most of this water gains aggressiveness, rather than losing it, as it enters the cave.

If the combined model of artesian flow and diffuse infiltration is valid, caves should be located (1) only beneath the broad, flat areas of thin sandstone (or where sandstone existed in the recent geologic past); (2) along the major phreatic drainage lines recharged from the limestone outcrop belt; (3) in the zone of greatest pre-solutional permeability; and (4) in laterally limited zones, with passages pinching out and varying greatly in cross-sectional area. All of these relationships are valid for the Black Hills caves, and no other hypothesis explains them so well.

Age of the Caves

The caves cut across the dogtooth spar that post-dates the Pennsylvanian paleo-fill, and they are also well adjusted to the structural and hydrologic pattern of the Black Hills. Therefore they are mainly of post-Laramide age, that is, less than about 65 million years old. Uranium/thorium age dating shows that the calcite wall crusts in Wind Cave range from modern to more than 500,000 years old (Bakalowicz et al., 1987; Ford et al., 1988). Those in Jewel Cave are older, but their exact age has not been determined. Since the wall crusts post-date the solutional phase, it is safe to say that most of the caves formed during the Tertiary Period. Reactivation of faults has displaced solutional cave walls as much as half a meter and has shattered calcite wall crusts, indicating that the caves pre-date the latest tectonic disturbance. The exact timing of cave origin is not yet known, partly because the reasoning used to determine the date of cave origin depends on which speleogenetic model is adopted.

From the standpoint of artesian flow, the most favorable time to form the Black Hills caves was late in the Eocene Period, when the limestone reached its present relief and the climate was humid. The only other feasible time would have been the late Pliocene, less than 5 million years ago, by which time erosion had removed much of the Oligocene sediment from the present spring areas. Although those favoring the thermal hypothesis prefer late Tertiary cave origin, rising thermal water would also have been favored during the Eocene because of the igneous activity, high relief, and availability of outlets along the limestone outcrop belt. Thermal anomalies persist today, but they do not disrupt the present artesian pattern of groundwater flow. Basinal fluid migration would have been favored during the early or middle Eocene, when the limestone was first exposed, or during the Oligocene or Pliocene, when thick sediments may have increased the lithostatic pressure rapidly enough to create overpressuring of the basinal water. Late Tertiary tectonic activity may have helped mobilize basinal fluids.

There is evidence for a lengthy stage of subaerial weathering separating the cave origin from deposition of the wall crusts (Deal, 1962; Palmer, 1981), but it is not universally accepted (Bakalowicz et al., 1987). Positive evidence includes the deep carbonate sediments containing evaporite pseudomorphs and overlain by wall crust, subaerial gypsum wedging of bedrock walls prior to deposition of the calcite wall crust, red weathering rinds and deep porous zones lithified by the first layer of wall crust, and crystal-coated stalagmites in Jewel Cave. If true, it implies early Tertiary cave origin followed by draining, with deposition of wall crusts later in the Tertiary and in the Quaternary when climatic oscillations and deposition of sediments around spring outlets caused frequent fluctuations in water level.

Conclusions

It is difficult to pinpoint the exact origin for the Black Hills caves because nearly all are inactive relics. Despite the abundance of field data, a fully convincing argument for any single mode of cave origin is still elusive. The most plausible is solution by artesian water, possibly enhanced by mixing with infiltrating water. We support this hypothesis mainly because it contains the fewest serious limitations. It considers the major cave solution to have occurred during the Tertiary Period by groundwater fed mainly by recharge from the limestone outcrop belt. Paleokarst and evaporite breccias afforded paths of high permeability through the limestone, and fracture zones and breccia pipes provided outlets, both past and present, for the water to rise through overlying insoluble strata. Renewed aggressiveness within the limestone aquifer was probably provided by mixing with water infiltrating through the sandstone caprock under quasi-closed chemical conditions, producing a strong contrast in CO_2 content. This combined model of cave origin best accounts for the distribution of caves and the nature of their passage terminations, but it does not explain superimposed cave levels very well.

Wall crusts, domed ceilings, and deep subaerial weathering might suggest an unusual cave origin, but the only implication is that the caves have been at least partly filled with stagnant supersaturated water late in their history, and that upward convection of warm, moist air has taken place in the absence of circulation between entrances at different levels. Sulfates dispersed locally within the bedrock also promote the kind of weathering observed in the Black Hills caves.

The resemblance to caves of deep-seated origin is misleading, because most of the features suggestive of a thermal or brine origin, such as the passage patterns, are inherited from Mississippian paleokarst and evaporite-related features. Much of the difficulty in establishing a widely accepted model of speleogenesis stems from this fact. The mode of Tertiary groundwater flow in fashioning the caves is less significant than originally thought. Any of the hypothesized groundwater models would tend to preserve and accentuate the existing porosity patterns rather than create new ones.

Evidence for this view is subtle but abundant: (1) known caves extend only through those parts of the Pahasapa containing paleokarst and evaporite recrystallization, fracturing, and brecciation; (2) the distribution of caves around the Black Hills matches the locations where Mississippian paleokarst and brecciation are most intense; (3) Mississippian breccia and boxwork are concentrated around the present caves but are sparse in non-cavernous zones in the same strata; (4) pre-paleokarst orange spar and granular quartz form resistant rinds outlining the present cave walls; (5) pre-Tertiary dogtooth spar partly lines some paleo-pockets, showing that they were not completely filled by Pennsylvanian sediment; (6) the difference in pattern between the upper and lower levels is accounted for by linear zones of breccia and fracturing below the bedded chert and by mixing-zone paleokarst above; (7) the present deep artesian flow is unlikely to have been established by normal karst solution without the advantage of considerable preexisting porosity; and (8) the uppermost passages form a crude level discordant to the bedding, apparently along a former water table.

If this is true, why is the pattern of the present caves so well adjusted to the shape of the Black Hills uplift? The dominant fractures that guide the caves radiate outward in exactly the pattern expected from Laramide deformation. Because the zones of breccia, boxwork, and orange spar are of Mississippian age, as are many of the fractures along which these features are aligned, the area must have experienced a minor amount of crustal movement during the Wendover phase of the Antler orogeny, which occurred immediately after the Pahasapa Limestone was deposited. Like many other large structural features of western North America, the basic outline of the Black Hills was probably established during the Precambrian and remained a slightly positive feature throughout the Paleozoic Era. One conclusion from this study is that the basic outlines of the caves, as well as of the Black Hills, were established long before these features are generally supposed to have formed.

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REFERENCES

Alexander, E. C., and Davis, M. A. 1987. Hydrogeologic study of Jewel Cave/Wind Cave, final report, year 2: National Park Service open-file report, 71 p.

Back, W., Hanshaw, B. B., Plummer, L. N., Rahn, P. H., Rightmire, C. T., and Rubin, M. 1983. Process and rate of dedolomitization: Mass transfer and ^{14}C dating in a regional carbonate aquifer: Geological Society of America Bulletin, v. 94, p. 1415–1429.

________, Hanshaw, B. B., and Van Driel, J. N. 1984. Role of groundwater in shaping the eastern coastline of the Yucatan peninsula, Mexico, *in* LaFleur, R. G. (ed.), Groundwater as a geomorphic agent: Boston, Allen and Unwin, p. 281–293.

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

Bowles, C. G. 1968. Theory of uranium deposition from artesian water in the Edgemont district, southern Black Hills: Wyoming Geological Association Guidebook, 20th Field Conference, p. 125–130.

Busby, J. F., Lee, R. W., and Hanshaw, B. B. 1983. Major geochemical processes related to the hydrology of the Madison aquifer system and associated rocks in parts of Montana, South Dakota, and Wyoming: U.S. Geological Survey Water-Resources Investigations Report 83–4093, 171 p.

Collignon, B. 1983. Spéléogenèse hydrothermale dans les Bibans (Atlas Tellien-nord de l'Algerie): Karstologia, v. 2, p. 45–54.

Conn, H. W., and Conn, J. 1981. The Jewel Cave adventure: St. Louis, Cave Books, 238 p.

Darton, N. H. 1918. Artesian water in the vicinity of the Black Hills, South Dakota: U.S. Geological Survey Water-Supply Paper 428, 64 p.

Davis, M. A., Alexander, E. C., and Alexander, S. C. 1988a. Water chemistry at Wind Cave and Jewel Cave [abs.]: National Spèleological Society, abstracts of 1988 annual convention, Symposium on Geology of Black Hills Caves, Hot Springs, SD, p. 12.

_________. 1988b. A chemical and isotopic investigation of the thermal springs of the southern Black Hills [abs.]: National Speleological Society, abstracts of 1988 annual convention, Symposium on Geology of Black Hills Caves, Hot Springs, SD, p. 12.

Deal, D. E. 1962. Geology of Jewel Cave National Monument, Custer County, South Dakota, with special reference to cavern formation in the Black Hills [M. S. thesis]: Laramie, WY, Univ. of Wyoming, 183 p.

_________. 1964. Scintillites: a variety of quartz speleothems: National Speleological Society Bulletin, v. 26, p. 29–31.

_________. 1968. Origin and secondary mineralization of caves in the Black Hills of South Dakota, U.S.A.: Proceedings of 4th International Congress of Speleology (1965), Ljubljana, Yugoslavia, v. 3, p. 67–70.

Dubljanskij, V. N. 1980. Hydrothermal karst in the alpine folded belt of the southern part of the USSR: Kras i speleologia (Univ. of Śl̓aski, Katowice, Poland), v. 3̓, p. 18–36.

Egemeier, S. J. 1981. Cavern development by thermal waters: National Speleological Society Bulletin, v. 43, p. 31–51.

Ford, D. C., and Bakalowicz, M. J. 1983. Preliminary report upon McMaster University geochemical and isotopic studies in Wind and Jewel Caves: open-file report, Wind Cave National Park, Hot Springs, SD, 16 p.

Ford, D. C., Palmer, A. N., and Palmer, M. V. 1988. U-series studies of calcite precipitation in lower Wind Cave, South Dakota [abs.]: National Speleological Society, abstracts of 1988 annual convention, Symposium on Geology of Black Hills Caves, Hot Springs, SD, p. 12.

Gott, G. B., Wolcott, D. E., and Bowles, C. G. 1974. Stratigraphy of the Inyan Kara Group and localization of uranium deposits, southern Black Hills, South Dakota and Wyoming: U.S. Geological Survey Professional Paper 763, 130 p.

Gries, J. P. 1969. Investigation of water losses to sinkholes in the Pahasapa Limestone and their relation to resurgent springs, Black Hills, South Dakota: Project completion report A-010-So. Dak. FY 1969, U.S. Department of the Interior, Office of Water Resources Research, 16 p.

Hackett, G. 1987. A review of chemical treatment strategies for iron bacteria in wells: Water Well Journal, Feb. 1987, p. 37–42.

Harksen, J. C., and MacDonald, J. R. 1969. Guidebook to the major Cenozoic deposits of southwestern South Dakota: South Dakota Geological Survey, Guidebook 2, 103 p.

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

_________ and Forti, P. 1986. Cave minerals of the world: Huntsville, AL, National Speleological Society, 238 p.

Howard, A. D. 1964. Model for cavern development under artesian ground water flow, with special reference to the Black Hills: National Speleological Society Bulletin, v. 26, p. 7–16.

Laury, R. L. 1980. Paleoenvironment of a late Quaternary mammoth-bearing sinkhole deposit, Hot Springs, South Dakota: Geological Society of America Bulletin, v. 91, p. 465–475.

Lisenbee, A. L. 1985. Tectonic map of the Black Hills uplift, Montana, Wyoming, and South Dakota: Geological Survey of Wyoming, Map Series 13.

Loskot, C. 1973. Deposition of cave material in Wind Cave [M.S. thesis]: Rapid City, South Dakota School of Mines and Technology, 66 p.

Lowry, D. C., and Jennings, J. N. 1974. The Nullarbor karst, Australia: Zeitschrift fur Geomorphologie, v. 18, p. 35–81.

Millen, T. M., and Dickey, D. N. 1987. A stable isotopic investigation of waters and speleothems in Wind Cave, South Dakota: an application of isotope paleothermometry: National Speleological Society Bulletin, v. 49, no. 1, p. 10–14.

Miller, T. E. 1979. Sampling of the atmosphere and carbonate aquifer at Wind Cave, South Dakota: Wind Cave National Park open-file report, 11 p.

_________. 1988. Wind Cave: an ancient artesian system modified by hydrothermal activity: Terre Haute, Indiana, Indiana State Univ., Dept. of Geography and Geology, Professional Paper No. 18, p. 16–29.

Mylroie, J. E. (ed.). 1988. Field guide to the karst geology of San Salvador Island, Bahamas: Mississippi State University, Friends of Karst, guidebook to 10th meeting, 108 p.

Ohle, E. L. 1985. Breccias in Mississippi Valley-type deposits: Economic Geology, v. 80, p. 1736–1752.

Owen, L. A. 1898. Cave regions of the Ozarks and Black Hills: New York: Johnson Reprint Corp. (1970 reprint), 228 p.

Palmer, A. N. 1975. The origin of maze caves: National Speleological Society Bulletin, v. 37, p. 56–76.

_________. 1981. The geology of Wind Cave: Hot Springs, SD, Wind Cave Natural History Association, 44 p.

_________. 1984. Jewel Cave—a gift from the past: Hot Springs, SD, Wind Cave/Jewel Cave Natural History Association, 41 p.

_________. 1987. Gypsum replacement by alternating open and closed systems in the vadose zone, Mammoth Cave, Kentucky: St. Louis, Cave Research Foundation, 1986 Annual Report, p. 27–28.

_________. 1988. Wind Cave: Hot Springs, SD, Wind Cave/Jewel Cave Natural History Association, 49 p.

_________, Palmer, M. V., and Queen, J. M. 1977. Geology and origin of the caves of Bermuda: Proceedings of 7th International Congress of Speleology, Sheffield, England, p. 336–339.

Pierre, C., and Rouchy, J. M. 1988. Carbonate replacements after sulfate evaporites in the Middle Miocene of Egypt: Journal of Sedimentary Petrology, v. 58, p. 446–456.

Plumley, W. J. 1948. Black Hills terrace gravels: a study in sediment transport: Journal of Geology, v. 56, p. 526–577.

Queen, J. M. 1973. Large scale replacement of carbonate by gypsum in some New Mexico caves (abs.): National Speleological Society, abstracts of 1973 annual convention, Geology Session, Bloomington, IN, p. 12.

Rahn, P. H., and Gries, J. P. 1973. Large springs in the Black Hills, South Dakota and Wyoming: South Dakota Geological Survey Report of Investigations 107, 46 p.

Retallack, G. J. 1986. Fossil soils as grounds for interpreting long-term controls on ancient rivers: Journal of Sedimentary Petrology, v. 56, p. 1–18.

Roth, J. E. 1977. Porosity evolution of the Pahasapa (Madison) Limestone at Jewel Cave National Monument, Custer, South Dakota [M. S. thesis]: Rapid City, South Dakota School of Mines and Technology, 103 p.

Rudnicki, J. 1978. Role of convection in shaping subterranean karst forms: Kras i Speleologia (Poland), v. 2, p. 91–101.

Sando, W. J. 1974. Ancient solutional phenomena in the Madison Limestone (Mississippian) of north-central Wyoming: U.S. Geological Survey Journal of Research, v. 2, no. 2, p. 133–141.

Sangster, D. F. 1988. Breccia-hosted lead-zinc deposits in carbonate rocks, *in* James, N. P., and Choquette, P. W. (eds.), Paleokarst: New York: Springer-Verlag, p. 102–116.

Schilberg, G., and Springhetti, D. (eds.). 1988. Caves and associated features of the Black Hills area, South Dakota and Wyoming: National Speleological Society, 1988 Annual Convention Guidebook, 156 p.

Shapiro, L. H., and Gries, J. P. 1970. Ore deposits in rocks of Paleozoic and Tertiary age, northern Black Hills, South Dakota: U.S. Geological Survey open-file report, 235 p.

Swenson, F. A. 1968. Recharge and movement of water in the artesian basin of the Dakotas: Wyoming Geological Association Guidebook, 20th Field Conference, p. 199–207.

Tullis, E. L., and Gries, J. P., 1938. Black Hills caves: Black Hills Engineer, v. 24, p. 233–271.

Turcotte, D. L., and Schubert, G. 1982. Geodynamics: New York, Wiley, 450 p.

White, W. B., and Deike, G. H. 1962. Secondary mineralization in Wind Cave, South Dakota: National Speleological Society Bulletin, v. 24, p. 74–87.

Ford, Derek C.—Features of the Genesis of Jewel Cave and Wind Cave, Black Hills, South Dakota: *NSS Bulletin* **51: 100-110.**

FEATURES OF THE GENESIS OF JEWEL CAVE AND WIND CAVE, BLACK HILLS, SOUTH DAKOTA

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Jewel Cave and Wind Cave, South Dakota, are here interpreted as multi-storey dissolutional mazes created during the present erosion cycle by deep phreatic waters that ascended through them. They formed where such groundwaters were focused to discharge through weaknesses in an overlying sandstone formation. The multi-storey structure is created by occurrence of different joint systems in adjacent beds or greater units, with inter-storey blocking layers such as thin clays often playing a role. In such structures, lower storeys tend to be more extensive; upper storeys may contain both outflow *and* adventitious *components. Mixing corrosion effects and migration of springs can complicate upper storeys. As such caves drain, condensation corrosion facets and pockets may be created.*

From U series dating and magnetic studies of normal speleothems and of subaqueous calcite encrustations, Jewel Cave drained more than 350,000 years ago. Its characteristic subaqueous spar sheets are certainly older than 1,250,000 years and probably greater than 2,500,000 years in age. Wind Cave has drained within the past 500,000 years or so. The mean rate of fall of the watertable in it is ~ 0.375 m per thousand years but the actual rate of fall has varied probably in response to Quaternary climatic fluctuations. Studies of stable isotope ratios indicate that the subaqueous deposits were precipitated from waters warmed to a probable range, 15–50°C. It is most likely that such waters were responsible for excavating the bulk of the caves as well, although there are older paleokarst remnants locally; thus, genesis of the caves is to be attributed to the type and scale of thermal water systems that feed the present hot springs in the Black Hills. Combinations of several different factors may account for the differing form, thickness and depth of deposition of the subaqueous calcites, including regulation of the rate of de-gassing by presence of caprocks.

INTRODUCTION

Jewel Cave and Wind Cave are complex cave systems that are of great interest to physical speleologists. This paper discusses four different but associated aspects of their complexity. Where it is relevant they are compared to maze caves in other regions. Readers who are not familiar with the Black Hills caves and their geology should first read the general reviews that A. N. and M. V. Palmer have prepared for this Symposium.

My knowledge of Jewel Cave is confined to the tourist paths plus short excursions from them. At Wind Cave I have also had several trips down to the Lakes. However, since 1982 my research group at McMaster University has been foremost in the isotopic analysis of samples of the characteristic cave calcites and the wall rock; therefore, conclusions and speculations derived from those analyses will be emphasized. Most samples were of small, naturally broken and fallen, speleothems whose precise site of origin could be established. At Jewel Cave there were also two samples of broken calcite spar that had been pushed aside as rubble during construction of the tourist path. Their sites of origin are uncertain but most probably lay within a few metres of the points where they were picked up.

MORPHOGENESIS OF THE CAVES

The first aspect is the gross morphology of these two systems of caves. They belong to a class that may be termed "lifting mazes." In it the formative waters flow upwards from a lower joint system into an upper and different system that immediately overlies it. There may be several more of such superimposed systems; each of them constitutes one 'storey' in a multi-storey cave.[1] Typically, the joint systems in successive storeys display different frequencies or orientations, with the result that usually they can be differentiated on a map. Another general characteristic is that the passages following one particular joint set (orientation) in a given storey tend to be roughly equal in their cross-sectional area. This demands that the formative waters be phreatic and flowing in an unusually uniform pressure field for karst rocks, as W. M. Davis recognized long ago (Davis, 1930).

The most extensive examples of such lifting maze caves are Jewel and Wind, and the mazes of the Podolia and Bukovina districts of the Ukraine. The latter include Optimists' Cave (165 kms as at December 1988), Ozernaya (105 kms), Zolushka (82 kms), and many lesser examples such as Atlantida Cave (Fig. 1). They are developed in just 12 to 18 metres of gypsum that rests upon a limestone and sand

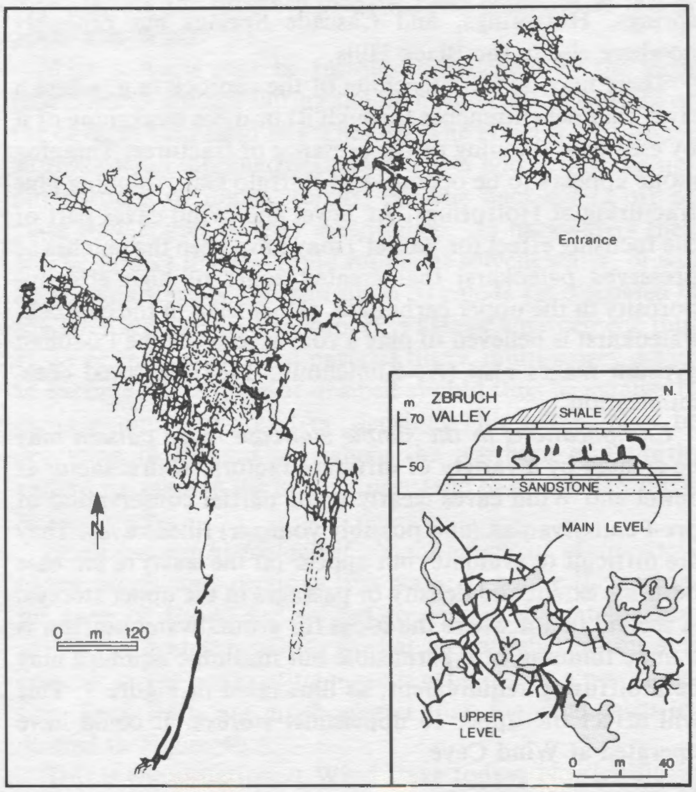

Figure 1. **Two examples of lifting mazes from the gypsum karst of the Ukraine, U.S.S.R. Optimists' Cave (principal illustration) now possesses 167 km of surveyed passages beneath an area less than one square km in extent. Atlantida Cave (inset) is a simpler example, with a large lower storey (or main level) and a fragmentary upper storey. (Adapted from Klimchouk and Andreichouk, 1986).**

aquifer and is capped (principally) by an impervious marine clay. They were created by shallow meteoric waters ascending from the sand stratum to spring lines that developed where rivers entrenched the clay (Klimchouk and Andreichouk, 1986).

Many shorter examples of lifting mazes occur at Budapest, Hungary. These are developed in clayey limestones capped by largely impervious marls and were created by deep thermal waters that ascended through them to springs along the Danube. There were possible mixing corrosion effects also. These caves have been intensively studied by Hungarian colleagues (cf. Muller and Sarvary, 1977). A final group of examples I have visited are the caves of McKittrick Hill, N.M. (Kunath, 1978). These are in well-bedded backreef limestones that are capped by a calcareous sandstone and were created by ascending meteoric waters or mixing waters with an H_2S component from the Pecos sedimentary basin (Hill, 1987).

From these examples a first conclusion we may draw is that the fundamental erosional morphology of lifting mazes

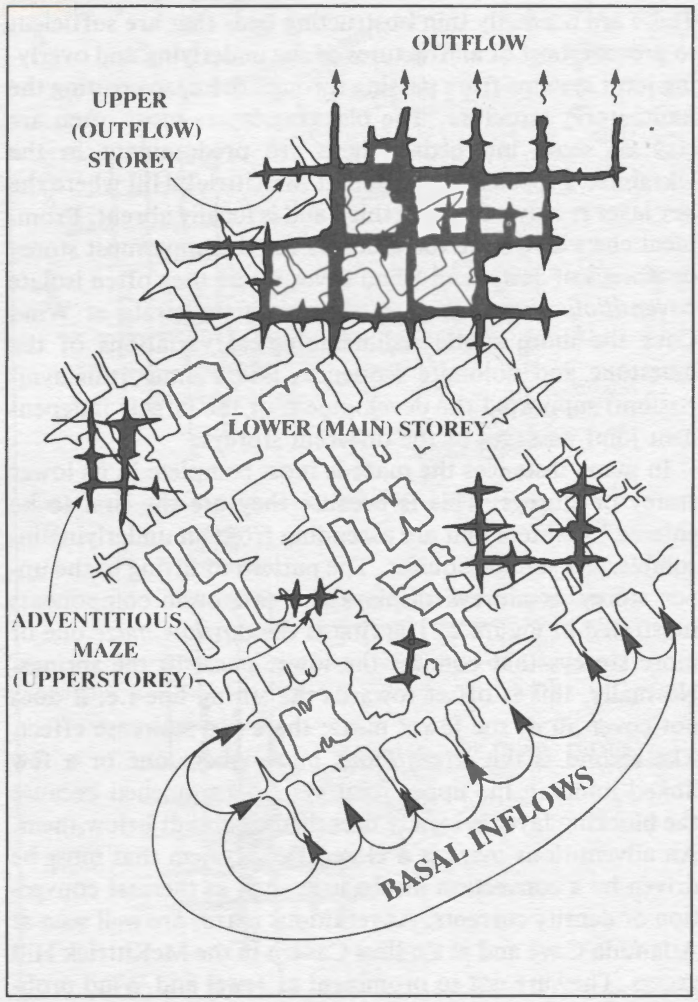

Figure 2. **The characteristic form and location of *outflow mazes* and *adventitious mazes* in upper cave storeys that are fed by water lifting from one or more lower storeys.**

can be created by a variety of groundwaters circulating through a variety of rocks in many different tectonic, geomorphic and climatic settings. What, then, are the significant features that they have in common? There appear to me to be three that are the most important:

1. *Occurrence of an underlying water source.* In most cases this is probably deeply circulating regional recharge in fracture aquifers (e.g. at Budapest; Muller, 1989). For Jewel and Wind caves the recharge area is the granitic and metamorphic core of the Black Hills, plus some of the exposed limestone. Volumes of flow generated per square km are usually small in such cases, with the consequences that discharge via the maze caves will be small and velocities in them will be low. In the Ukraine the recharge is a matter of more local, inter-valley flow between river channels at differing elevations (Klimchouk and Andreichouk, 1986).

2. *Presence of one or more inter-storey blocking layers.*

These are normally thin obstructing beds that are sufficient to prevent most or all fractures of the underlying and overlying joint systems from passing through them, so creating the multi-storey structure. The blocking layers most often are clay or shale interbeds. These are predominant in the Ukrainian gypsum mazes, and at McKittrick Hill where the key layer is only a few cms thick and is locally absent. Prominent chert layers are the blockers for the uppermost storey or storeys of Jewel and Wind caves where they often isolate *adventitious mazes* (Fig. 2). Lower in the strata at Wind Cave the more subtle sedimentological variations of the limestone and dolomite sequences (see Palmer, this symposium) supported the development of the largely independent joint passages of the different storeys.

In many instances the maze is most complete in its lower storey or storeys. This is because they are the first to be entered by waters that are ascending from an underlying insoluble/less soluble aquifer. The pattern of lifting to the upper storey or storeys displays the two basic components illustrated in Figure 2. The first is the *outflow maze,* one or more storeys that connect the lower maze to the springs. Normally, this is offset towards the spring line i.e. it does not cover all of the lower maze; there is a staircase effect. The second is the *adventitious maze* where one or a few linked joints in the upper joint system are opened because the blocking layer is locally breached or absent below them. An adventitious maze is a closed flow system that must be driven by a convection mechanism such as thermal convection or density currents. Adventitious mazes are well seen at Atlantida Cave and at Endless Cavern in the McKittrick Hill group. They are not so prominent at Jewel and Wind probably because the joints in the successive storeys are closely aligned and because mixing corrosion effects have extended the upper storeys there (see below). However, simple examples displaying cupola-form pocketing develop above the chert band at the Fairgrounds (Wind Cave) and elsewhere.

3. *Presence of a caprock* is also essential. The caprock is either insoluble or significantly less soluble than the maze cave rock and it functions either as an aquitard (less pervious) or an aquiclude (impervious). It is a barrier to the discharge of the artesian water. It has a much greater areal extent than the maze caves that develop beneath it. Where it is offset by faulting or attacked by subaerial erosion processes, etc. breaches or zones of lesser weakness will inevitably be created in it. The deep groundwaters are focused to discharge via those zones and they create the maze caves toward and beneath them. This is to assert that lifting maze caves are not present everywhere in the Pahasapa Limestone of the Black Hills but only where sufficient ascending flow could be concentrated. Such focusing of flow is quantitatively essential to create them, given that the rate of recharge to the deep aquifers is very low. Thus in my opinion modern lifting mazes are being created beneath the Buffalo Gap springs, Hotsprings, and Cascade Springs but probably nowhere else in the Black Hills.

There may be full breaching of the caprock (e.g. where a river channel entrenches through it) or mere weakening of it by erosional thinning or the presence of fractures. Thinning alone appears to be operative at Buffalo Gap, thinning plus fracturing at Hotsprings. At Jewel and Wind caves part of the focusing effect (or 'target') may have been the patches of preserved paleokarst that created zones of high effective porosity in the upper carbonate, rather than in the caprock. Paleokarst is believed to play a role in some of the Podolian gypsum mazes also (A. Klimchouk, 1987; personal communication.

Complications in the simple staircase maze pattern may be created by a variety of different factors. A first factor at Jewel and Wind caves clearly is the partial conservation of pre-Pennsylvanian (and possibly younger) filled caves. They are difficult to evaluate[2] but appear (at the least) to increase both the extent and density of passages in the upper storeys. A second factor where the focus for groundwater outflow is a mere thinning of a permeable but insoluble aquitard may be a diffusing requirement, as illustrated in Figure 3. This will affect the upper to uppermost storeys. It could have operated at Wind Cave.

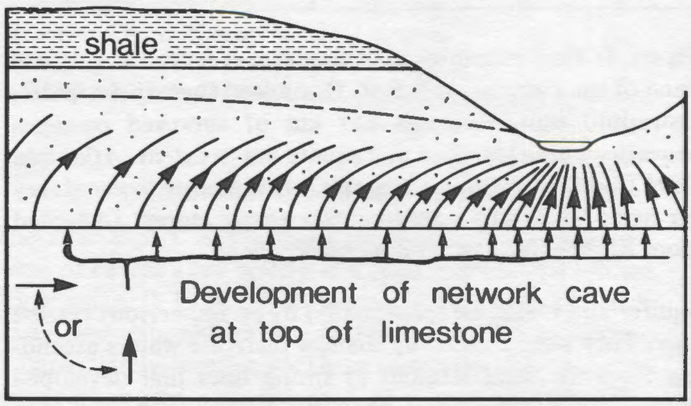

Figure 3. **Water flowing upwards in dissolutional conduits in limestones is compelled to diffuse in order to discharge through a sandstone cover. An upper storey maze at the top of the limestone marks the start of the diffusing process.**

Mixing corrosion is a third factor that is probably much more widely important in carbonate rock mazes. It is especially powerful where deep, warmed waters enriched in exhalative CO_2 and/or H_2S from reducing processes encounter cool meteoric waters with dissolved O_2. Figure 4a suggests that at Jewel and Wind there was potential for meteoric water supply both down dip and along strike to mix with the ascending waters. Mixing corrosion normally should be most effective in upper to uppermost storeys because meteoric waters will not circulate far below the caprock, but the presence of paleokarst complicates this at Jewel and Wind.

Maze patterns may be further complicated by the local migration of springs during the hydrologically active life of the cave, as suggested in Figure 4b. The effect is that upper storeys (or, in dipping rocks as at Jewel and Wind, the up-dip portions of upper storeys) become backwaters off of main flow routes or they are drained. As backwaters they will be subject to a less rapid supply of solvent water, or will switch to a state of net deposition of calcite (as discussed in the later sections of this paper). Passage enlargement thus slows or ceases. The first parts of tilted, multi-storey mazes to become backwaters or drained should thus be smaller in passage dimensons and/or density. This effect may be operating in Wind Cave where the northern and north-eastern passages in the upper storey tend to be smaller; dip is southeasterly.

Drainage and hydrological abandonment of multi-storey lifting mazes may occur in two ways: 1) where the springs (or zone of focus) are shifted to a new, distant location within the host formation, so that an entirely new maze is created, or the pre-existing maze is extended in a new direction. Such shifts are much greater than the local shifts indicated in Figure 4b.

This is the situation at Wind Cave today. No significant eroding waters flow in it but the lowest storey dips below a regional water table to become phreatic. It is an up-dip backwater. The springs probably shifted to Buffalo Gap, 8 km distant, but the pioneer dye tracing work that Professor Calvin Alexander is undertaking hints that the shift could have been to Hotsprings, 12 km away (C. Alexander, personal communication, 1988). 2) where the base level of erosion is lowered below the rock formation hosting the cave; in most instances, this means that the river creating the spring line valley has entrenched its channel below the formation. That is the case for a majority of the Podolian mazes. Jewel Cave appears to be essentially in this situation also. We cannot be sure because exploration has not yet attained the base of the host formation there. However, any basal standing water that is encountered in the future seems likely to be of local meteoric origin and perched, rather than backwater on a regional water table as at Wind Cave. In the past, however, Jewel Cave appears to have been a backwater for a very protracted period, as evidence cited in the next section suggests.

A note on condensation corrosion. This is a cave-enlarging process insufficiently appreciated by English-speaking speleologists of my generation, perhaps because most of us trained in cool, humid meteoric water caves where it is not of much significance. Carol Hill has stressed it recently from her work in the warm, arid Guadalupe caves (Hill, 1987) but most work comes from Hungarian and Soviet speleologists working in thermal water caves or those

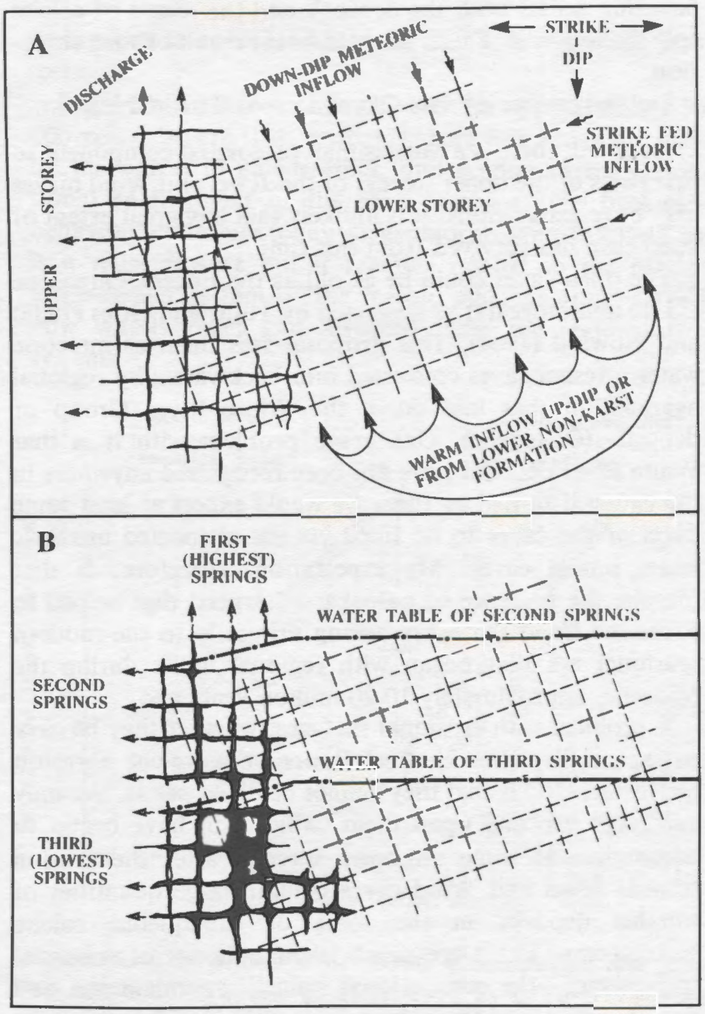

Figure 4. A. **An illustration of the potential for mixing corrosion to occur in geological situations such as those at Jewel and Wind caves. Mixing corrosion will preferentially create or enlarge upper storeys in a multi-storey maze.** ***B.*** **Illustrating the effect that local shifts of springs may have upon the enlargement of conduits in upper storeys.**

of the semi-arid Asian regions of the U.S.S.R. Of particular importance is theoretical modelling by Szunyogh (1984) who suggests that highly rounded (cupola-form) ceiling solution pockets in thermal caves are created by this process. Condensation corrosion would appear to be most potent where hot, CO_2-surcharged waters degas into confined caves. The saturated air is quickly condensed onto cooler upper walls.

Wind Cave displays cupola-form pockets above the chert layer at the Fairgrounds and elsewhere. These may be adventitious phreatic maze components, as suggested above, but they warrant consideration as features of condensation corrosion. More convincingly, around the Loft and other uppermost parts of Jewel Cave there is smooth dissolutional bevelling across both the bedrock and the sheets of calcite spar that cover it. This is likely to be the result of such corrosion.

THE ANTIQUITY OF THE CAVES AND OF THEIR DRAINING

As noted, there is a Mississipian paleokarst component so that parts of the upper storeys of the Jewel and Wind mazes may be re-excavations. It is unlikely that any great extent of open cave has survived from this time.

The open caves could be as old as the Eocene-Oligocene (35-40 million years) as suggested by Tullis and Gries (1938) and Howard (1964). This proposal sees them as meteoric water artesian caves converted into backwaters by regional aggradation that laid down the White River Group of deposits (Oligocene). One grave problem with it is that White River deposits have not been recognised anywhere in the caves; if buried by them we would expect at least some parts of the caves to be filled via the purported meteoric water intake caves. My expectation, therefore, is that (despite the presence of paleokarst 'targets' that helped to focus the flow) the caves belong primarily to the modern erosional era that began with regional uplift during the Miocene, approximately 10-20 million years ago.

A problem with erosional surfaces (whether they be cave passage walls, river channel floors or wave-cut abrasion platforms, etc.) is that they cannot be dated *per se*. We may date only deposits upon them, which will have begun to accumulate at some unknown interval after the erosion ceased. Jewel and Wind caves contain large quantities of suitable deposits in the form of subaqueous calcite speleothems. There are much lesser volumes of subaerial speleothems—the conventional stalactites, stalagmites and flowstones. All of these deposits may be dated by U series methods (Harmon and others, 1975; Ivanovich and Harmon, 1982) if they contain a sufficient concentration of trace uranium and are not highly contaminated by detrital clay.

Jewel Cave

Sample JC11 is a fragment from a large stalactite curtain that had fallen and shattered in a grotto in the lowest part of the tourist cave. Isotopic details of it and of other samples representative of those discussed here are given in Bakalowicz and others (1987). ^{230}Th:^{234}U dating reveals that its outer part grew around 100 000 years ago (100 ka). The inner part is older than 350 ka (the dating limit of the standard method), but not much older. This is from one of the largest subaerial speleothems seen in the tourist cave, which suggests the broad generalization that most conventional speleothem deposits in Jewel Cave may be younger than 500 ka. More certainly, JC11 indicates that the tourist cave has been drained and relict for at least 350 ka.

More significant are the sheets of subaqueous calcite spar that cover most of the walls. Their thickness is generally between 6 and 15 cm. I have studied eight samples taken from the highest to lowest occurrences close to the tourist path. All are similar, displaying euhedral form and very regular layering. They contain between 0.2 and 0.5 ppm (parts per million) of uranium, which is significantly less than in any other Jewel and Wind deposits. For example, stalactite JC11 has 1.7-7.5 ppm U which its meteoric feedwaters scavenged from the overlying paleokarst fill or Minnelusa sandstone. The waters that deposited the calcite spar cannot have followed that overhead route to their depositing site. This is the first of many separate geochemical and isotopic evidences tending to show that the spar waters are from deeper sources.

All spars are greater than 350 ka in age i.e. ^{230}Th:^{234}U = 1.00. In fact, within measuring error the ratio, ^{234}U:^{238}U, also equals 1.00. For technical reasons that cannot be discussed here (see Ivanovich and Harmon, 1982, page 68) an excess of ^{234}U atoms is normally precipitated (trapped) in cave calcite. Any excess at Jewel has decayed away. This suggests that the spars are older than 1 250 000-1 500 000 years (1.25-1.5 ma).

Spar samples from the lowest part of the tourist cave are particularly interesting because they display a systematic depletion in ^{234}U. The ratio, ^{234}U:^{238}U, is slightly less than 1.00 at their base (oldest part) and becomes lower towards the top (youngest part). This is significant for two reasons:

1) it suggests a principal source aquifer that was largely insoluble so that ^{234}U (the more readily erodible isotope of the ^{234}U,^{238}U pair) became depleted in the walls that enclosed its flow paths i.e. flow had been very long sustained before any deposition of the spar began. As shown by the U concentrations and by stable isotope evidence for the spars given in the next section, this source aquifer was most probably beneath the limestone and it supplied warm water.

2) a 'daughter deficient' dating technique can be applied to the spars (Ivanovich & Harmon, 1982, p. 61). This does not give their absolute age but does suggest the time that elapsed between deposition of the basal spar (least deficient) and that at the top. The calculated growth period for a sample 6 cm in thickness is approximately one million years i.e. a mean deposition rate of 0.06 mm per 1000 years. As noted, the texture of the spars suggests that they accumulated steadily rather than in an episodic manner, so that this mean rate is a significant one. It is very low indeed when compared to meteoric water speleothem deposition rates (Hill and Forti, 1986); it helps build a picture of very slow deposition from waters rising from an ancient aquifer. This is extended into a general model in the last section of the paper.

One of the daughter deficient spar samples from a low elevation in the cave was tested for its magnetic signal by Dr. A. G. Latham, the pioneer of magnetic studies of speleothems (Latham, 1981). The entire sample is magnetised normally (i.e. positive North). From the ^{234}U:^{238}U studies it is known that it is older than 1.25 ma and that it took ~1.0 million years to accumulate. The youngest available 'normal' magnetic interval of sufficient duration is the "Gauss," which extended from 2.5 to 3.4 ma BP. It is therefore concluded that the principal wall deposits in Jewel Cave most probably accumulated at some time before 2.5 ma BP.

Wind Cave

Wind Cave is 400 m lower in elevation than Jewel Cave, much closer to the major regional springs, and its lowest known parts are occupied by static backwaters today. It is expected to be significantly younger than Jewel Cave and this will be shown to be the case. Its quantities of precipitates are much smaller but have proven to be most useful for paleodrainage reconstruction.

Those parts of the cave that are highest in elevation are without calcite deposits except occasional stalactites from meteoric waters. As we descend downdip to upper storey areas such as the Fairgrounds (~1220 m above sea level) small nodular and sheet encrustations appear. Most of them are of subaqueous origin although there is some redeposition as evaporites. The abundance and thickness of subaqueous sheets increases markedly on the descent to the Lakes (lowest storey, 1120 m above sea level) perhaps being thickest in the area between Base Camp 1 and the L.A. Freeway. Locally the greatest thickness appears to accumulate on protrusions into the centres of passages where the rate of paleoflow will have been greatest; they are up to 4 cm thick (N.B. the same feature is seen in some of the Budapest thermal water caves). From isotopic analysis of two shards, the famous heligmite bushes such as "the Emperor Maximus" are also subaqueous growths. From Boxwork Chimney (1180 m) downwards there are accretions of calcite raft debris marking paleowaterlines. They continue to accumulate on the water table (Calcite Lake) today.

I have made more than 90 ^{230}Th:^{234}U determinations on 57 samples of crust and raft debris collected between the Fairgrounds and the modern water line. Their detailed analysis and the hydrological implications are being presented in a separate paper (Ford, Palmer, and Palmer *in litt.*). The principal conclusions are summarised in Figure 5.

The main point is that the lower 100 m of Wind Cave (i.e. the majority of its known passages) has drained within the past 300 000 years or so. From the highest dated deposits we may track this process of water table fall extending back beyond 500 ka BP—at that time most or all of the cave will have been water filled. The mean rate of fall has been ~0.375 m per 1000 years. This cannot be attributed to progressive lowering of the spring elevation—it is too rapid for that and the magnitude of the lowering is too great. Rather, most of the lowering must be attributed to increasing permeability in the Minnelusa sandstone that the groundwaters must ascend through to gain the modern spring points (whether these be at Buffalo Gap or Hotsprings).

Wind Cave has been a backwater for approximately 500 000 years; before that, probably there was active discharge up through it to paleosprings, and accompanying dissolutional enlargement of the cave passages. The backwater behaviour continues today. In geomorphic terms Wind Cave is a relict feature but it remains related to the modern groundwater system of the region, whereas Jewel Cave is now wholly relict and divorced from the regional flow systems.

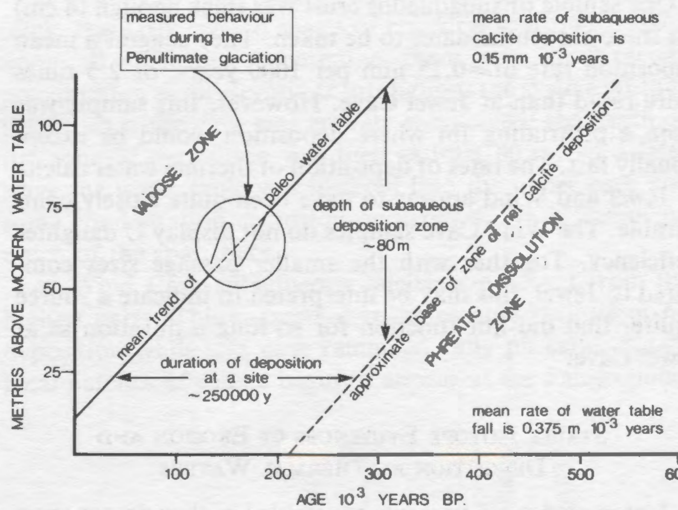

Figure 5. **The drainage of Wind Cave during the past 300,000–500,000 years, as determined by ^{230}Th:^{234}U dating of waterline and subaqueous calcite deposits.**

Figure 5 emphasizes several other features. The fall of the water table has not been steady. It has fallen about 10 m during the last 10 000 years (the Holocene period), a rate of ~1.0 m per 1000 years. At least half of this fall might be attributable to pumping of the aquifer; there is an important well 2 km distant. More significantly, dating of wall crusts, rafts and pool rims at the base of Boxwork Chimney show the water table falling, rising, then falling again through a range of 10 m at least once in the interval between (broadly) 180 and 130 ka BP. This is the upper part of the penultimate glaciation of North America (often termed the 'Illinoian Glaciation' in American literature although the type deposits in Illinois are not accurately dated). This water table fluctuation was probably a response to more arid and then more humid periods inside that cold phase. Or it could be a response to some non-systematic events such as a single catastrophic flood, though this seems less likely.

It may be supposed that detailed work would reveal more of these fluctuations during the measured decline of the water table. They will explain the deposition of subaqueous crusts following vadose events such as the fall of grus (pulverised dolomite) and other detritus in the boxwork zone.

From a strict interpretation of the dating results there was contemporaneous deposition of calcite from the water surface to a depth of 80 m beneath it; this is the "subaqueous calcite deposition zone." Below this depth no deposition was possible and, perhaps, late stage and/or incongruous dissolution of the rock was occurring. In my opinion, the value of 80 m is a maximum. The true depth of deposition is more likely 50-60 m, being extended to the apparent maximum depth of 80 m by unresolved fluctuations of the water table and error margins in the dating.

One sample of subaqueous crust was thick enough (4 cm) for three sequential dates to be taken. They suggest a mean deposition rate of ~0.15 mm per 1000 years, or 2.5 times more rapid than at Jewel Cave. However, this sample was from a protruding fin where deposition would be exceptionally fast. The rates of deposition of thermal water calcite in Jewel and Wind appear to have been quite closely comparable. The Wind Cave samples do not display U daughter deficiency. Together with the smaller passage sizes compared to Jewel, this may be interpreted to indicate a source aquifer that did not function for so long a duration as at Jewel Cave.

STABLE ISOTOPE EVIDENCES OF EROSION AND DEPOSITION BY THERMAL WATERS

Isotopes of most elements are stable i.e. they do not spontaneously emit nucleons from their nuclei. Stable isotopes of interest here are those of C and O that make up the CO_3 (carbonate) in the speleothems and the wall rocks of the caves. We compare the ratio of the rare isotope, ^{13}C, to the abundant isotope, ^{12}C, and similarly ^{18}O (rare) to ^{16}O (abundant). Different ratios in different samples may reflect the dominance of different processes, or the same process operating at different temperatures.

Bakalowicz and others (1987) have reported in some detail upon $^{13}C/^{12}C$ and $^{18}O/^{16}O$ relationships in Jewel and Wind caves. The subaqueous calcites display a distinctive thermal water isotopic signal when compared to the ordinary (meteoric water) stalactites and stalagmites. There are also alteration trends in the wall rocks.

These results are summarised in Figure 6. Normal marine limestones and dolomites (such as those containing Jewel and Wind caves) will plot within ± 5‰ of the origin of the figure. The "Hydrothermal Calcite Box" is drawn as a boundary to enclose all previously published values for North American calcites that were unquestionably deposited from thermal waters (e.g. from the gour terraces at Yellowstone National Park; Truesdell and Hulston, 1980). In comparison to the marine rocks, these calcites are somewhat depleted in ^{13}C and strongly depleted in ^{18}O.

The Jewel and Wind subaqueous calcites all plot within 'the Box.' The Wind Cave boxwork calcites are 'deepest' in it: from a temperature relationship proposed by O'Neil and others (1969) they were deposited from waters between 30° and 60° C. They include the oldest thermal deposits in Wind and (it is argued below) were deposited from waters still capable of dissolving dolomite. The calcite crusts and ices of lower Wind Cave are similar to the younger boxwork or a little less depleted in ^{18}O, a relation consistent with the supposition that they were deposited from progressively cooling, progressively falling, backwaters.

Isotope ratios of the Jewel Cave spar sheets form a tightly clustered group that plots on the trend between unaltered bedrocks and Wind Cave deposits. They are only just within 'the Box' and yield O'Neil temperatures between 15° and 35° C. This implies cooler waters from a longer groundwater flowpath, which fits very well with the picture of a long-lived, insoluble source aquifer that was obtained from the ^{234}U:^{238}U data on these same spars.

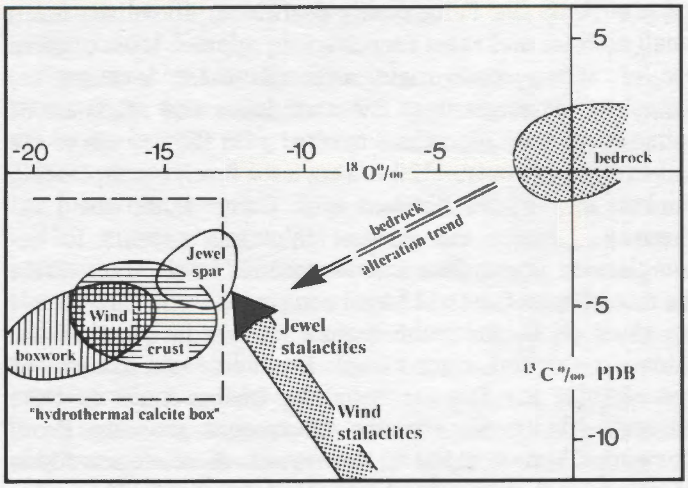

Figure 6. **Carbon and oxygen stable isotopic data for sample bedrocks, subaqueous calcite deposits and normal (vadose) speleothems from Jewel and Wind caves. Generalized from data reported in Bakalowicz and others, 1987. See the present text for discussion.**

We investigated rock samples from the walls of the caves and determined that their outermost few cm had been altered by isotopic exchange with the hot waters. There is a very clear trend (see Bakalowicz et al., 1987, Fig. 6). The Jewel Cave rock is the more deeply and extensively altered, in keeping with the longer duration of its exposure to the waters.

Conventional stalactites and stalagmites in Wind Cave plot around -10‰ ^{13}C and ^{18}O. This is well outside of 'the Box' and in equilibrium with the modern air temperatures in the cave. These are ordinary deposits from meteoric waters. However, Jewel Cave stalactites are notably enriched in ^{13}C

(–8 up to –5‰) and somewhat depleted in ^{18}O. This suggests that their meteoric source waters have exchanged isotopes with rock that is thermally altered to quite considerable depths along the microfissures that those source waters must flow through. Once again, this observation fits with the picture of longer exposure to thermal waters at Jewel Cave.

That the stable isotope results show that the subaqueous deposits were precipitated from thermal waters does not prove that the caves were largely or entirely excavated by such waters, but it is strongly suggestive of the point. Figure 7 emphasizes this. The boxwork is calcite encrustation (overgrowth) upon vein calcite in dolomitic beds that are densely fractured. It formed in thermal waters capable of dissolving dolomite (Mg ion), but only with concomitant deposition of Ca ion (= "incongruous dissolution of dolomite"). In the boxwork zone of Wind Cave it is common to find boxwork 30 cm deep in passages one metre wide (Fig. 7a); 60% of the width of such passages must have been excavated by the thermal waters and we may confidently suppose that 100% of it was. Such a direct relationship cannot be observed in wider passages because surviving boxwork is rarely deeper than 30–50 cm (Fig. 7b).

In summary, temperature and isotopic conditions essentially identical to those measured in the hot springs of the Black Hills today readily explain the subaqueous deposits. Some of these latter are probably greater than 2.5 million years in age, others are forming now. Ages of the deposits indicate that the caves were slow to drain, and probably slow to form also. Cave excavation thus is concordant also with the amounts of water being discharged at the thermal springs today. It does not appear necessary to postulate some precursor, radically different, hydrological system to explain their formation.

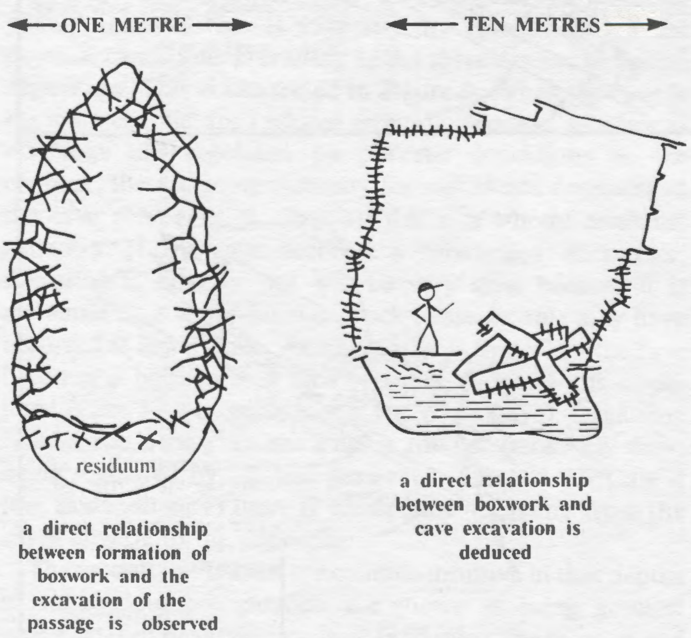

Figure 7. **Relationships between the deposition of boxwork and the dissolutional excavation of passages.**

INTERPRETING THE DEPTHS OF SUBAQUEOUS CALCITE DEPOSITION IN THERMAL WATER CAVES

The basis of this final topic is illustrated in Figure 8. This depicts the modes of subaqueous precipitation and the depths to which it extended at a given time, in the three groups of thermal water maze caves that I have had an opportunity to study.

At Jewel Cave the deposits consist almost exclusively of thick, regularly layered sheets of euhedral spar that formed upon all surfaces. Apparently, there was simultaneous precipitation at all elevations throughout the system. It thus began when the cave was entirely submerged; the depth of the zone of deposition was then at least 80 m. At a later stage the water table was lowered into the upper cave and some condensation corrosion occurred above it, removing much of the spar there.

At Wind Cave subaqueous calcites are absent from the highest parts. This suggests that there could be little or no deposition while this cave remained fully phreatic. Small, local patches of calcite begin to appear at the Fairgrounds,

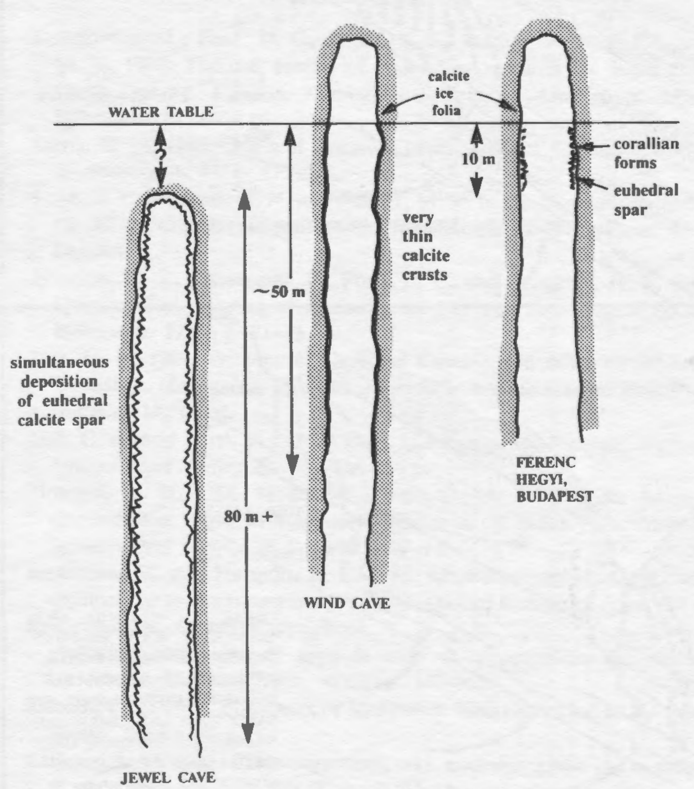

Figure 8. **Types of subaqueous deposits and the depths of their deposition at Jewel Cave, Wind Cave and a Budapest cave. The figure is simplified and schematic.**

and increase in thickness and extent below them. The calcites are always thin, and contain some interlaminations of red clay apparently reworked from the paleokarst fillings. In the lower half of the cave there are also calcite raft debris and shelfstone to mark the rest positions of a falling water table. The depth of the zone of deposition beneath any given water table elevation was 50–60 m or possibly less.

At Budapest exploration thus far has discovered only comparatively small fragments of maze caves scattered at different elevations in well dissected hills. With some exceptions, the subaqueous deposits are sparse, being notably smaller in volume than in middle and lower Wind Cave. They display strong vertical zonation. Deposits are thickest at the paleo water line, comprising shelfstone, folia and raft debris. Botryoidal coatings grow for a few metres below, then are succeeded by patches of euhedral spar 0.5–2.0 cm in thickness. There is no precipitation below approximately ten metres.

The intriguing problem is to establish why these differing modes and depths of deposition should occur. I have no pat answers. Below are some suggestions and speculations that may assist future work in these fascinating caves.

(1) *Solute concentrations and saturation state in the waters* may have differed substantially between the three sites. Potentially, such variation could explain all of the difference observed in the deposits. However, considered as the only category of variables, this requires that supersaturation be greatest at Jewel Cave (because deposition extended to the greatest depth there); that is not supported by the evidence for the *rate* of calcite deposition which tends to show that it was probably least at this cave.

(2) *Time available for deposition at a given site* is a second possible type of control. There are two different effects: *first*, is the longevity of a stable water fill (table) with hydrochemical conditions permitting deposition. At Jewel

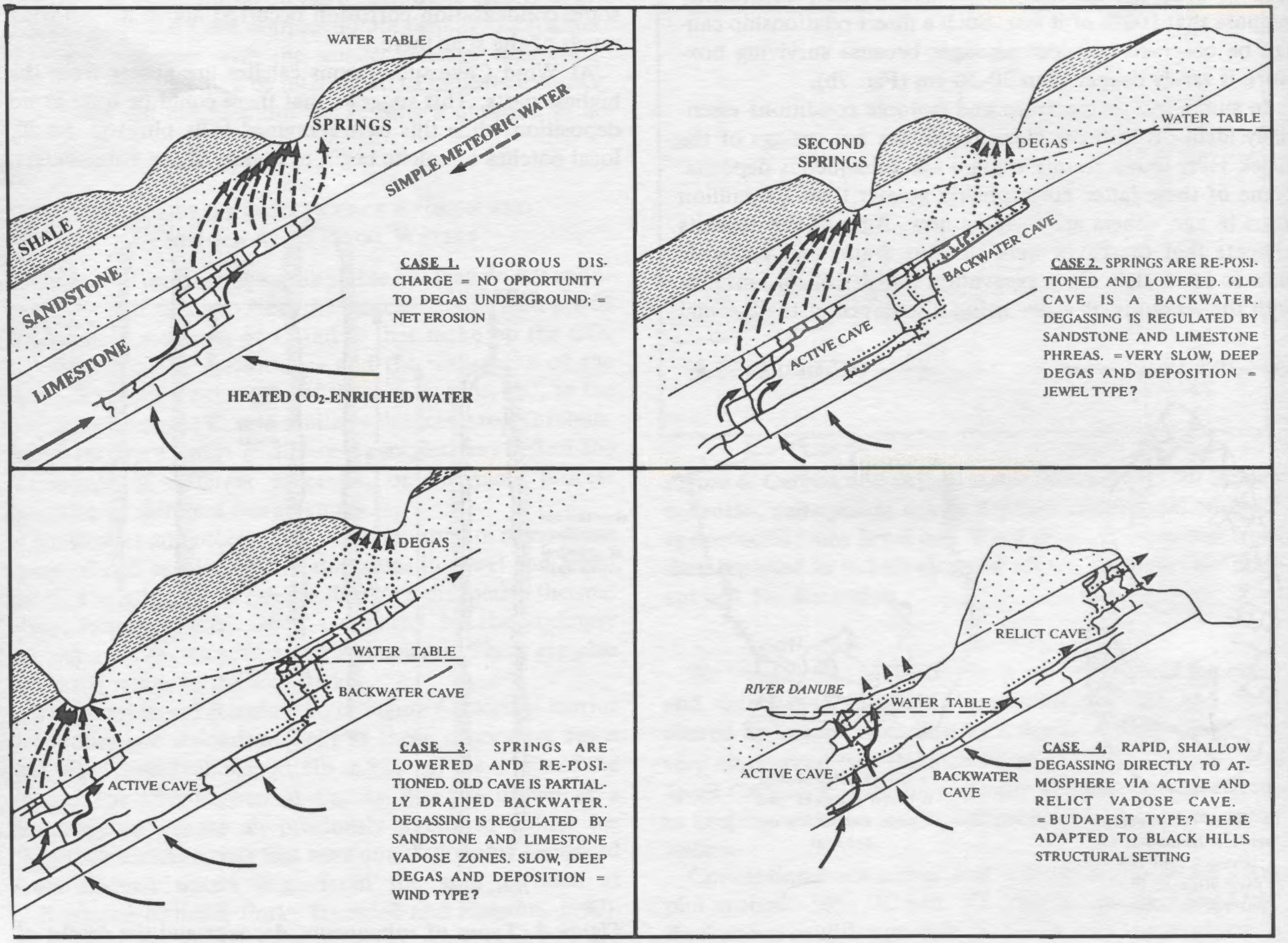

Figure 9. **Four hypothetical models to depict differing relationships between caprock regulation of groundwater flow, the degassing of deep phreatic groundwater, and consequent dissolution or deposition of calcite in the phreatic zone.**

Cave the $^{234}U:^{238}U$ evidence indicates that the entire system was a submerged and passive receptor of spar accumulation for a period not less than one million years. It was then at some unknown depth beneath the water table. At Wind Cave, extrapolation of the $^{230}Th:^{234}U$ dating evidence shows that the water table has descended rather steadily through the cave during the last one half million years. When that period began the entire cave probably was in a net eroding condition; i.e., it was still enlarging and thus no calcite crusts could accumulate. At Budapest the timespan available for subaqueous deposition at a given site was probably even shorter because this is a region of active tectonic uplift that also suffers much greater perturbations from Quaternary glaciation effects than do the Black Hills.

The *second effect* of time is the rate of exchange of phreatic water from which deposition may occur. If a given parcel of water is discharged at the springs before it has had the opportunity to degas there can be no subaqueous precipitation from it. One suspects that as a rule when a cave (even a maze) is functioning as the main discharge channel for thermal waters either there is no deposition or it is largely limited to the waterline. Divers report no deposits below one-two metres in the active outlets of the Budapest caves today. Deeper deposition can begin only where part or all of a system has become a backwater, as in modern Wind Cave.

Longevity of a given water table position and low rate of exchange of water can be mutually reinforcing effects. Jewel Cave may display the greatest and deepest deposits because it was an especially sluggish backwater for a particularly long span of time.

(3) *Caprock regulation of the rate of degassing* is a further possible control that is suggested by comparison of the physical conditions prevailing at the three cave sites during deposition. This is illustrated in Figure 9. When the cave is the main channel for regional groundwaters and all rates of discharge are regulated by phreatic conditions in the caprock, there is no opportunity for significant degassing in the cave zone (Fig. 9, Case 1): this is a wholly erosional situation. If the cave becomes a submerged backwater, degassing is possible but will be very slow because it is regulated by a water-filled caprock (Case 2); this may have occurred at Jewel Cave. An alternative is that when the cave becomes a backwater it also becomes vadose in its upper parts (Case 3—a possible model for Wind Cave). Degassing is more rapid than in Case 2 but is still comparatively slow, being regulated by the less permeable caprock. In Case 4 (the Budapest case) there is unimpeded degassing from the caves directly to the atmosphere.

The models in Figure 9 are counter-intuitive in that depths of subaqueous precipitation are shown as being greatest where rates of degassing are least (and, also, the thermal and pressure gradients). However, they do accord to what is understood of the differing hydrophysical environments of the caves at time of deposition. Other things being equal, rate of exchange of the water would also be least in a Case 2 situation, so that rate of exchange effects may be combining with caprock regulation to achieve the great depth of subaqueous deposition that apparently occurred at Jewel Cave.

Conclusions

Jewel Cave and Wind Cave may be interpreted as multistorey lifting mazes created during the present erosion cycle by waters ascending through them. Mixing corrosion involving lateral (dip- and strike-sourced) groundwaters may have played a role. The ascending groundwaters were heated, probably of the type and magnitude being discharged at Black Hills hot springs today. The subaqueous precipitates were deposited from these heated waters when the caves became backwaters as a consequence of shift of springs. Combinations of several different factors may account for the differing crystal habits, thicknesses and sub-water table depths of calcite precipitation.

The subaqueous precipitates in Jewel Cave are probably older than 2,500,000 years B.P. Wind Cave is much younger. Most of its deposits have formed, and it has drained to expose them, within the past 500,000 years.

References

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

Davis, W. M. 1930. Origin of limestone caves. *Bulletin, Geological Society of America, v. 41;* p. 475–628.

Ford, D. C., Palmer, A. N. and Palmer, M. V. in litt. Uranium series dating of the draining of an aquifer: the example of Wind Cave, South Dakota.

Harmon, R. S., Thompson, P., Ford, D. C. and Schwarcz, H. P. 1975. Uranium Series Dating of Speleothems. *National Speleological Society Bulletin, v. 37(2),* p. 21–33.

Hill, C. A. 1987. Geology of Carlsbad Caverns and other caves of the Guadalupe Mountains. *New Mex. Bureau Mines and Mineral Resources.* Bulletin 117, 149 p.

Hill, C. A. and Forti, P. 1986. "Cave Minerals of the World." *National Speleological Society of America,* 238 p.

Howard, A. D. 1964. Model for Cavern Development under Artesian Groundwater Flow, with special reference to the Black Hills. *National Speleological Society, Bulletin 26,* p. 7–16.

Ivanovich, M. and Harmon, R. S. 1982. *Uranium series disequilibrium: applications to environmental problems,* Oxford University Press. 571 p.

Klimchouk, B. A. and Andrejchouk, V. N. 1986. Geological and hydrogeological conditions of gypsum karst development in the western Ukraine. *Le Grotte d'Italia,* 4(xii), p. 349–358.

Kunath, C. E. 1978. The Caves of McKittrick Hill, Eddy Co., N.M. *Texas Speleo. Survey.* 87 p.

Latham, A. G. 1981. Paleomagnetism, rock magnetism and U-Th, dating of speleothem deposits. Ph.D. thesis, McMaster University, 507 p.

Muller, P. 1989. Hydrothermal paleokarst in Hungary in P. Bosak, D. C. Ford, J. Glazek and M. Horacek (editors). *Paleokarst; a world review.* Academia Praha/Elsevier, The Hague.

Muller, P. and Sarvary, I. 1977. Some aspects of developments in Hungarian speleology theories during the last ten years. *Karszt-es Barlang*, p. 53–59.

O'Neil, J. R., Clayton, R. N., and Mayeda, T. 1969. Oxygen isotope fractionation in divalent metal carbonates. *Journal of Chemical Physics, v. 51*, p. 5547–5558.

Szunyogh, G. 1984. Theoretical investigation of the origin of spherical caverns of thermal origin (Second Approach). *Karszt-es Barlang*, p. 19–24.

Truesdell, A. H. and Hulston, J. R. 1980. Isotopic evidence on environments of geothermal systems. in P. Fritz and J.-C. Fontes (Editors) *Handbook of Environmental Isotope Geochemistry*. Elsevier, The Hague. p. 179–226.

Tullis, E. L., and Gries, J. P. 1938. *Black Hills caves*. Black Hills Engineer, v. 24, p. 233–271.

Miller, Thomas E.—Evidence of Quaternary Tectonic Activity, and for Regional Aquifer Flow at Wind Cave, South Dakota: *NSS Bulletin* 51: 111-119.

EVIDENCE OF QUATERNARY TECTONIC ACTIVITY, AND FOR REGIONAL AQUIFER FLOW AT WIND CAVE, SOUTH DAKOTA

THOMAS E. MILLER

Radiometric dating of extensive sub-aqueous precipitates from Wind Cave, South Dakota, shows deposition in the period from 350,000-1,250,000 years B.P., and establishes a minimum age of 350,000 years for the cave. Tilting and faulting of these precipitates demonstrate tectonic activity in the Black Hills at least since the mid-Pleistocene. A local dike intrusion beneath the cave is indicated by geophysical evidence, and could possibly relate the present high geothermal gradient in the cave, the dated precipitation event, and the recent tectonic activity.

In the lower cave levels, the well-known "boxwork" of Wind Cave is covered by the calcite precipitates, and therefore predates them.

The first hydrochemical survey of surface and cave waters undertaken in Wind Cave National Park (1976–1978) shows the chemical similarity of cave waters to nearby artesian well waters, and supports the idea that they are part of a regional flow system, rather than merely local. Seasonal solutional enlargement also appears to be possible well into the aquifer.

INTRODUCTION

Wind Cave is among the largest cavern systems in the world, with in excess of 80 km of mapped passage. It has been cited as a classic example of "artesian" flow development (e.g. Palmer, 1975; and Ford and Ewers, 1978), largely because it contains extensive three-dimensional mazes. Only recently have studies attempted to test this hypothesis, or to determine the age of the cave. Located in the Black Hills of South Dakota (Fig. 1), the cave offers numerous advantages for such testing: it is relatively accessible, well mapped, and contains large quantities of dateable precipitates. It is also a rare window into the carbonate aquifer fronting the Black Hills.

Geology

The cave has formed in the Mississippian Pahasapa Limestone, which is the South Dakotan equivalent of the Madison Formation of Montana and Wyoming. The Pahasapa Limestone is about 90 m thick in the Wind Cave area, as much as 180 m thick elsewhere in the Black Hills (Rahn and Gries, 1973), and in the cave area dips four to seven degrees to the southeast (White and Deike, 1962; Howard, 1964). The Pahasapa is underlain by non-cavernous Paleozoic sedimentary rocks and Precambrian basement rock. It is unconformably overlain in the cave area by the Pennsylvanian Minnelusa Sandstone, and younger shales, limestones, and sandstones. Following exposure of the limestone, a mature karst surface developed in many of the Western states (Sando, 1974). Dolines and caves that formed in this period were largely filled by rock and finer debris from the Pennsylvanian Absaroka transgression. These two events are clearly exposed in the bedrock hosting the present cavern, but the pre-existing caves

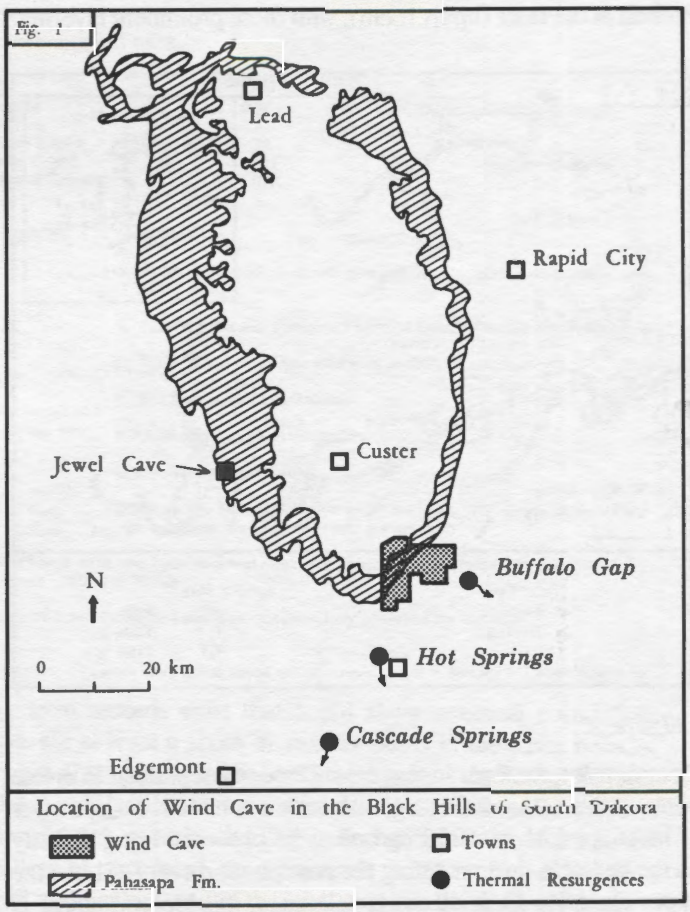

Figure 1.

of the Mississippian era have had an unknown influence in the location of the present passages.

Cave Morphology

The presently known internal relief of Wind Cave is about 215 m (1079-1294 m elevation)—in profile view, it dips southeast accordant with the bedding. The cave developed almost entirely in phreatic conditions, with concentrations of known passage occurring in "upper levels" generally above 1160 m in the northwest area of the cave, and in "lower levels" below about 1150 m in the southeast area.

Much of the lowest part of the cave (e.g. the lower half of Fig. 2) is covered with a continuous precipitated crystalline calcite to a thickness of 2.5-5.0 cm on the walls. This coating is sub-aqueous in origin, and resembles that of nearby Jewel Cave (where it can be 15 cm thick). Internal layering can be found at many (but not all sites): this is also similar to Jewel Cave, and shows that more than one episode of deposition occurred. This large precipitation event (or group of episodes) was limited to the lower area, although smaller events of unknown significance have occurred elsewhere in the cave (e.g. the section known as the Blue Grotto).

The Wind Cave wall coating has a discernable upper limit or boundary below which it increases in thickness with depth. It is thickest at the floor (up to 10cm), with more prominent layering.

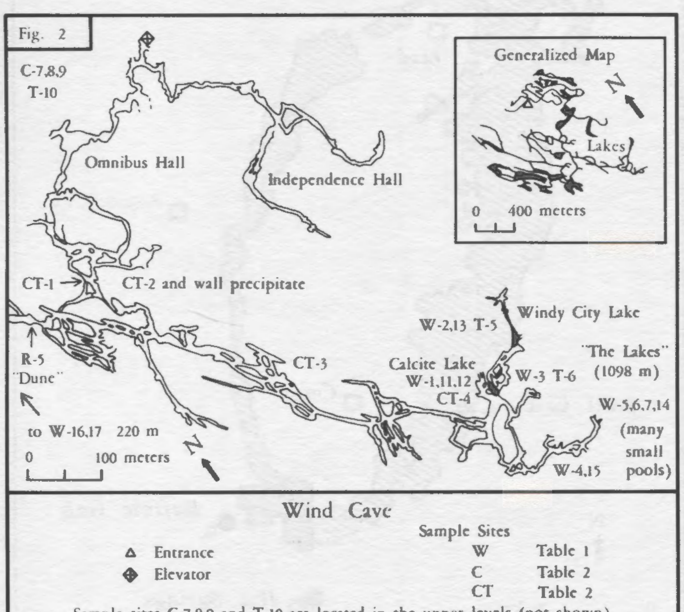

Figure 2.

Occasionally, the floor layers are sufficiently separated to create hollow "false floors" (Plate 1). Where these have broken, a layer of loose, pinkish material can often be observed lying directly on the bedrock, thus predating the precipitation event. Above the upper elevation limit of the precipitates, this loose material is frequently almost dunelike in *appearance*: The particles are

***Plate 1.* Collapsed false floors in lower levels of Wind Cave, with typical calcite precipitate coating above.**

powdery or silty in size, with horizontal "waterlines" present on the outside surface of some of the "dunes". Maximum thickness exceeds 50 cm for some dunes. Jennings (1985) described similar deposits in an Australian cave, interpreting them as in-situ weathering accumulations from salt crystal growth in the cave walls.

Wind Cave is perhaps best known for its "boxwork" features, described as thin, intersecting "blades or fins of crystalline calcite that project into [passages] from the walls, ceiling, or floor" (Palmer, p. 33, 1981). In the lower cave, they also are covered by the later precipitation event.

A final phenomenon is the occurrence in Wind Cave of aerosol solution of bedrock, with associated mineral precipitation. In some passage constrictions, freshly-weathered bedrock is exposed, while in larger adjacent cross-sections, carbonate and/or sulphate minerals have precipitated in apparent "leeward" or "eddy" regions. Such effects are rare elsewhere (e.g. Caverns of Sonora, Texas, Carlsbad Caverns, New Mexico, and Belize, Central America), and always found in fossil passages.

Faults, of generally minor vertical movement (<50 cm), are common features in the cave: they frequently intersect, and offset, conduit passages and small phreatic conduit tubes, indicating post-solutional movement.

Hydrology

The cave locale is arid, with mean annual precipitation of about 43 cm, and recharge of about 2.5 cm. Runoff from the central Precambrian and Paleozoic core of the Black Hills flows radially outward across the Pahasapa limestone. Streams entering Wind Cave National Park (all are east-flowing) are pirated underground within the Park boundaries.

Hydrothermal discharge occurs south and east of the Park in three groups of springs (Fig. 1)—Buffalo Gap, Cascade, and the combined springs creating the Fall River at Hot Springs (e.g. Hot Brook and Evan's Plunge). Temperatures of 17-34° C have been measured (Rahn and Gries, op cit.). The Buffalo Gap springs have long been postulated as the resurgence for the Park's sinking streams because they are both downdip and topographically lower. All of these springs lie above anticlinal flexures which raise the Pahasapa closer to the surface.

Within the cave, only one tiny (<1 l sec^{-1}), permanently-flowing stream (Rebel River) is known (it is southwest of CT-3 on Fig. 2). As previously noted, the cave is overlain by sandstone. It is of very low permeability: except in a few near-surface sites, there are no known seeps into the cave, and few drips. This is highlighted by the extreme scarcity of calcite speleothems, active or fossil, throughout the cave.

Most of the water in the cave occurs as numerous pools found at the extreme downdip end (below 1100 m) of the cave, in the "Lakes" area (Fig. 2). The largest (Windy City Lake, Plate 2) is 55 m long and 9 m deep. No visible flow enters it, yet the author has noted fluctuation of as much as 30 cm over a period of several months, and has been told of past change of as much as a meter. This has been attributed to hidden conduits or seeps; no published

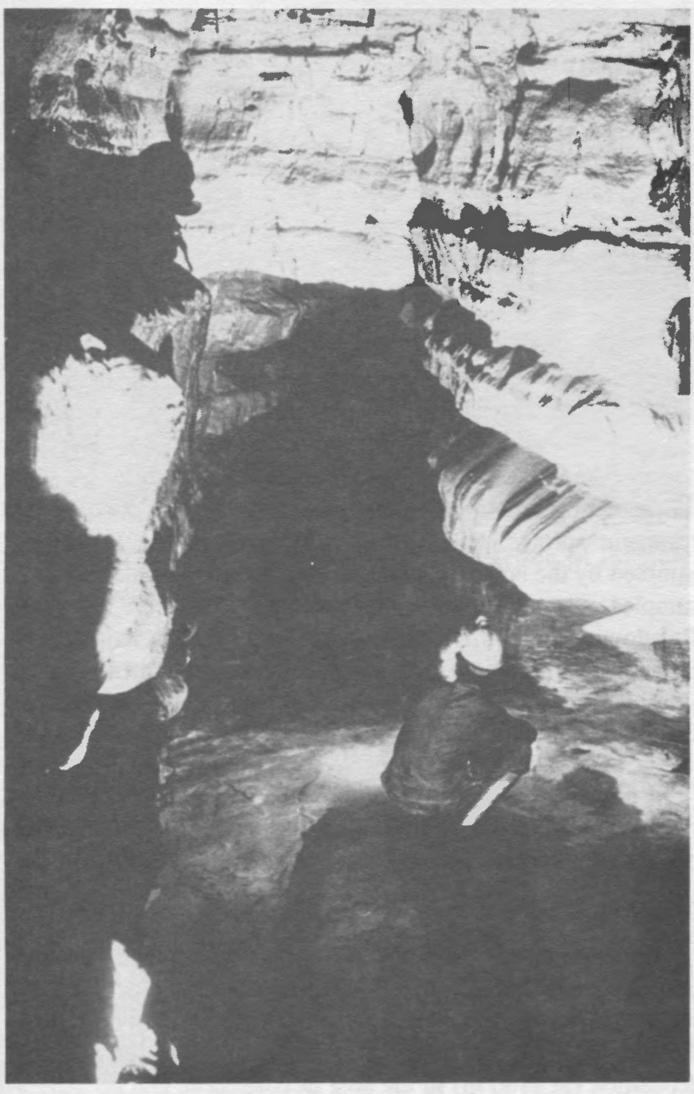

Plate 2. **Windy City Lake, looking northeast.**

TABLE 1 HYDROCHEMICAL AND HYDROLOGIC DATA

1 A. WATER ANALYSES

Location	ID #	Date	Temp.	pH	Ca^{2+}	Mg^{2+}	HCO^{3-}	SIc	SId	P_{CO_2}
Wind Cave Lakes										
Calcite Lake	W-1	1974	13.7	7.80	108	78	180	0.13	0.09	0.30
Calcite Lake	W-11	1976	13.7	7.85	90	84	176	.10	.11	.26
Windy City Lake	W-2		14.0	8.20	79	74	148	.33	.34	.10
	W-3		13.8	8.05	89	86	174	.29	31	.16
	W-4		13.8	8.30	73	84	151	.40	.46	.08
	W-5		13.7	8.35	79	89	164	.51	.57	.08
	W-6		13.7	8.10	75	87	151	.21	.27	.13
	W-7		13.7	8.15	83	87	173	.35	.38	.13
Calcite Lake	W-12	1978	13.7	7.8	87	68	155	-.01	-.04	.26
Windy City Lake	W-13		14.0	7.6	109	76	171	-.08	-.13	.46
	W-14		13.7	7.65	99	74	162	-.10	-.13	.38
	W-15		13.8	7.8	72	83	152	-.10	-.13	.26
Crystal Lake	W-16		11.4	7.75	84	81	156	-.11	-.11	.29
Petey's Lake	W-17		11.4	7.80	86	87	162	-.04	-.02	.27
Surface Streams										
Cold Spring	W-8	1976	2.2	7.82	187	116	273	.33	.21	.38
Beaver Creek	W-9		0.2	7.70	160	68	199	.04	-.20	.36
Highland Creek	W-10		5.4	7.95	98	33	126	-.01	-.27	.14
"Madison" Flowing Well (Streeter Well)										
(Knirsch, 1980, p. 28)		—	—	7.6	110	52	—	—	—	—
(Bakalowicz et al., 1987)		1982	19.0	7.56	109	48	163	-.12	-.49	.56
			SIg = -3.19							
Thermal Resurgences										
Buffalo Gap	W-18	1976	10.7	7.95	1300					
Cascade Spring			19.4	7.0	1420	400	193	0.18	-.11	1.91
Evan's Plunge			30.6	—	738	213				
		SIg = -0.04 (Cascade Spring)								

Other Ions (mg/1)	Na^+	SO_4^{2-}
Cascade Spring	130	1604
Evan's Plunge	177	767
"Madison" Well (Knirsch, 1980)	8	13
(Same) (Bakalwicz, et al., 1987)	—	3

1 B. HYDROLOGIC DATA

Site	Q(Mean)	S.D.	C.V.	
Cascade	0.67	0.06	0.09	Thermal Springs
Fall River	.71	.11	.15	
Buffalo Gap	.24	.60	.27	
Beaver Creek	.017			Surface Streams
Highland Creek	.020			

1 C. NOTES

a. All solute concentrations given as mg/L $CaCO_3$ except Na+ and SO_4^{2-}

b. Fall River is the combined flow of Evan's Plunge and Hot Brook

c. P_{CO_2} is the % partial pressure of CO_2 in solution

d. S.D. . = standard deviation

e. C.V. = coefficient of variation

f. Q = discharge in m^3/sec

g. SIm = the saturation index of mineral "m" (see text), where c = calcite, d = dolomite, and g = gypsum

All discharge data, and hydrothermal chemistry (except for Buffalo Gap), are from Rahn and Gries (1973, pp. 29-32)

Sample W-1 from Calcite Lake was commercially analyzed by the Park.

long term records exist that could show seasonal correlation. There are at least a score of smaller pools in the cave: because they are also located in the lowermost unit of the limestone (and in the lowest levels of known cave passage), do not have meteoric sources (such as stalactite drips), and generally conform to the position of the local potentiometric surface of the Madison (1108 m elevation west of the Park, 1112 m immediately south—both figures from unpublished U.S. Geological Survey data, Rapid City South Dakota office—and 1087 m at the cave lakes) they are likely to represent the water-table surface in the cave.

PURPOSE OF THE STUDY

Although the waters that formed Wind Cave drained long ago, it is possible to attempt some understanding of their role in its development. This can be approached through study of the present hydrochemistry of the cave, and examination of the extensive precipitates that coat its lower part. Dating of these precipitates can also provide a time framework for the development of the caves.

The Park first granted a permit to the author in 1976. The initial objectives of study were to:

1) collect, analyze and compare surface and subsurface waters in the Park for the first time, to relate them in terms of origin;

2) measure the temperature gradient and atmospheric CO_2 pressures in the cave for the first time; and

3) collect a sample of the extensive lower-level precipitates for uranium-series (^{230}Th / ^{234}U) dating to provide the first estimate of the age of the cave.

Collection of bedrock and "dune" material was also done at that time, and additional water analysis was made in 1978. A preliminary report of the hydrochemistry and age findings was made to the Park in 1978, with a final submission placed on file in the park in 1979 (Miller, 1979), and copies distributed to other researchers. During a subsequent visit in 1982, the author measured the elevations of the top of the precipitate layer to determine if it was horizontal. Preliminary publication of all these results (including the rock and dune analyses) was in 1988 (Miller, 1986-87).

SAMPLING PROCEDURES

Water Sampling

Collection of water in the park was conducted in two phases; the first was in December, 1976, and the second in July, 1978. Each sample was collected in a 500ml Nalgene bottle that was filled beneath the water surface and tightly capped to prevent degassing CO_2. Temperature and pH were recorded on-site, with the exception of some remote cave sites where difficulties in transportation of the pH meter prevented immediate measurement. These short delays (2-3 hours) have been shown by other researchers to be of little significance, especially where the temperature of the sample is kept constant (as it was during transport within the cave).

Bicarbonate alkalinity was determined immediately at the cave by titration with HCl. Following acidification, titration with EDTA was made after transport to the lab. Each entry in Table 1 is the mean of at least two readings that differ by less than 3 mg l^{-1} (as $CaCO_3$).

Specific conductivities of the water samples were taken to ensure that the analysis was complete for the significant ions.

These data are listed in Table 1; relevant samples from other studies were included if they were sites not visited by the author or for which chemical equilibria had not been calculated. The three thermal spring groups in the area (from Rahn and Gries, op cit.), and two nearby artesian wells that penetrate the Pahasapa Limestone (Knirsch, 1980) are also shown on Figures 1 and 3 (with the exception of the Kaiser Well which is immediately

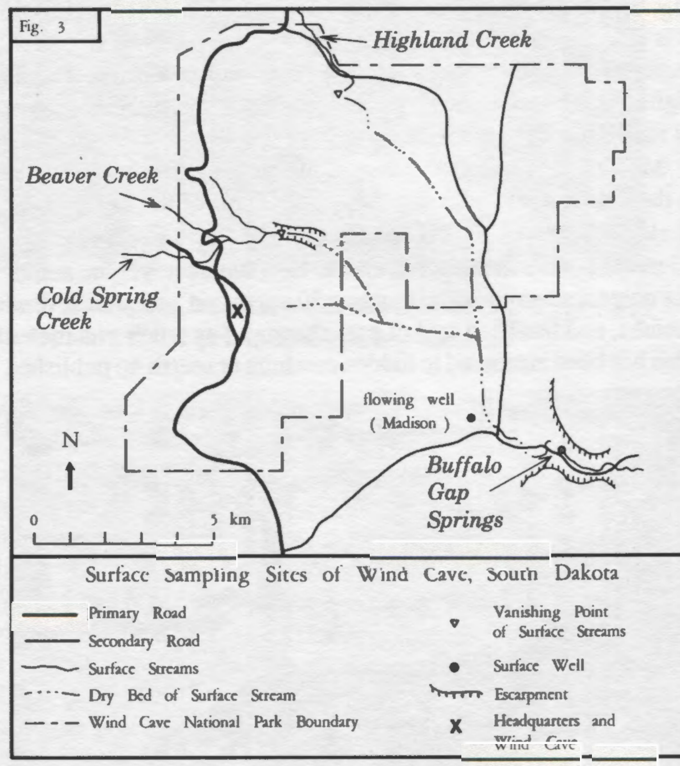

Figure 3.

south of the southwestern border of the Park), and Figure 2 shows the sampling locations within the cave. Of the three major resurgent springs, only Buffalo Gap (east of the Park) was sampled by the author. The more complete of the Madison Well samples was collected by the author, D.C. Ford, and M. Bakalowicz in 1982 (Bakalowicz, Ford, Miller, Palmer and Palmer, 1987). Sample W-1 was collected for commercial analysis by the Park in 1974: its temperature was not measured at that time, and is assumed to be the same as that measured at the site in 1976 and 1978. Samples W-3 to 7, and W-13 and 14 were from unnamed pools in the vicinity of Calcite and Windy City Lake. Crystal River and Petey's Lake (W-16, W-17) were small pools a kilometer northwest of these large lakes and 100 m higher.

Precipitates

A section of the extensive lower-level precipitates was removed (in 1976) to the lab at McMaster University for uranium-series dating. The sample was a 1.2 cm-thick uniform coating of the cave walls. It was near the upper boundary of sub-aqueous deposition (at 1160 m) in the area of collection, and because it had no internal banding, was from the oldest precipitation episode. Its surface was not weathered or re-dissolved, limiting the likelihood of post-depositional alteration.

Elevation changes in the upper boundary of the precipitates could be measured because it rapidly decreased in thickness near its upper boundary, and terminated within a one-meter interval. These elevations were noted over a 570-meter linear transect in 1982 (Plate 3), and were calculated from the vertical difference between the precipitate boundary and permanent marked known survey station nearby.

Plate 3. **The "Garden Gate," site of collection of the dated precipitate sample.**

Atmospheric Temperature and CO_2

Eight different determinations for carbon dioxide content in the cave air were made with a Draeger hand pump and ampules. Air temperatures were measured by a mercury thermometer (graduated to 0.1° C) from the entrance (1245 m) to the lowest part of the cave (1079 m elevation in the Lakes area) (Table 2).

TABLE 2 ATMOSPHERIC CO_2 AND TEMPERATURE

Site	ID #	Elev	%CO_2	Temperature (°C)	
Top of Boxwork Chimney	CT-1	1173	0.031	11.3	Upper Part of Cave
Devil's Lookout	C-7	1194	.045	10.0	(Park Service)
Coliseum	C-8	1219	.024	10.0	
2nd Crossroads	C-9	1189	.039	10.0	
Garden Gate	CT-2	1157	.045	11.5	Lower Part of Cave
Station JF-80	CT-3	1128	.038	12.1	
Calcite Lake	CT-4	1097	.041	13.6	
Windy City Lake	T-5	1097		14.3	
Transition Lake	T-6				
Surface (2 readings)			.018		

Note: Elevations are in meters

RESULTS AND DISCUSSION

Dating of Cave Events

The sample of calcite precipitate provided some absolute figures for preliminary estimation of the age of the cave and more recent events. Uranium-series dating (^{230}Th / ^{234}U) was done at McMaster University (Schwarcz and Gascoyne, 1984), with the results shown in Table 3. Because the ^{230}Th / ^{234}U ratio is greater than unity (1.099), Sample 76020-1 was deposited at least 350,000 years before present, the limit of the method at the time of analysis in 1977. It provides a *minimum* age estimate of the oldest precipitation event that coated the lower part of the cave.

The decay series of ^{234}U / ^{238}U could provide age estimates beyond 350,000 years, but use of this method requires knowledge of their initial ratios. These ratios are quite variable for freshwaters, and cannot be assumed to be the same as those of the modern waters at the cave. However, because this ratio achieves secular equilibrium (ratio= 1.0) within 1.25 million years, a ratio greater than 1.0 (showing disequilibrium) demonstrates the crystallization of a calcite at a more recent age. The ratio of the Wind Cave sample is quite high (1.57), and shows the event took place in the interval 350,000-1,250,000 years B.P.—since the mid-Pleistocene.

Plate 4. **The large fault in the lower levels, as it appears on the northeast side of the passage. This limestone unit is abruptly lower to the right, and it is thus a normal fault, in spite of its local superficial appearance.**

The upper surface of the calcite is not horizontal: in one measured section of the lower cave this upper boundary drops 27 m in a horizontal distance of about 570 m, a 2.7° overall slope (shown diagrammatically in Fig. 4b). The slope is increased midway (from 1 to 3 degrees) by the presence of a large normal fault (Plate 4): the fault has shattered the calcite coating for many

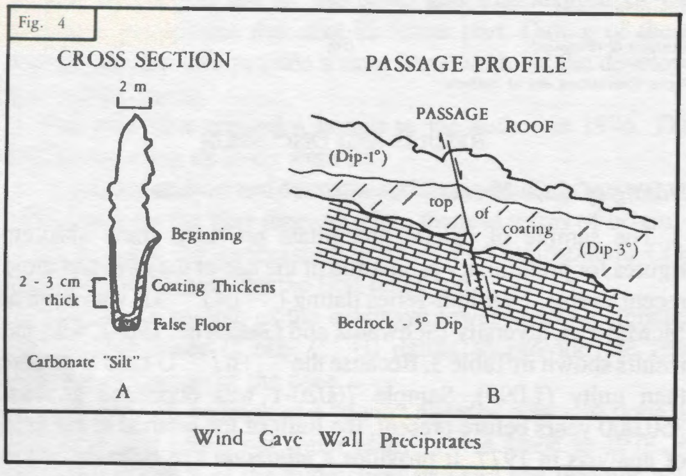

Figure 4.

meters on either side. Both a major local tilting of the cave area, and fault activity, are indicated within Pleistocene times, after the latest precipitation event.

Tectonic activity in the area is poorly described: modern seismic activity in the region shows a minor concentration in the Black Hills (Reagan, 1981), and state highway surveys suggest neo-tectonic uplift in the area immediately to the north of the Black Hills in the Belle Fourche area (B. Gomez, Department of Geography, Indiana State University, 1989, personal communication). James Martin (Museum of Geology, South Dakota School of Mines and Technology, 1990, personal communication) reports a preliminary age of three million years for the oldest of three major river terraces incised following uplift in the eastern Black Hills (based on fossil identification), and as recently as 20,000 years B.P. for the youngest.

Hydrochemistry

The method outlined by Langmuir (1971) was used in the author's computer program MILLKEM to calculate the degree of saturation with respect to calcite and dolomite (SIc and SId), and the partial pressure (P_{CO_2}) of the carbon dioxide in solution:

$$SIm = \log_{10} (IAP/Km) \qquad (1)$$

Where SIm = the saturation index of a given mineral "m" (e.g. an SIc of 0.0 indicates calcite saturation)

IAP = the ion activity product of those ions in solution

Km = the equilibrium constant for "m" at a given temperature

In Table 1, SIc and SId are logarithmic expressions of the measured to predicted activity ratios of the calcite and dolomite ions in solution. Positive values show supersaturation of the cave water with respect to the specific mineral; negative values show undersaturation. Calculations for some of the gypsum solubilities (SIg) are also shown.

These indices were also calculated for samples reported elsewhere in the area during other studies: the large thermal resurgences of the area, and two capped artesian wells ("Madison" Flowing Well and Kaiser Well) penetrating the Pahasapa Limestone between the cave and the Buffalo Gap Springs (Table 1A).

Table 3. Uranium Series Dating of the Calcite Precipitate
Location: Survey Station JF-45 ("Garden Gate"), Elevation 1160 meters
Analysis Run: 11-22-77
Sample ID: # 76020-1
U-concentration: 3.57 ppm

Isotope	Ratio	Error
$^{234}U / ^{238}U$	1.568	0.0328
$^{230}Th / ^{234}U$	1.099	0.0935
$^{230}Th / ^{232}Th$	33.2	11.83

All of the Park waters sampled were at or near saturation for both calcite and dolomite, especially the cave lake waters. Some of the latter were supersaturated, a finding supported by the presence of precipitated calcite "ice-rafts", floating due to surface-tension.

These results are somewhat surprising for the surface waters. Two of the sinking streams of the Wind Cave area are not only higher in calcium content and P_{CO_2} than the cave waters, but all are already at saturation with respect to calcite: for Beaver and Highland Creeks (collected during stable winter low flow to minimize the effects of greater variability expected for surface waters) to equilibrate with the water in the cave lakes and pools, they would not only have to lose CO_2, and precipitate calcite, but simultaneously gain magnesium through incongruent solution of dolomite (Wigley, 1973). This process is unlikely, and would ultimately seal the aquifer to further flow.

Rahn and Gries (1973) surveyed discharges of springs throughout the Black Hills, and discussion of their findings has relevance to the interpretation of the Wind Cave data. The total flow of the sinking streams in the Park is insufficent to account for more than a small fraction of the water resurging at Buffalo Gap: Rahn and Gries (op cit., p. 29) have estimated the losses of sinking streams in the park to average about 37 liters per second (1.3 cfs): Table 1B. Yet Buffalo Gap alone, smallest of the major thermal springs in the area, has a mean discharge of 240 l sec^{-1}. If Rahn and Gries' estimate of mean areal discharge for the Cascade spring of 0.79 l sec^{-1} km^{-2} (0.07 cfs $mile^{-2}$) is similar to that for Buffalo Gap, then an additional limestone recharge area of 250 km^2 is needed to supply the remaining 203 l sec^{-1} for Buffalo Gap. This would extend far outside the park boundaries.

Rahn and Gries have postulated a substantial regional flow passing through the aquifer of the Pahasapa, and resurging at the several hydrothermal springs of Figure 1—such a flow would be a more logical source for the cave lake waters than the surface streams.

Comparison of the Park (surface and cave) waters with nearby artesian wells (Table 1A) does not lead to unequivocal categorization, however. The artesian well waters are most like the cave waters in terms of total ion concentrations (especially calcium, <100 mgl^{-1}), but the calcium-magnesium ion ratios, and saturation indices (the wells are undersaturated with respect to calcite and dolomite), have similarities to both Park water types. It might not be unreasonable to assume that the well waters are the resulting mixture of regional flow (as represented by the cave waters) with local inputs from the sinking surface streams.

Although there is remarkable similarity in the values of all the variables measured in the cave lakes, the samples collected in winter, 1976, were all supersaturated, while those of summer, 1978, were slightly undersaturated. This difference is greater than the error range of the instruments, but additional samplings are needed to determine whether it represents a significant seasonal effect. The P_{CO_2} in the wells is surprising: it is considerably higher (>0.5 %) than either of the Park waters, almost as though there were a CO_2 source within the aquifer contributing to the undersaturation and aggressiveness of the artesian waters with respect to the carbonate rocks. If the seasonal undersaturation of the cave lakewaters is also confirmed, positive evidence for modern, widespread solutional enlargement deep in the aquifer would be confirmed, with implications for the past morphological development of the cave under artesian conditions.

Buffalo Gap is unfortunately not permanently gauged: the only major spring group in the area that is gauged is that of Cascade Springs. Because both share similarities of low flow variability (Table 1B), and a thermal, Pahasapa Limestone source, some of Buffalo Gap's internal organization may be inferred from analysis of Cascade discharge data.

Discharges for Cascade spring reported by the USGS (Hoffman, 1982) show an extremely small range for the water years of 1977-1982. In these six years, mean daily discharge varied only from a minimum of [510 1 sec^{-1}] to a maximum of [708 1 sec^{-1}]. Ten baseflow recessions, none with a range greater than [170 1 sec^{-1}], were noted in this period, and individual recession slopes (k) were calculated using

$$Q = Q_o e^{-kt} \qquad (2)$$

where Q= instantaneous discharge at the start of the recession

Q_t= discharge at some "t" after the start of the recession

e= base of natural logarithms

k= recession constant (slope of the recession) for the basin

Recession slopes ranged from [-1.0×10^{-5} to -3.1×10^{-5} sec^{-1}], and integrating (2) to calculate total storage:

$$\text{Storage} = Q_o / k \qquad (3)$$

gave estimates of [23 – 108 million m^3], or from 1-5 years of annual discharge (Table 4). The Buffalo Gap spring has a coefficient of variation of flow (Table 1B) that is greater than Cascade Springs (0.27 versus 0.09), but the implication is that the low range of Buffalo Gap flow (157 to 446 1 sec^{-1}) still indicates a storage sufficiently large to be of regional extent, and a residence time long enough to make difficult a positive dye-tracing of a connection from Wind Cave to Madison Well to Buffalo Gap.

Table 1 demonstrates the existence of two primary chemical water types in the park vicinity: 1) the sinking inflowing streams, the cave waters, the artesian wells—and 2) the thermal springs. A partial analysis of the thermal resurgence at Buffalo Gap

Table 4. Recession Discharges of Cascade Springs

Location: Latitude 43 20'10"; Longitude 103 33' 07"
Elevation: 1049 meters a.s.l.

Mean Discharge 567 1 sec^{-1}
Mean Annual Flow approx. 19.8 X 10^6 m^3

	Individual Recessions					
	Start	Q (1 sec-1)	End	Q	Slope (k)*	Storage**
1.	10-76	567	1-77	510	-1.04	54
2.	7-77	567	9-77	510	-1.19	47
3.	12-77	709	5-78	538	-3.1	23
4.	6-78	652	10-78	510	-2.1	28
5.	7-79	680	10-79	567	-2.78	24
6.	12-79	595	3-80	538	-0.95	62
7.	6-80	623	9-80	510	-2.1	30
8.	12-80	538	4-81	510	-0.5	108
9.	9-81	567	12-81	510	-1.94	29
10.	7-83	709	9-82	567	-2.2	30

Note: Recession 8 was of *monthly* means.
*Slopes x $10^{-5} s^{-1}$
**Storage x 10^6 meters^3

(W-18) is included in Table 1 with two analyses of Cascade Spring and Evan's Plunge (both from Rahn and Gries (1973). All of these springs showed high values of dissolved anhydrite or gypsum, and are difficult to relate directly to the purely bicarbonate waters of the cave lakes. The chemical similarities of the well water to the cave waters suggests broad areal movement through a Pahasapan aquifer to the vicinity of resurgence, then evolution to sulphate-type water by solution of gypsum and anhydrite in the overlying Minnelusa and redbed formations.

Atmospheric Temperature and CO_2

The surface CO_2 values were lower than expected (at 0.018 % they were half the global mean), but may be explained by local influences of altitude and cold weather. Significantly higher percentages were found in the cave than outside (0.024- 0.045%), which is normally attributed to loss of CO_2 to the cave atmosphere from seepage water charged to soil atmosphere levels (Miotke, 1974). The atmospheric carbon dioxide levels increased with depth in the cave: the mean was 0.035% in the upper levels, versus 0.041% in the lower sections. All are low compared to global atmospheric values, and are as much as an order of magnitude below the P_{CO_2} of the cave pools and lakes, and the artesian well waters.

Wind Cave was named for the impressive atmospheric exchanges that occur through its main entrance due to barometric pressure differences. The author has measured windspeeds in excess of 65 km hr^{-1} in the entrance, and airflow may persist with the same direction for several days. Air currents can be detected thousands of meters into the cave.

Considered together, the previously-noted lack of dripwaters, increase of atmospheric CO_2 with depth, and the considerable air movement throughout the cave, may indicate that CO_2 and other gases diffuse from the aquifer into the cave, to be ultimately discharged by the cave winds.

Hydrothermal Activity

The major calcite precipitation events that occurred at Wind and Jewel caves are among the attributes that make these caves truly remarkable. White and Deike (1962) inferred high-temperature sub-aqueous deposition at Wind Cave, and Deal (1964) suggested the events at Jewel Cave were caused by a temperature rise that made the carbonate in solution less soluble. Analysis of wall precipitates collected by the author and others from Wind Cave has found evidence of deposition under hydrothermal conditions (Bakalowicz et al., op cit.). This is not unexpected in light of the neighboring warm springs and high geothermal gradients previously noted in the area; ironically, theorists as long ago as the 1880's suggested "geyser" origins for the cave (e.g. Owens, 1898).

Rahn and Gries (op cit.) report a geothermal gradient in the Hot Springs area of 7.2 C per 100 m. The gradient measured in Wind Cave (3.3 C in 122 m or 2.7 per 100 m—Table 2) is less than half this, but still indicative of a thermal anomaly. Rahn and Gries suggested several theories for the high gradients, among them a shallow magma body, radioactive decay of elements at depth, locally high rock conductivity, and the exothermic reaction of the alteration of anhydrite to gypsum: $CaSO_4^{-2} + 2H_2O = CaSO_4^{-2}\ 2H_2O$.

Knirsch's (op cit.) chemical and temperature study of 29 wells penetrating the Pahasapa limestone in the southern Black Hills dismissed all but thermally-conductive Precambrian granite as a likely heat source for thermal gradients near Edgmont 45 km southwest of Wind Cave. She found that the thermal gradients in the Kaiser and Madison Flowing wells to be nearly identical to that measured in Wind Cave by this study.

Gravity and aero-magnetic data in the Black Hills were summarized by Kleinkopf and Redden (1975). Their aero-magnetic map shows an extremely anomalous, positive ridge running directly beneath the cave, and paralleling it in orientation. It correlates with a less developed, but still noticeable, gravity ridge located in the same area. Elsewhere in the Black Hills, correspondence of high gravity readings and high magnetic readings usually indicates an intrusion, often laccolithic (Kleinkopf and Redden, p. 68). The linearity of this feature suggests it is a vertical dike intrusion.

It is tempting to infer that the precipitation event is due to the emplacement of a dike, but the dating of the event at less than 1.25 million years would pose a problem: the most recent intrusive events in the Black Hills have been dated at Eocene in age (30 m.y. B.P.), and well over 100 km distant in the northern part of the range. The hot springs may themselves be relatively recent phenomena, indicating a rebirth of igneous activity in the area.

SUMMARY

The widespread calcite precipitation events in the lower levels of Wind Cave allowed initial time frames to be established for various events in the cave. The cave itself is at least 350,000 years old, as determined from dating the oldest precipitation event to the interval 0.35 to 1.25 million years B.P. Boxwork and in-situ weathering accumulations in the lower cave predate the precipitation and may be older than 1.25 million years. Faulting and tilting of the precipitates clearly show tectonic activity since at least the mid-Pleistocene.

A survey of water chemistry in Wind Cave's lower levels. and of sinking surface streams in the Park, was able to establish that the surface streams were an unlikely source for the cave waters. Two artesian wells in the vicinity have sufficient similarities to both to suggest mixing of the three waters in the subsurface. Calculation of calcite and dolomite saturation indices show the Park waters are at or near saturation; the well waters are surprisingly undersaturated, with elevated P_{CO_2}, and are capable of solution deep under artesian conditions. The limited hydrologic and hydrochemical evidence from the major thermal springs in the area are at present best explained by the existence of large-storage, regional groundflow through the limestone aquifer to a few resurgent points.

The transect of air and water temperatures from higher to lower levels in the cave showed a geothermal gradient of 2.7° C per 100 meters, comparable to two well waters in the area. Aeromagnetic and gravity data from the Black Hills implies the presence of a dike beneath the cave—this may have relevance to the thermal gradient and the precipitation event.

Future studies in Wind Cave need to expand the sampling of both the precipitates and the lake and pool waters in the lower cave: the only known stream in the cave, and numerous pools, remain to be analyzed. Determination of the upper boundary surface of the precipitates throughout the lower cave could establish the general dip and direction of uplift. Additional dating elsewhere in the lower levels would clarify chronologic relationships and rates of events.

Permanent recording of stage levels and chemistry in both Windy City Lake within the cave, and of Buffalo Gap springs, are needed to better determine the extent, storage, and flow relations of the apparent regional aquifer. Dye-tracing from the Park lakes to the artesian wells and thermal springs is desirable, but may require too lengthy a time. The more variable chemical response of the surface streams in the Park are also poorly characterized by the few present samples.

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REFERENCES

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

Deal, D.E. (1964) Origin and secondary mineralization of caves in the Black Hills of South Dakota, U.S.A. *4th International Congress of Speleology Proceedings*, 1965 p. 67-70

Ford, D.C. and Ewers, R. (1978) The development of limestone cave systems in the dimensions of length and depth *Canadian Journal of Earth Sciences* v. 15 p. (1783-1798)

Hoffman, E.B. (1982) Water Resources Data, South Dakota: Water Year 1982 *U.S. Geological Survey Water-Data Report* SD-82-1

Howard, A.D. (1964) Model for cavern development under artesian flow, with special reference to the Black Hills NSS *Bulletin* v. 26: p. 7-16

Jennings, J.N. (1985) *Karst Geomorphology* Basil Blackwell, Oxford

Kleinkopf, M.D. and Redden, J.A. (1975) Bouguer gravity, aeromagnetic, and generalized geologic maps of part of the Black Hills of South Dakota and Wyoming *U.S. Geological Survey* Map GP-903

Knirsch, K. (1980) *Possible heat sources in geothermal waters at Edgemont, South Dakota* Unpld. M.Sc. Thesis, South Dakota School of Mines and Technology

Langmuir, D. (1971) The geochemistry of some carbonate groundwaters in central Pennsylvania *Geochimica et Cosmochimica Acta* v. 35: p. 1023-1045

Miller, T.E. (1979) *Sampling of the atmosphere and carbonate aquifer at Wind Cave, South Dakota* Open File Report to Wind Cave National Park 11 p.

Miller, T.E. (1986-87) Wind Cave: An Ancient Artesian System Modified By Hydrothermal Activity Professional Paper No. 18: p. 16-29, *Department of Geography and Geology, Indiana State University, Terre Haute* (actual date of publication 1988)

Miotke, F.D. (1974) Carbon dioxide and the soil atmosphere *Abhandlungen zur Karstund)Holenkunde* Reihe Al Heft 9

Owen, L.A. (1898) *Cave Regions of the Ozarks and the Black Hills*, Cincinnati, Ohio; The Editor Publishing Company 228 p.

Palmer, 1975 The origin of maze caves NSS *Bulletin* v. 37 p. 56-76

Palmer, A.N. (1981) *The Geology of Wind Cave*, Wind Cave Natural History Association 44 p.

Rahn, P.H. and Gries, J.P. (1973) Large springs in the Black Hills, South Dakota and Wyoming *South Dakota Geological Survey*, Report of Investigations #107 46 p.

Reagan, B.G. (1981) Miscellaneous Field Studies MF 1325, *U.S. Geological Survey*

Schwarcz, H.P. and Gascoyne, M. (1984) Uranium-series dating of Quaternary Deposits from *Quaternary Dating Methods* ed. by W.C. Mahoney New York, Tokyo

Sando, W.J. (1974) Ancient solution phenomena in the Madison Limestone (Mississippian) of North-Central Wyoming *Journal of Research U.S. Geological Survey* 2 (2) p. 133-141

White, W.B., and Deike, G.H. (1962) Secondary mineralization in Wind Cave, South Dakota *National Speleological Society Bulletin* v. 24 (2) p. 72-87

Wigley, T.M. (1973) The incongruent solution of dolomite *Geochemica et Cosmochimica Acta* v. 37: p. 1397-1402

Davis, Donald G.—Helictite Bushes—A Subaqueous Speleothem?: *NSS Bulletin* 51 120-124.

HELICTITE BUSHES— A SUBAQUEOUS SPELEOTHEM?

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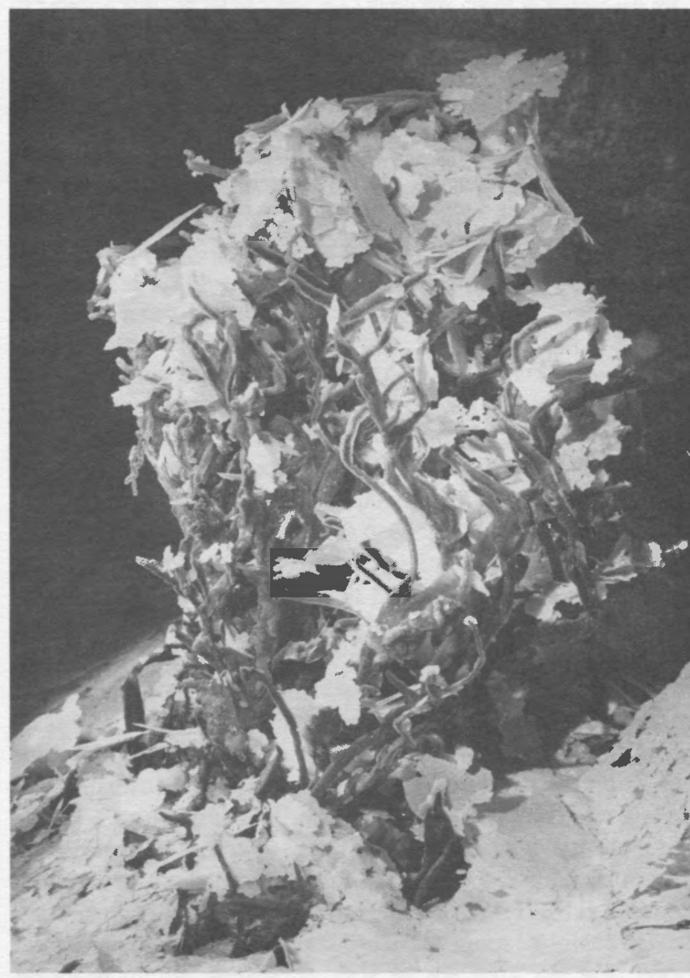
Several lines of evidence suggest that the "helictite bushes" of Wind Cave grew underwater, probably where localized plumes of thermal water once welled up into ambient groundwater then standing in the cave. In their physics of origin, helictite bushes may be more closely related to features like submarine "smoker chimneys" than to normal helictites. The chemistry of helictite-bush growth has not been investigated, but may involve calcite crystallization by the "common ion effect" where calcite-rich and gypsum-rich waters interacted.

"Helictite bushes" (Fig. 1; see also the cover of the February 1988 *NSS News*)—are a much admired, but little studied, feature of Wind Cave, South Dakota. Writers on Wind Cave have treated them as exceptionally large and complex variants on ordinary helictites, but have not regarded them as fundamentally different in character or origin (A. Palmer, 1981, 1988; Shafer, 1988). However, there is now reason to suspect that helictite bushes are far more unusual and interesting speleothems than has been assumed.

"A helictite," as defined in Hill and Forti's *Cave Minerals of the World* (1986, p. 42; 45), "is a contorted speleothem which twists in any direction, seemingly in defiance of gravity . . . All helictites originate by capillary flow under hydrostatic pressure," the rate of feeder flow being so slow that no hanging water drop is formed and crystallographic forces override gravity in directing the helictite's growth. This explanation is applied specifically to Wind Cave by Shafer (1988): "The helictite bushes . . . may form when water seeps from the cave through pores so small that the flow is controlled by capillary action and not gravity. This allows water to move uphill and deposit calcite against the force of gravity." Such interpretations assume that all helictites are subaerial (i.e., grow in air-filled, not water-filled, caves).

Wind Cave helictite bushes, however, show conspicuous evidence of former submergence, such as calcite rafts lodged in their branches. Arthur N. Palmer's *Geology of Wind Cave* (1981, p. 28) accounted for this as follows: "The helictites formed above water, and the rafts settled over them during a later rise in the water table."

Are these explanations correct? In Lechuguilla Cave, New Mexico (Carlsbad Caverns National Park), I recently discovered helictites which have grown beneath the water of shelfstone pools (*NSS Bulletin*, publication pending). This raises the question whether some helictites in other caves—particularly the helictite bushes of Wind Cave, with their peculiar morphology and signs of inundation—have also grown underwater.



Helictite bush near Calcite Lake in Wind Cave, adorned with calcite raft fragments that settled onto it when the lake level was several meters higher. Height of bush is approximately 40 cm. Photo by Arthur N. Palmer.

On May 28 and 29, 1988, with Wind Cave National Park permission and Ed LaRock's guidance, I examined several dozen helictite bushes between the Garden Gate and Calcite

Lakes for evidence bearing on this question. Nearly all Wind Cave helictite bushes are found in or near this downdip section of this axial passage complex which bisects the known cave. They typically grow upward from floors, often in intricately intertwining branching patterns, and are uniquely distinct in morphology from either ordinary subaerial helictites or from the downward-angling, usually unbranched spaghetti-like Lechuguilla subaqueous form. Several observations relevant to their origin were made:

(1) *All* Wind Cave helictite bushes observed show evidence of former submergence: calcite rafts lodged in branches; encrustations continuous with crystalline subaqueous calcite wall crust; and in some cases, films of compact, apparently water-laid sediment on upper surfaces of helictite arms.

(2) *No* bushes observed are now "alive," (i.e., none can be shown to be actively growing in air).

(3) Subaerial helictites are created by seeping water which will normally have grown dripstone, flowstone or shields in the same area; these vadose speleothems are totally absent from this section of Wind Cave.

(4) The great majority of helictite bushes grow upward from floor crusts which were apparently calcified before drainage of the cave. The bushes often originate beside precrust breakdown blocks or other obvious discontinuities that provide permeable places in the crust. During or after drainage, these floor crusts were extensively undermined and broken by subsidence of their sediment substrates, but no helictites appear to have grown after this episode of crust breakage. Even when floors were intact, it would seem unlikely that descending vadose moisture would take the most indirect possible route by bypassing outlets in walls and ceilings only to ooze up from the floors.

(5) The capillary-seepage growth mechanism of subaerial helictites typically constrains their internal feeder canals to a very small diameter (specifically .008 to .5 mm, according to Hill and Forti, p. 43). In striking contrast, Wind Cave bush branches (whose cross-sections are observable in many pieces broken naturally or by careless cavers) usually have large and irregular internal canals, often 3 mm or more in minimum diameter, and 12 mm or more along the wider axis in flattened branches (the canals are occasionally wider than the wall of the branch is thick). In air, it would be quite improbable that a helictite with a large internal tube could grow very far before gravitational effects on overflow or leakage caused conversion to a soda-straw mode. Straws, indeed, are often seen growing from subaerial helictites but *never* from any Wind Cave helictite bushes examined.

George W. Moore (pers. comm., 1989) has pointed out that helictite growth in air requires that the terminal orifice be small, but that, because of high acidity created by confined CO_2, the internal canal may be enlarged by resolution. Helictite bush sections have not been critically examined for evidence of truncated growth layers that would result from resolution. However, in bushes not heavily overcrusted, the irregularities of internal and external outline tend to coincide, suggesting that the cavity dimensions are original and not secondary features. Moreover, internal resolution would lend itself to perforation, leakage, and consequent soda-straw growth, which, as noted above, is not observed.

(6) One specific helictite bush, in an alcove on the west side of the passage a few meters south of the Emperor Maximus bush, grew upward from the floor crust several cms, where its branches encountered an overhanging ledge. The branches thereupon merged into a thin crystalline crust, whiter than the adjacent walls, for several cms up the overhang. At a sharp breakover to an upward-facing slope, helictite branches at once emerged from the upper edge of the crust and resumed independent upward growth for several more cms. Such phenomena would be consistent with a deposit created by an ascending fluid plume within a surrounding water body—but not with behavior of subaerial helictites, which, if their feeder canals were disrupted by intersecting a wall, would have no way to "remember" original orientation and resume coherent growth above.

A. Palmer—who had been informed of the subaqueous Lechuguilla Cave helictites—has recently presented, in his updated Wind Cave geology book (1988, p. 33), a revised subaerial hypothesis which addresses some of the above points: "It is possible for helictites to grow under water, but only in unusual chemical conditions. These [Wind Cave bushes] most likely formed in air, growing upward from deposits of weathered limestone powder that accumulated from upper levels, apparently when the powder was moistened by pools of water immediately below. Later rises in water level coated them with layers of calcite, making them thicker and sturdier."

Palmer's new model does account for their prevalent growth from floors and the lack of association with vadose decorations, as well as the universal evidence of inundation (assuming that growths originating near the water table would inevitably be flooded by normal fluctuations) and the failure of bushes away from the water table to remain active. It does not, however, explain their having feeder canals larger than capillary size (resolutional enlargement of canals would be unlikely in Palmer's wicking model, because CO_2 in helictite-forming water derived from open pools should already have reached equilibrium with the relatively low CO_2 level to be expected in cave air). Helictite growth in a water medium would place no constraint on the dimensions of the canals. Nor does Palmer's new mechanism account for the bush that reorganized its upward growth after interruption by a ledge. Finally, a wicking process, such as Palmer suggests, should take place anywhere that standing cave water abuts a shore of absorbent material and where the cave atmosphere allows evaporation. This is a situation common enough that one might expect helictite bushes, if caused by it, to be widespread in Wind Cave and other caves; yet they are extremely localized without apparent correlation with factors that would have favored wicking.

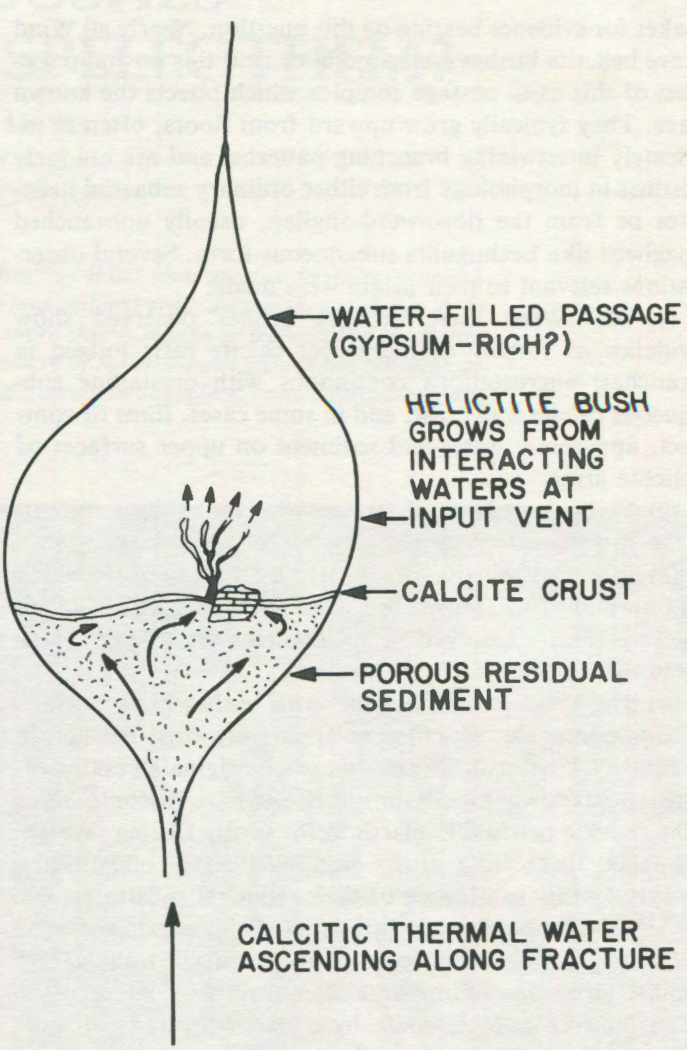
Points 1 through 5 in the above observations, though not impossible in a subaerial model, are simpler to account for in a subaqueous one. Point 6 appears to be explainable only by subaqueous growth. Accordingly, it seems highly probable that the helictite bushes in this section of Wind Cave grew underwater before the water table fell beneath their levels (this conclusion would be strengthened if diving were to find that such bushes continue to substantial depths below the present water table). Speleologists in the past had not seriously considered this possibility because they had neither seen helictites growing underwater nor thought of any plausible means by which this might happen. What mechanisms, then, could allow for the growth of subaqueous helictites and for the differences between the Wind Cave and Lechuguilla forms?

In both cases, they appear to be composed of calcite. The Lechuguilla helictites, unlike those in Wind, are usually unbranching and grow erratically inward and downward from the walls of shelfstone-rimmed pools, never turning above the shelfstone level. They are finely crystalline, with uniformly circular cross-section and relatively uniform diameter usually between 1 mm and 8 mm. They have tiny, regular central canals. In all 14 known localities, partially-dissolved gypsum blocks or crust lie upslope from the helictite pools.

I hypothesize that in Lechugulla Cave, mixing of gypsum-enriched stringers of water into the calcite-saturated pools has triggered growth of helictites at the mixing points where calcium ions are in excess, via the so-called "common ion effect." I speculate that similar mixing chemistry may have been involved at Wind Cave, but with variations in the mechanics to account for the different morphologies.

The Wind Cave helictite bushes have intricate, usually ascending dendritic branching (possibly to a degree unknown in helictites from any other cave), with branch dimensions varying from needle-thin (as in terminal segments of Emperor Maximus) to 25 mm or more wide. They are relatively irregularly and coarsely crystalline, with individual crystal domains apparently to 25 mm or more in length in those not too overcrusted to see the basic structure. Cross-sections vary from circular to angular or flattened into a strap-like form. The internal canals vary correspondingly in shape, with a variety of irregular partitions and obstructions common. These blockages are probably the cause of the numerous bifurcations that result ultimately in the elaborately arborescent pattern.

Their growth up from floors of passages at that time beneath the water table, and the absence of gypsum blocks, imply differences in mineral sources from those in the



restricted shelfstone-pool environments in Lechuguilla. There are at least two instructive parallels to the Wind Cave situation: the cave "geysermites" reported in *Cave Minerals of the World* (pp. 41–42); and the recent discovery of tubes extended from the sea bed where submarine springs vent in ocean floors.

Geysermites are interpreted as forming "from upwelling, thermal water," but it is not clear whether the geysermite cones were submerged when formed. Hill and Forti (1986) do not report any specific investigation of the chemistry of geysermite development, but hypothesize (p. 42) that "water rich in dissolved carbon dioxide releases carbon dioxide gas as the water ascends into the cave . . . this release . . . causes the precipitation of calcite." This model would work only if the geysermite were surrounded by air into which the CO_2 could disperse. However, substantial flow of upwelling water in the enclosed cave environment would be likely to be associated with flooding and submergence of the exit point, even if it was not already at a water table—a situation similar to that I propose for the Wind Cave helictite bushes.

The suboceanic spring-vent tubes were discovered on the sea floor by deep-diving vehicles. Those initially sampled (e.g., Rona, 1986) were composed of sulfide minerals and had grown where mineralized submarine hot springs ("smokers") interacted with seawater. Subsequently (Ritger et al., 1987), similar "chimneys" up to 1 m high—but composed of carbonate minerals—were found where methane-rich seeps reacted with seawater. The chemistry, even in the carbonate case, may not be close to the Wind Cave analogue, but the existence of submarine "chimneys" certainly establishes that rising fluid plumes can create ascending tubular mineral structures in a subaqueous medium.

It is likely that the more delicate and complex Wind Cave deposits are also creations of former thermal water rising along fractures underlying this major axial passage of the cave. Higher temperature in the entering water than in the ambient water would help account for the ascending growth tendency of most helictite bushes. In mechanics of development, Wind Cave helictite bushes are probably more akin to geysermites and submarine "chimneys" than to either ordinary helictites or the non-thermal subaqueous helictites of Lechuguilla Cave.

A related case has been reported in the speleological literature: Peck (1979) described small tubular speleothems, up to several inches long, some of which had large central canals and which Peck believed had been created by injection of ascending fluid into geode-like phreatic caves. Peck defined the growths as subaqueous "helictites" and "stalactites" (notwithstanding that most of the "stalactites" had grown upward). Like some submarine chimneys, these speleothems were composed of sulfide minerals, and it would probably be more appropriate to classify them as a spelean variant of the marine "smoker" phenomenon.

Although the physical characteristics of subaqueous helictites differ between Wind and Lechuguilla Caves, the chemistry of deposition may have been related. It is possible that Wind Cave helictite bushes were created, like those in Lechuguilla, by interaction between calcium bicarbonate- and calcium sulfate-rich waters. The distinctions between the helictite geometries in the two caves might mean that the relationship between the ambient and inflowing waters in one cave was in mirror image in the other.

In Lechuguilla, calcitic water bodies have been invaded by lesser stringers of sulfate-bearing seepage. In Wind Cave, study by Margaret Palmer (1988) indicates that the dolomitic bedrock surrounding Wind Cave originally contained zones of gypsum or anhydrite which were subsequently dissolved or replaced by silica and calcite. If these evaporites had not been fully removed from the Pahasapa at the time of helictite-bush growth, the ambient groundwater might then have been high in gypsum. The intruding thermal plumes, coming from lower strata, might have been rich in calcite but not gypsum. This would enhance the tendency toward ascending helictite growth; calcite is far less soluble than gypsum, so the less gypsiferous solution should (temperature differences aside) be less dense than the surrounding gypsum-rich water and would therefore rise through it. The comparatively large size of the Wind Cave helictites and of their feeder canals, with respect to those in Lechuguilla, may reflect a faster fluid input rate in the Wind Cave situation, though the larger crystal size in Wind Cave may imply slower growth of the speleothems themselves, perhaps implying less intense reactions in the Wind Cave case.

Fig. 2 illustrates the proposed mechanism of helictite-bush growth. A phreatic passage is flooded with water which is presumed to have dissolved gypsum from the surrounding bedrock. Thermal, calcite-rich water rises into porous residual floor sediment from underlying fractures. At the sediment surface, the two waters react via the common ion effect to create a calcite-crust barrier. This directs venting preferentially to weak points in the crust, where vent tubes in the form of helictite bushes then extend out from the exit point.

Certain anomalies in the growth of a minority of helictite bushes do pose challenges to the ascending-water framework I propose. Especially problematic is the famous Emperor Maximus, the largest known bush complex, in which scores of intertwined branches, growing downward from a ceiling joint, extend 2 meters or more from the ultimate point of origin. After examining Emperor Maximus in its context, I believe its unusual orientation is attributable to peculiarities in the local "plumbing." Within a few meters before Maximus, many smaller bushes grow upward from floor crust in the normal way. Maximus itself is situated where the main passage pinches into a joint. If a thermal plume had been rising along that joint beyond the end of the open passage, it would have found no floor crust to issue from, and would have been diverted laterally to emerge at the most open level of the joint, in this case along the ceiling. The branches of Maximus are also thinner-walled and more slender than in most others, and if temperature gradient was the primary determinant of growth direction, Maximus may have equilibrated with surrounding water faster than most bushes, allowing it to grow atypically downward.

Near Calcite Lake, a few smaller bushes also grow from ceiling joints and wall crust, in association with more numerous floor bushes; the above considerations probably apply to these as well. The same area has a more puzzling phenomenon: interspersed with typical bushes on floor crusts are a number of cryptic stalagmite-like objects, cylindrical or parabolic with round tops, averaging 75–100 mm in diameter by 30 cm high, from whose irregular surfaces bristle small, tentacle-like, downward-angling helictite-bush arms up to 10 cms long. These are almost certainly not true stalagmites; none have stalactites above, nor do they correlate with obvious ceiling drip points. They are apparently another subaqueous form created by water rising from the floor, but I have no explanation for their stalagmitic symmetry.

In certain passages, especially off the Garden Gate area, groups of helictite bushes show consistent deflections, usually in the downslope direction, as if influenced during growth by currents in the surrounding water (in one case the direction is reversed behind the inner angle of a corner, where there was probably an eddy). The deflections are generally consistent with fossil ripple marks near Calcite Lakes, which appear to indicate flow in the downdip direction.

If I am correct in proposing that these are subaqueous speleothems, many factors beyond the scope of my observations may have been involved, and my chemical scenario is offered only as a speculative possibility pending analytical study. Ed LaRock intends to do sampling and analysis of broken pieces to confirm the composition of the helictite bushes and to elucidate the temperature and geochemistry of development. This should be rewarding; understanding the origin of helictite bushes is not only interesting in itself, but is important to more accurate knowledge of the history of events in Wind Cave.

REFERENCES

Hill, C. and Forti, A. *Cave Minerals of the World*. National Speleological Society, Huntsville, AL. 238 p. 1986.

Palmer, A. *An Ancient World Beneath the Hills: Wind Cave*. Wind / Jewel Caves Natural History Association, Hot Springs, SD. 49 p. 1988.

Palmer, A. *The Geology of Wind Cave*. Wind / Jewel Caves Natural History Association, Hot Springs, SD. 44 p. 1981.

Palmer, M. "Influence of Former Evaporites on South Dakota Caves." *Abstracts NSS Convention Program 1988*. National Speleological Society, Huntsville, AL. 1988, p. 31.

Peck, S. "Stalactites and Stalagmites of Marcasite, Galena and Sphalerite in Illinois and Wisconsin." *NSS Bulletin*, Jan. 1979, p. 27–30.

Ritger, S., Carson, B., and Suess, E. "Methane-derived Authigenic Carbonates Formed by Subduction-induced Pore-water Expulsion Along the Oregon-Washington Margin." *Geological Society of America Bulletin*, Feb. 1987, p. 147–156.

Rona, P. "Mineral Deposits from Sea-floor Hot Springs." *Scientific American*, Jan. 1986. p. 84–92.

Shafer, D. "The Geology of Wind Cave National Park," *Rocky Mountain Caving*, Autumn 1988, pp. 23–25.

Nepstad, J. and Pisarowicz, J.—Wind Cave, South Dakota: Temperature and Humidity Variations: *NSS Bulletin* 51: 125-128.

WIND CAVE, SOUTH DAKOTA: TEMPERATURE AND HUMIDITY VARIATIONS

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It has long been thought that caves possess constant temperatures and humidities; never changing, even during periods of extreme temperature fluctuations on the surface. Although this may be true in very remote parts of large cave systems, most caves exhibit enough variations in this respect to warrant serious investigation (Wigley and Brown, 1976; Bramberg, 1973; Stark, 1969; Cropley, 1965; Davis, 1960; Little, 1952).

In his investigations into temperature fluctuations at Greenbrier Caverns and Ludington's Cave, Cropley (1965) noted that cave temperatures would vary widely under the influences of surface temperature fluctuations and the varying flow patterns of air and water currents within the studied caves. Such changes in cave temperatures were attributed to a combination of cold wintertime air and water entering these caves, evaporative cooling by less-than-saturated cave winds, and, to a lesser extent, by the conduction of heat from the surface through the rock above the caves.

Wind Cave, located in the southern Black Hills of South Dakota, is aptly named. Winds in excess of 120 kph have been recorded at the entrance to the cave. Conn (1966) provided evidence that such winds are caused by barometric pressure changes with the magnitude of the wind related to the cave's vast size (currently 84 km surveyed) and the small size and number of entrances to the cave (2 natural entrances each approximately 300 cm^2; two man-made entrances, one 3 m^2 and the other sealed by an elevator door).

Few caves experience the volume of airflow which Wind Cave exhibits. On average, almost 3000 m^3 of air enter or leave the cave per hour when the Walk-In Entrance is open. With such huge exchanges of air, Wind Cave provides an excellent study site to evaluate not only the effects of cave winds upon temperature and humidity variations but also the effects of man-made intrusions such as digging entrance tunnels and shafts into the delicate cave environment.

METHOD

Between 11 November 1984 and 25 March 1988 temperature and humidity readings were recorded on the surface and at 14 standardized sites located along the tour route in Wind Cave (Fig. 1). These readings were taken on 109 different days during this time period although all sample sites were not sampled each day (Table 1). A sling psychrometer (Princo Instruments, model D-430) was used to record the temperatures and humidities using the standard wet bulb/dry bulb method. Wind speed and direction at the Walk-In Entrance to the cave were recorded as well as whether or not the door to the cave was open or closed. Miscellaneous observations such as sky cover, precipitation, unusual wet or dry conditions in the cave at the recording sites, etc., were also noted. It took approximately 1.5–2 hours to traverse the cave to make the required measurements with the sites sampled in the same sequence on each

Table 1. Means, standard deviations, and sample sizes at each of the sample locations.

Sample	Location	Temperature (C) Mean	SD	n	Relative Humidity Mean	SD	n
PO	Post Office	9.14	1.25	109	88.01	8.23	109
RKY	Rookery	9.12	1.02	109	88.36	6.83	109
FLO	Flowstone	10.16	0.53	109	89.96	4.75	109
CHR	Church	10.25	0.52	109	91.67	4.38	109
T87	Top of the 87	12.90	0.18	71	93.77	2.51	71
COL	Coliseum	12.78	0.20	71	94.07	2.72	71
FG	Fairgrounds	12.87	0.13	71	92.94	2.40	71
SXR	Second Crossroads	12.73	0.29	71	90.81	2.77	71
TEM	Temple	11.63	0.35	101	90.95	3.92	101
ASS	Assembly Room	11.64	0.33	109	91.20	5.47	109
ES	Eastern Star	11.74	0.26	109	92.05	3.77	109
GDN	Garden of Eden	11.81	0.28	109	92.22	4.00	109
LE	Lower Elevator	11.16	0.46	109	86.61	9.05	109
UE	Upper Elevator	11.27	0.44	109	88.73	7.79	109

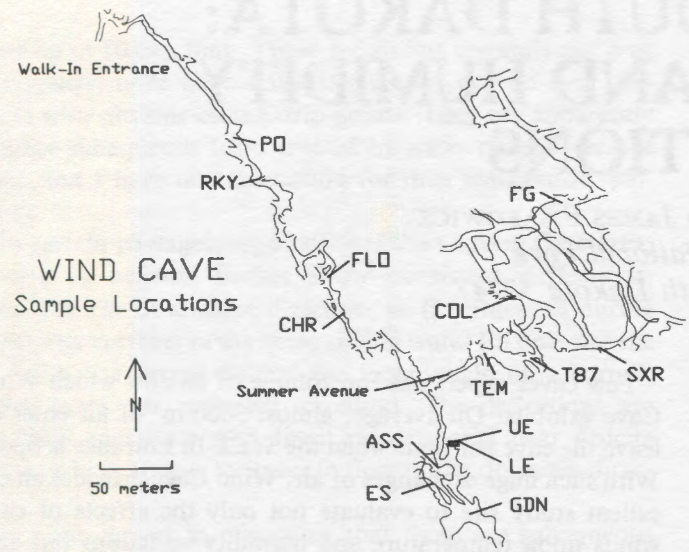

Figure 1. **Map of Wind Cave's Tour Routes indicating the sampling locations. This map represents approximately 2 km of the 84 km of known cave passage.**

day. As all the data collection locations were on the tour trails, travel time from location to location was not a significant factor. Thermometers were read to within 0.5° F. The resulting uncertainty in relative humidity is approximately 3% (Wefer, 1989). In addition to the standardized sample locations several "deep" caving trips collected temperature and humidity readings during the study.

Ten different thermometers were used in the study because of breakage in the cave. Each thermometer was standardized against a laboratory thermometer in a constant temperature environment in the temperature ranges normally found in the cave. The thermometers chosen for use in the cave were matched pairs in that each pair gave identical readings in the temperature ranges expected to be found in Wind Cave. Thermometers were not calibrated beyond the temperature ranges found inside the cave.

RESULTS AND DISCUSSION

Figure 2 indicates variations in temperature throughout the sampled area as a function of air direction at the Walk-In Entrance. Several items are significant in this figure. First note the large temperature differentials in the Post Office [PO] (t = 9.619, df = 107, $p < .001$), Rookery [RKY] (t = 12.353, df = 107, $p < .001$), Flowstone [FLO] (t = 6.618, df = 107, $p < .001$) and Church [CHR] (t = 6.182, df = 107, $p < .001$). Significant differences also observed at the upper [UE] and lower [LE] elevator landings (t = 2.222, df = 107, $p < .05$ and t = 2.290, df = 107, $p < .05$, respectively). The temperature differentials are not significant at the other sample locations because of air flow direction.

In general the air outside the cave is colder and drier than the air underground. During the period of this investigation,

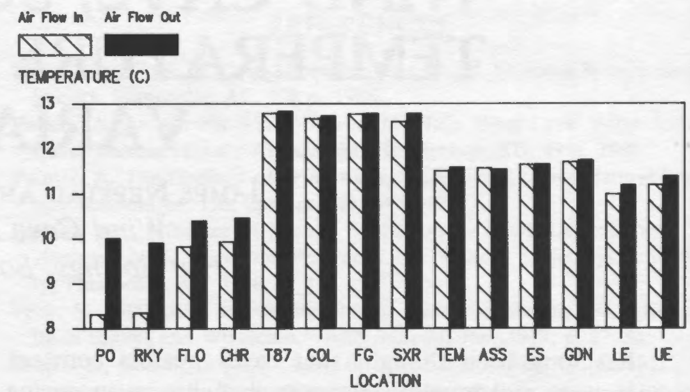

Figure 2. **Mean cave temperature at sample locations as a function of the direction of air flow at the Walk-In Entrance. See Figure 1 for the relative locations of these points to each other.**

the outside temperature averaged 6.9° C and the relative humidity 56.7%. When rising barometric pressure forces this colder, drier air into the cave, the temperatures in the cave near the entrance drop as the cave gives up heat to warm the colder surface air. These data indicate that air movement into the cave at the Walk-In Entrance results in significant differences in cave temperature as far into the cave as the Church (approximately 300 m direct distance from the Walk-In Entrance). Although it is tempting to use distance measures in this analysis, because of the extremely complicated, three-dimensional nature of Wind Cave, what those distances would be are unknown. Nevertheless, it appears that significant temperature fluctuations result from air movement shifts as far along the tour route as the Church.

Also noteworthy are the significant temperature differentials at both elevator landings. Despite the elevator doors and a building over the elevator shaft, air flow down the shaft significantly effects cave temperature at these locations.

The general shape of the data in Figure 2 is also revealing. Moving along the graph from the left, temperatures can be seen to rise until they appear to stabilize in the area of the cave called the Top of the 87 [T87] and remain fairly constant until the Second Crossroads [SXR] where the temperatures drop again. The next four readings are relatively constant with more variability in the last two sets of data, the Lower Elevator [LE] and Upper Elevator [UE] landings.

Using correlated t-tests on overall temperature readings reveal that the data separate into five separate clusters. The first cluster involves the Post Office and Rookery temperatures. The temperatures at these locations are not statistically different from each other but are statistically different from the other cave temperatures. In a similar manner, the Flowstone and Church make another cluster. A third cluster involves the Temple, Assembly Room, Eastern Star and Garden of Eden locations. The Top of the 87, Coliseum, Fairgrounds, and Second Crossroads comprise the fourth cluster. The final cluster includes the two elevator landing data sets.

The entrance cluster (PO and RKY) is separated from the second cluster (FLO and CHR) by a series of trails which constricts before opening into larger passage again in the vicinity of the second pair of locations. Perhaps this constriction along the main corridor of the tour trail accounts for the separation of these four locations into two separate clusters.

The third cluster locations, (TEM, ASS, ES, GDN) have in common that they are near the elevator entrances, but separated by some distance from the direct effect of the air movement up and down the elevator shafts. In a similar vein, the two elevator data collection locations are colder than adjacent sampling points probably because of air leaking into the cave down the elevator shaft. Such leakage of surface air into the cave at these locations tend to depress these temperatures despite the elevator doors and the building which sits atop the elevator shaft.

Finally, the eastern cluster (T87, COL, FG, SXR) consists of locations either in the upper levels of the cave (T87, COL, FG) or, as in the case of the Second Crossroads [SXR], at the farthest distance (straight line) from the Walk-In Entrance. The temperatures in this cluster appear to be the more stable and correspond most closely to temperature readings that have been taken on "deep" trips into Wind Cave (12.5–12.8° C).

Figure 3 indicates the relationship between location, the direction of air flow and relative humidity. Interestingly, all of the locations with statistically significant differences—the Temple [TEM] (t = 2.619, df = 99, $p < .05$); Assembly Room [ASS] (t = 2.621, df = 107, $p < .05$); Second Crossroads [SXR] (t = 2.439, $p < .05$); Eastern Star [ES] (t = 2.814, df = 107, $p < .05$) and Garden of Eden [GDN] (t = 3.210, df = 107, $p < .01$)—are relatively far from the Walk-In Entrance, fairly close to the elevator entrances, or seem to connect at about the same cave elevation through a passage named "Summer Avenue."

Summer Avenue is an interesting place along the tour route in Wind Cave. No matter which direction the air is moving at the Walk-In Entrance, the air movement is always from east to west through this passage. The reason behind this unusual air flow is not currently understood and has not been addressed in any literature about Wind Cave. Whatever the reason for this air-flow pattern, each of locations with significant differences in humidity as the air flow changes, are along passages on approximately the same level near the elevator entrances.

The humidity data collected show less of a tendency to

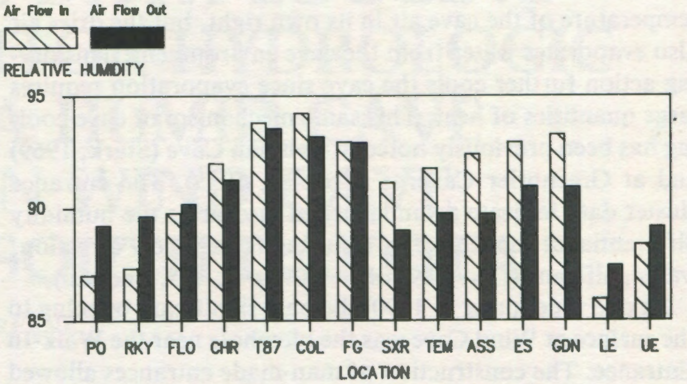

Figure 3. **Mean relative humidity at sample locations as a function of the direction of air flow at the Walk-In Entrance. See Figure 2 for the relative locations of these points to each other.**

"cluster" than the temperature data. In this regard, only the elevator locations (LE and UE) may be referred to as a cluster. These locations relative humidities are not statistically different from each other. Yet the lower Elevator/Assembly Room and Upper Elevator/Eastern Star differences are statistically significant using a correlated t-test (t = 3.166, df = 107, $p < .01$ and t = 2.511, df = 107, $p < .05$, respectively). Leakage of colder, drier air into the cave down the elevator shaft is responsible for the dry conditions in the cave near the elevator shaft.

Although vast quantities of air move into the cave through the Walk-In Entrance, statistically significant differences were not observed at the entrance sampling locations. This finding can be attributed to two factors. First, there was great variability in the humidity readings in this part of the cave. For example, in the Post Office, relative humidities as low as 60% were recorded during winter months when the cave is inhaling large quantities of air and as high as 100% when the cave was exhaling. Second, a large measurement uncertainty factored into all of the humidity data. This was because of the relatively gross precision of the temperature data. With temperatures read to within 0.5° F the uncertainty in relative humidity was approximately 3% (Wefer, 1989). Such an uncertainty in the measurement of this variable would tend to mask any real differences that might actually exist.

Statistically significant differences were also noted between the Fairgrounds and Second Crossroads (t = 2.913, df = 69, $p < .05$) and the Second Crossroads and the Assembly Room (t = 3.021, df = 69, $p < .01$). Perhaps these differences are related in some way to the Summer Avenue phenomenon.

A reexamination of results from Figures 2 and 3 for the elevator cluster indicates the double effect of dry air entering the cave. Not only does this surface air lower the temperature of the cave air in its own right, but the drier air also evaporates water from the cave environment. This drying action further cools the cave since evaporation requires large quantities of heat. This same mechanism of cave cooling has been previously noted at Lehman Cave (Stark, 1969) and at Greenbrier Caverns (Cropley, 1965). The entrance cluster data indicate a similar trend insofar as the humidity differential at the Rookery, resulting for air flow direction, was significant ($t = 1.89$, df $= 107$, $p < .05$, one-tail).

At one time (prior to 1890) the only significant opening to the surface at Wind Cave was the blowhole near the Walk-In Entrance. The construction of man-made entrances allowed for a huge increase in airflow into and out of the cave, bringing with it changes in the cave climate.

The greatest harm to the cave itself may come from the evaporation of moisture in the cave. Many of the cave's speleothems are directly dependent upon the amount of water available. Stalactite growth may be slowed or even stopped when less dripping water is available. Such was the case at Carlsbad Caverns until air locks at the elevators slowed this process.

There is also considerable evidence showing that aragonite tends to form in preference to calcite in areas with high evaporation rates (Hill and Forti, 1986). Thus, a change in cave climate could possibly change the very chemical structure of the speleothems in the cave.

Cave fauna will also be disturbed by a change in the cave climate. Today Wind Cave's fauna is relatively sparse and one can wonder whether it was more abundant sometime in the recent past. Animals which have evolved in the cave's environment over thousands of years probably have little tolerance for major temperature changes. Many of these animals live on moist surfaces. When evaporation takes place, these surfaces can become remarkably cool. Different species of bats prefer different environments in the cave for roosting and they also could be disturbed by a change in the cave climate.

Unfortunately, there is no base line to assess the changes that opening up large entrances into Wind Cave has wrought. The climate of Wind Cave is complex and probably affects the cave in ways that we have yet to understand.

REFERENCES

Bamberg, S. A. "Environments in Lehman Caves, Nevada." *NSS Bulletin*, v. 35(2), p. 35–47. April 1973.

Conn, H. W. "Barometric Wind in Wind and Jewel Caves, South Dakota." *NSS Bulletin*, v. 28(2), p. 55–69. April 1966.

Cropley, J. B. "Influence of Surface Conditions on Temperatures in Large Cave Systems." *NSS Bulletin*, v. 27(1), p. 1–10. January 1965.

Davis, W. E. "Meteorological Observations in Martens Cave, West Virginia." *NSS Bulletin*, v. 22(2), p. 92–100. July 1960.

Hill, C. and Forti, A. *Cave Minerals of the World*. National Speleological Society. Huntsville, AL. 1986.

Little, W. "Significant Air-streams in Ogof Ffynnon Ddu." *Trans. Cave Research Group*, v. 2(2), p. 135–148. 1952.

Stark, N. "Microecosystems in Lehman Cave, Nevada." *NSS Bulletin*, v. 31(3), p. 73–81. July 1969.

Wefer, F. L. "On the Measurement of Relative Humidity in Cave Meteorology Projects." *Nittany Grotto News*, v. 36(1), p. 6–14. Winter 1989.

Wigley, T. M. L. and Brown, M. C. "Cave Meteorology." In T. D. Ford and C. H. D. Cullingford (eds.). *The Science of Speleology*. p. 329–344, Academic Press, New York. 1976.

Greene, Earl A.—Use of Electronic Data-logging Equipment to Monitor Hydrologic Parameters in a Humid Cave Environment in Wind Cave National Park, South Dakota: *NSS Bulletin* 51: 129-131.

USE OF ELECTRONIC DATA-LOGGING EQUIPMENT TO MONITOR HYDROLOGIC PARAMETERS IN A HUMID CAVE ENVIRONMENT IN WIND CAVE NATIONAL PARK, SOUTH DAKOTA

EARL A. GREENE

The U.S. Geological Survey has installed electronic data-logging equipment at Windy City Lake in Wind Cave to monitor changes in water level, water temperature, air temperature, and barometric pressure. These parameters are being monitored in response to the concern of park managers that water levels in this area of the cave are progressively declining. Description of equipment and the methodology used to collect data in this remote area of the cave are presented. Preliminary results for the first 6 months of the study indicate that the lightweight, durable data-logging equipment works successfully in the humid cave environment. Water levels at Windy City Lake appear to slightly fluctuate in response to barometric pressure changes, and have declined 0.176 m from January 21, 1988, to July 18, 1988. Water temperature in Windy City Lake and temperature of the air in the cave are constant at 13.8 and 15.0°, respectively.

INTRODUCTION

Managers of Wind Cave National Park, located in the southern Black Hills of South Dakota, are concerned that water levels in the cave at the lakes region (Calcite Lake and Windy City Lake) have declined since the lakes were discovered in the early 1970's (Fig. 1). Because of this concern, the U.S. Geological Survey in cooperation with the U.S. Park Service have initiated a study to monitor changes in water level, water temperature, air temperature, and barometric pressure continuously at Windy City Lake with electronic data-logging equipment. Barometric pressure at the land surface is being monitored by park personnel with a recording barograph during the study. Data collection began in January 1988 and is scheduled to end in October 1989.

Windy City Lake and Calcite Lake, at 169 m below the elevation of the visitor center, are the deepest known points in Wind Cave. The lake may represent the water table in the Madison aquifer and is assumed to be the groundwater table at that point (Palmer, 1988). Windy City Lake is accessible only by a 4-km route through some of the most scenic and physically demanding portions of the cave. This trip may take as long as 4 hours one way. Because of the remote site, difficult access, and humid environment, lightweight and durable equipment are required to collect hydrologic data at the lakes.

This paper describes the methods used to make hourly measurements of water level, water temperature, air temperature, and barometric pressure at Windy City Lake. The equipment used, installation procedures, and calibration results are discussed and preliminary data are presented.

EQUIPMENT

The collection and recording of hydrologic data at Windy City Lake is best accomplished by using electronic equipment that is miniaturized, battery operated, has large data storage capacity, and is capable of multiprobe monitoring. Even though there are a number of data loggers available commercially that have these attributes, the Campbell Scientific Instrument's CR-10[1] has been selected as the data-logger for use in the study. The CR-10 is a battery-operated, small, lightweight, unattended field-data recording and storage module capable of operating for many months without maintenance. In addition, the data logger is programmable and may be used with a number of different sensors. The main workings of the data logger are: (1) Output signals of the sensors are converted to digital values, (2) measurements are processed by internal programs to provide data values, such as minima, maxima, and averages, and (3) processed results are stored for retrieval at a later date (Campbell Scientific Instruments, 1987). A number of different peripheral devices, such as a portable keyboard/display module, external storage module, external power source, wiring panel, and wiring interface help give this instrument the portability and miniaturization that is needed in the cave environment. Data may be downloaded to a microcomputer that is outfitted with an accessory

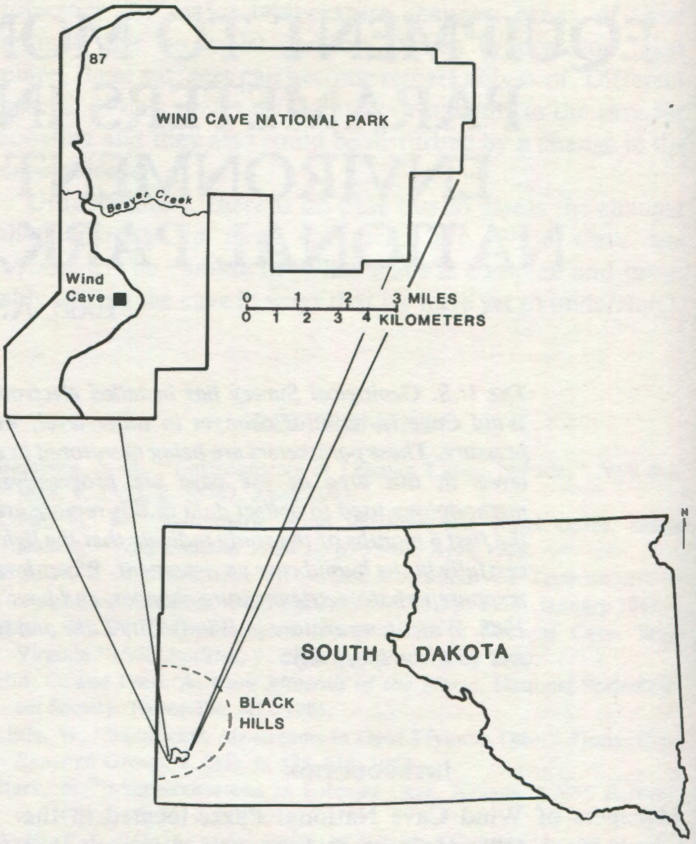

Figure 1. **Location of study area.**

processor board. Manipulations of the data may be performed with computer graphics and statistical software.

Power to the data logger is from either an internal or an external 12-volt battery source. To prolong the life of the batteries, two batteries are wired in parallel, and the data logger will monitor voltage and draw power from the battery with the most charge. When the voltage in the battery that is being used drops to less than the unused battery voltage, the data logger will automatically switch to the charged battery.

The data logger, wiring panel, wiring interface, external storage module, and internal battery source are stored in a molded plastic, air-tight box. Because the data logger is an electronically powered unit, it is important that the wiring and electronic parts in the box be kept dry. Humid air or moisture in the box may cause the instruments to short circuit resulting in a malfunction of the sensor and loss of the collected data.

Air in the cave near Windy City Lake has a relative humidity of about 95%, which caused the instruments to malfunction shortly after installation. This malfunction problem was solved by installing an external desiccant drying system. A portable vacuum pump was used to vacate the humid air from the air-tight box after the lid was closed. The humid air outside the box was allowed back into the box through the external desiccant system. The outside air was vented through this external desiccant system to allow equalization of the barometric pressure inside and outside of the box.

Two Geokon model 4500 vibrating wire pressure transducers (sensors) were used to measure fluid pressures (air and water) in the cave environment. These transducers are designed for low pressures of 34.475 kilopascals (kPa) (0 to 3.515 m) and measure water levels to accuracies of 0.005 m. The transducer uses a sensitive diaphragm that is coupled to the vibrating wire. When electromagnetic coils pluck the wire that is connected to the diaphragm, the wire vibrates and the frequency output is converted to a digital pressure value. As the fluid pressure changes, a deflection of the diaphragm takes place causing a change in the wire tension, and changing the frequency output.

The first transducer was mounted in the lake and makes hourly water-level measurements. This transducer is pressure compensated by venting the back side of the transducer diaphragm to the cave atmosphere. Measurement of water levels with this transducer was referenced to a staff gage that was mounted in the lake and is referenced to an arbitrary datum. This transducer also has a built-in thermistor to correct the output of the vibrating wire for fluid-temperature changes and measures water-temperature changes of the lake. Temperature measurement of the thermistor was calibrated to a standard American Standard Temperature Measurement (ASTM) thermometer and is accurate to ± 0.2 °C. Water temperature was measured during each visit to the site to make sure the sensor was still in calibration.

The second transducer was mounted to the ceiling of the cave near Windy City Lake and sealed so that only the front side of the diaphragm is exposed to changes of air pressure in the cave. This transducer was referenced at the land surface to a barometric pressure measurement made by the National Weather Service at the Rapid City Regional Airport and corrected for elevation changes. The transducer has a built-in thermistor that was calibrated to an ASTM thermometer to correct the output of the vibrating wire and measure air temperature changes in the cave. Air temperature of the cave was measured during each visit to the site to make sure the transducer was still in calibration.

RESULTS AND DISCUSSION

Three site visits to Windy City Lake to service and conduct calibration field checks have been completed. During each site visit, the temperature measurement of each transducer was compared to water temperature and air temperature measurements made manually with an ASTM thermometer. Water-level measurements of the transducer was compared to the staff gage. For the first 6 months of data collection the transducers were still in calibration. However, calibration checks on the transducer measurement of barometric pressure have not been conducted. These checks were to begin using a portable field barometer during subsequent site visits.

Water levels on July 18, 1988, of Windy City Lake were 0.175 m lower than when the study began on January 21, 1988. Hourly water-level data from January 21 through July 18 for Windy City Lake are presented in Figure 2. The water-level decline may be seasonally related based on the increased rate of decline during the summer months. The daily water-level fluctuations probably are caused by changes in barometric pressure in the cave (Fig. 2). The relation between Windy City Lake water levels and cave barometric pressure will be examined throughout the study.

Water and air temperature have been constant at 13.8° C and 15.0° C, respectively (Fig. 3). The transducer and ASTM thermometer water temperature measurement of 13.8° C at Windy City Lake compares closely with a water temperature of 14.0° C measured by Millen and Dickey (1987). The 1.2° C difference in water and air temperature was not expected at this level of the cave. Water and air temperature were expected to be the temperature of the surrounding rock. This temperature difference may be caused by evaporation from the lake, air movement in the cave, or influx of cooler water into the lake.

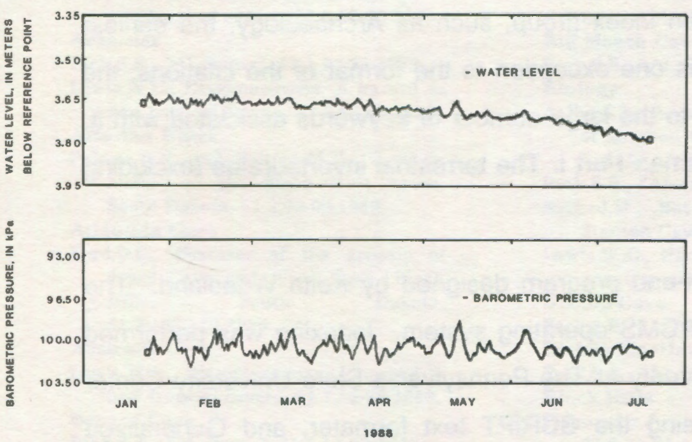

Figure 2. **Comparison of water-level and barometric pressure fluctuations at Windy City Lake from January through July, 1988.**

The 21 days of missing data shown by the dashed line in Figure 3 was caused by a programming error. The spikes of higher air temperature on March 3 and April 7 are the measurement of body heat warming the air near the probe while personnel were servicing the data-logging equipment.

SUMMARY AND CONCLUSIONS

Electronic data-logging equipment to monitor hydrologic conditions will operate in the remote, inaccessible, and humid environment of Wind Cave. After 6 months the transducers remained calibrated with only a 21-day loss of water temperature data caused by a programming error. A desiccant drying system was developed to keep the instruments from short circuiting due to the humid cave atmosphere.

The water level, water temperature, air temperature, and barometric pressure data at Windy City Lake, from January 21 to July 18, 1988, indicate that the water level at Windy City Lake has declined 0.175 m. Water levels appear to fluctuate daily in response to barometric pressure changes in the cave atmosphere.

Water temperature has been constant at 13.8° C and air temperature is constant at 15.0° C. The 1.2° C difference between the water and air temperature at Windy City Lake was not expected, but may be caused by evaporation from the lake, air movement in the cave, or influx of cooler water into the lake.

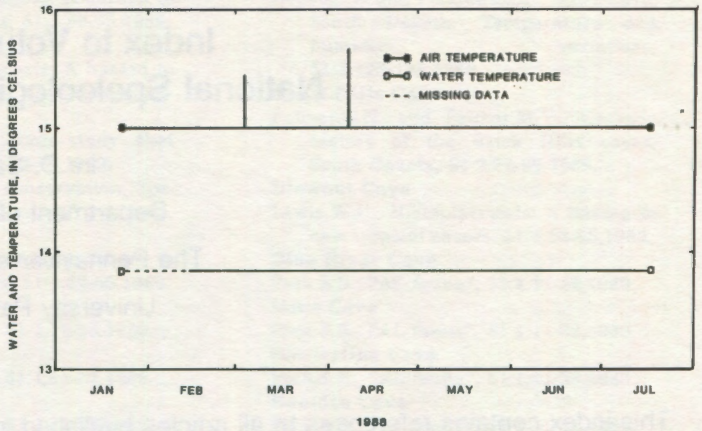

Figure 3. **Comparison of water and air temperature at Windy City Lake, January through July 1988.**

REFERENCES

Campbell Scientific Instruments. 1987. CR10 measurement and control module operator's manual. Campbell Scientific, Logan, Utah, 194 p.

Millen, T. M., and Dickey, N. D., Jr. 1987. A stable isotopic investigation of waters and speleothems in Wind Cave, South Dakota: An application of isotopic paleothermometry: NSS Bulletin, v. 49, p. 10–14.

Palmer, A. N. 1988. An ancient world beneath the hills, Wind Cave: Wind Cave National History Association, 49 p.

FOOTNOTE

[1]Use of brand names is for identification purposes only and does not constitute endorsement by the U.S. Geological Survey.

Index to Volume 51 of the National Speleological Society Bulletin

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This index contains references to all articles published in volume 51 parts 1 and 2. The index consists of three parts. The first of these is a **keyword index** which starts on **page 133.** Keywords include: unique words from the article title, cave names, geographic names, and descriptive terms. The second part is a **biologic names index** begining on **page 141.** These terms are Latin names of organisms discussed in articles. The third part is an alphabetical **author index** starting on **page 142.** Articles with multiple authors are indexed under each author.

Citations are of the following form: names of all authors in the order which they appear in the journal; title of the article or abstract; volume number and part number (separated by a colon); beginning and ending page (separated by a dash); and year of publication from the cover of the issue. Volume number and year are included to match previously published indices for the Bulletin. Within an index group, such as Archaeology, the earliest article is cited first, followed by consecutive articles. **There is one exception to the format of the citations; the title of an article by Peck was abbreviated to "AL fauna" due to the large number of keywords assiciated with it. The complete citation is: Peck,S.B., The cave fauna of Alabama: Part I. The terrestrial invertebrates (excluding insects), 51:1,11-33, 1989.**

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

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Keyword Index

Biologic Names Index

Author Index

Guide To Authors

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

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

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

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

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

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

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

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CONTENTS

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