Thermal genesis of dissolution caves in the Black Hills, South Dakota

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ABSTRACT

Jewel Cave (118 km of mapped passages beneath an area of 2.7 km^2) and Wind Cave (70 km beneath 1.8 km^2) are, respectively, the fourth and tenth longest known cave systems and the world's foremost examples of three-dimensional, rectilinear networks of solutional passages. Other caves in the Black Hills are similar. They occur in 90–140 m of well-bedded Mississippian limestone and dolomite. Walls throughout Jewel Cave are lined with euhedral calcite spar as much as 15 cm thick. Wind Cave displays lesser encrustations and remarkable calcite boxwork. Since 1938, opinion has favored cave excavation by slowly circulating meteoric waters in artesian confinement similar to that surrounding the Black Hills.

We believe that the caves were developed by regional thermal waters focusing on paleospring outlets in outlying sandstones. Four sets of criteria are evaluated: (1) morphological—the three-dimensional, one-phase maze form having convectional features is similar to known and supposed thermal caves in Europe; (2) petrographic and mineralogical study of the chief precipitates shows a record of carbonate solution → calcite precipitation consonant with a model of cooling, then degassing, waters; (3) a thermal anomaly at regional hot springs is shown to extend beneath Wind Cave, where basal lake-water samples show chemical and isotopic affinities with the thermal waters; and (4) $\delta^{13}C$ and $\delta^{18}O$ measurements place all suspected paleothermal water precipitates in the domain of thermal calcites reported by others and being deposited at the modern hot springs. Finally, U-series dates show that the Wind Cave deposits are Quaternary and that the cave is still draining. Jewel Cave is truly relict and divorced from the modern thermal ground-water system; its great calcite spar sheets are probably older than 1.25–1.50 Ma.

INTRODUCTION

Jewel Cave and Wind Cave, in the Pahasapa Limestone of the Black Hills, South Dakota, are the world's fourth and tenth longest known caves, respectively. They are the foremost examples of three-dimensional, rectilinear cavern networks. Each displays, in great abundance, types of calcite precipitates that are rare in caves formed by direct infiltration of meteoric water. Most parts of Jewel Cave are encrusted with coatings of euhedral calcite spar that average 15 cm in thickness. Wind Cave displays a variety of lesser coatings and also remarkable wall and ceiling boxwork of composite solutional-depositional origin. Shorter caverns are known elsewhere in the Black Hills, and with few exceptions, they appear to be fragments of network complexes similar to Jewel and Wind Caves. They contain the same exotic forms of calcite.

The Black Hills caves are composed of two or three levels of solution galleries, most of which are disposed in rectilinear arrays. Passage size varies greatly, with no trend to increase downstream of junctions. The multi-storey characteristic makes them true three-dimensional mazes. All levels appear to have developed simultaneously in the same phase or sequence of phases. This is a rare phenomenon. Two-dimensional rectilinear maze caves (that is, one storey, one phase) are common, being formed where mete-

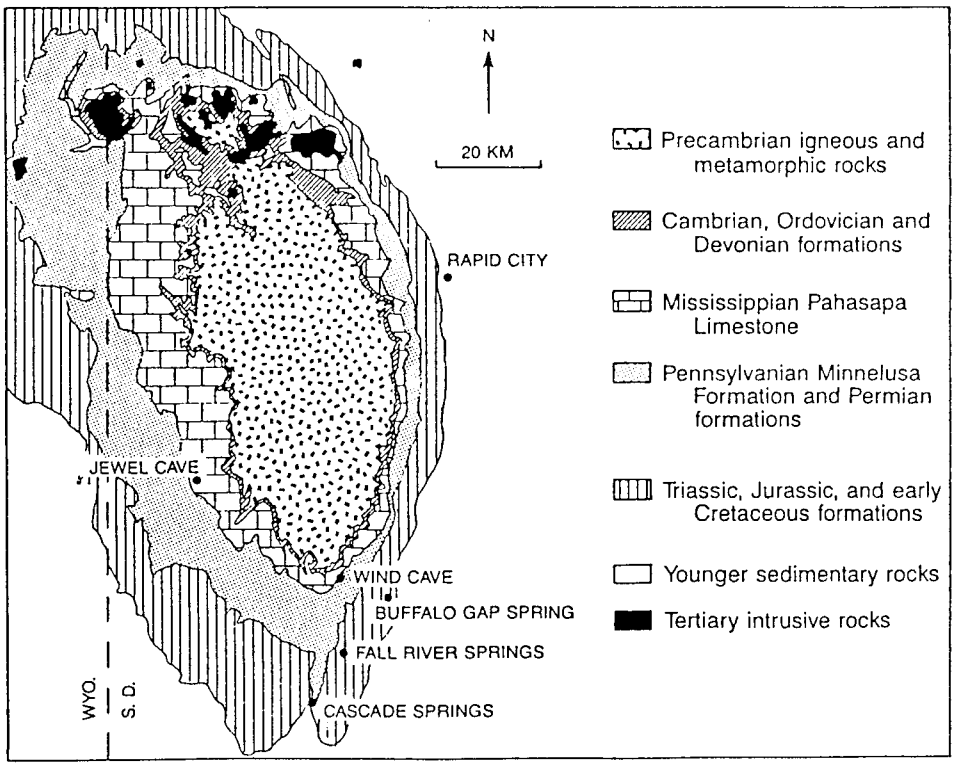

Figure 1. Geologic map of the Black Hills, showing locations of described sites.

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oric waters are guided into a well-jointed and soluble limestone either as uniform infiltration or as periodic flood waters. Multi-level caves with crisscrossing galleries are also common, but only in cases in which each level represents a different phase of development. Normally, these are series of passages developed at successively lower levels, displaying dendritic patterns rather than rectilinear ones; passage size tends to increase systematically downstream of junctions.

The problems posed by the Black Hills caves, therefore, are to explain the development of these multi-level but single-phase solution mazes of exceptional extent and to account for their exotic mineralization. A majority of previous authors have advocated variations of a confined or artesian flow hypothesis with meteoric waters. In this paper, we present evidence that both the dissolution and the mineralization of the caves are the product of rising thermal waters.

Geologic Setting

The Black Hills are a dissected domal structure of Laramide age (Fig. 1). The core consists of rugged mountains of Precambrian igneous and metamorphic rocks that were further intruded by igneous rocks during the early Tertiary. Around the perimeter, there are cuestas of radially dipping Paleozoic and Mesozoic sedimentary rocks, mainly sandstones and shales, breached by a few wind and water gaps. By the end of the Eocene, dissection had advanced close to the modern base levels (Palmer, 1981). Much of the landscape around the perimeter of the Black Hills was then covered by extensive terrigenous sediments of the Oligocene White River Group. There has been renewed uplift and dissection in the later Tertiary and Quaternary.

The lowest sedimentary rocks are 20–70 m of Cambrian to Mississippian sandstones, shales, and arenaceous limestones, resting unconformably on the Precambrian basement (Fig. 2). They are succeeded conformably by the Pahasapa Formation, a platform carbonate of Mississippian age. It is 90–140 m thick in the vicinity of the caves. The lower Pahasapa is massive, limy dolomite with prominent joints, favoring a simple fissure form of passage. Middle strata, in which the principal boxwork is found, include medium-bedded limestones and dolomites, locally highly fractured and brecciated, with some prominent chert beds near the top. Passages are less regular in form, with lower ceiling heights. Upper strata are massive limestones with sparse chert nodules. Passages are well rounded.

The top of the Pahasapa Limestone is a Mississippian paleokarst that has a preserved relief of ~50 m buried by Pennsylvanian sandstones. The paleokarst extends deep into the Pahasapa in the form of filled solutional clefts, sinkholes, and caves. Pennsylvanian detrital fillings vary from collapse breccia to water-laid allochthonous sediments. The modern caves primarily follow later fracture systems but ramify into the palaeokarst cavities, complicating the modern patterns. Reworked pakeokarst fill is a major component of the detrital veneers in the modern caves.

The sandstone cover (100–200 m thick) seals the paleokarst and the caves from overhead penetration by all but diffuse infiltration of meteoric water, except where recent shallow canyons have approached or intersected upper galleries.

Physiography of the Caves

Jewel Cave comprises 118 km of surveyed galleries contained within an area of no more than 2.7 km^2 (Fig. 3). It extends between 1,511 and 1,645 m above sea level in strata dipping a few degrees southwest. The cave is fully drained today, except for a few small perched pools, and is without flowing water. A nearly ubiquitous wall coating of calcite spar is an outstanding feature of this cave (Fig. 4). In the uppermost passages, this encrustation has been partly removed during at least one solutional episode. There is also some local boxwork and a few small occurrences of the travertine (stalactites, stalagmites) typical of most caves. The cave air temperature is ~8.3 °C.

Wind Cave contains 70 km of known passages beneath an area of 1.8 km^2. It extends between 1,120 and 1,265 m above sea level. Its solutional form is very similar to that of Jewel Cave, except that the average cross-sectional dimensions are smaller. Spar coatings on walls are concentrated mainly in the lowest levels and are much thinner than at Jewel Cave. There are many other unusual calcite precipitates, including horizontal fins and false floors that appear to mark growth at former pond surfaces. Normal travertine deposits are rare. The famous "boxwork" of this cave is, in scale, extent, and complexity, probably the finest that has been described. It comprises skeletal structures of vein

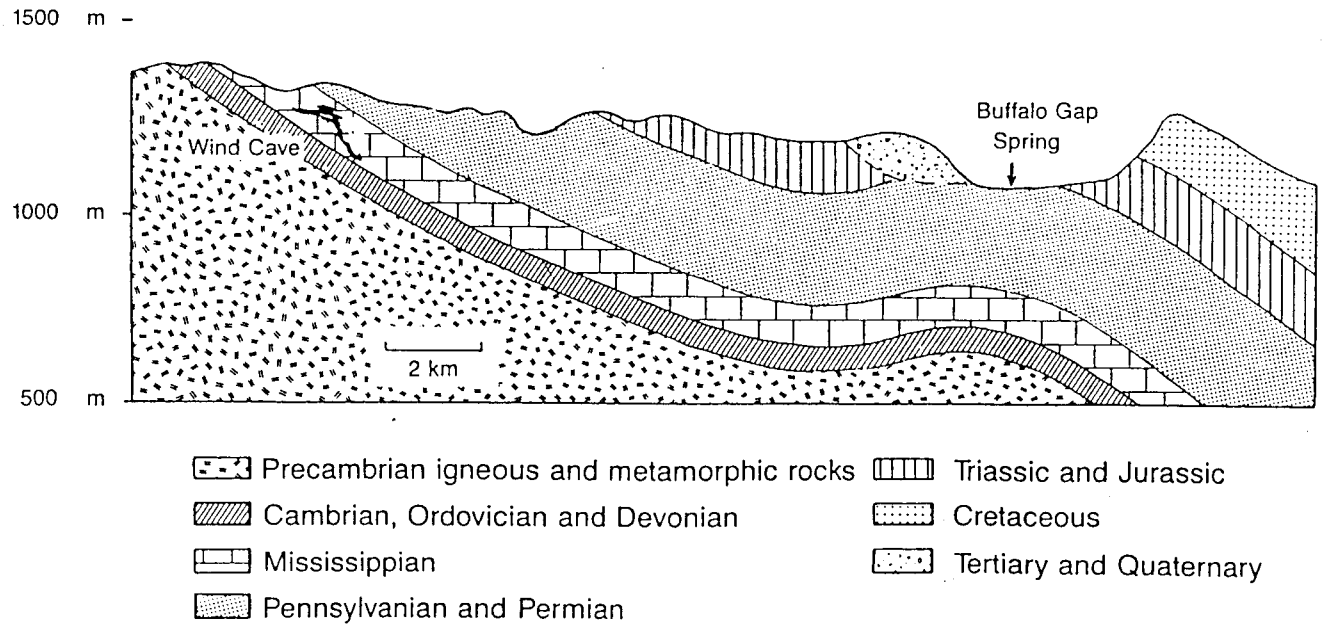

Figure 2. Cross section through the Wind Cave area in the southeastern flank of the Black Hills.

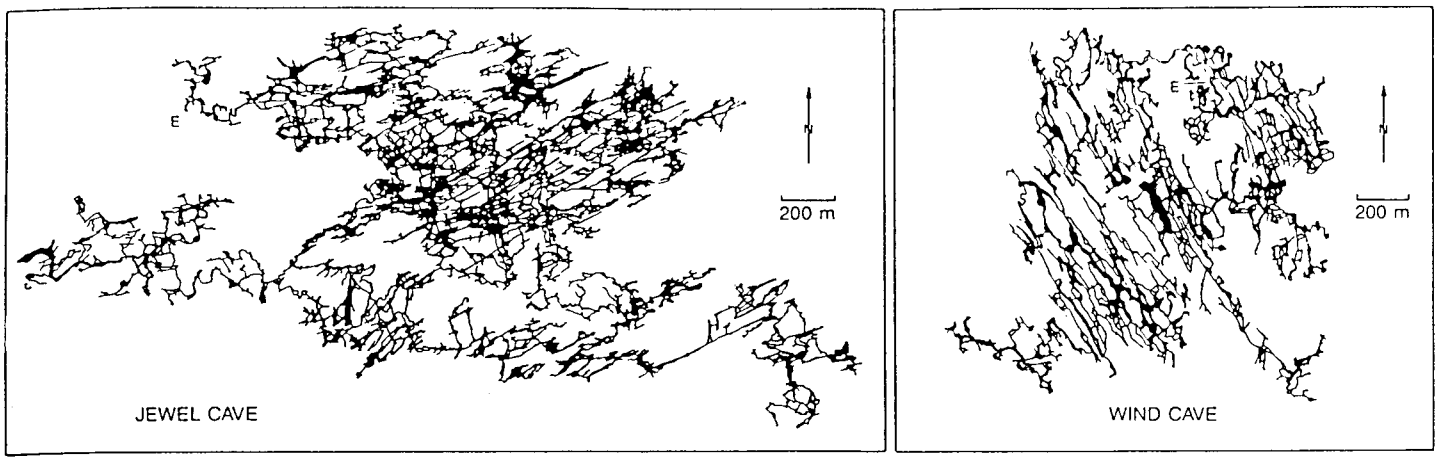

Figure 3. Maps of Jewel Cave and Wind Cave, drawn at identical scales, showing all passages mapped as of 1985. E = natural entrance.

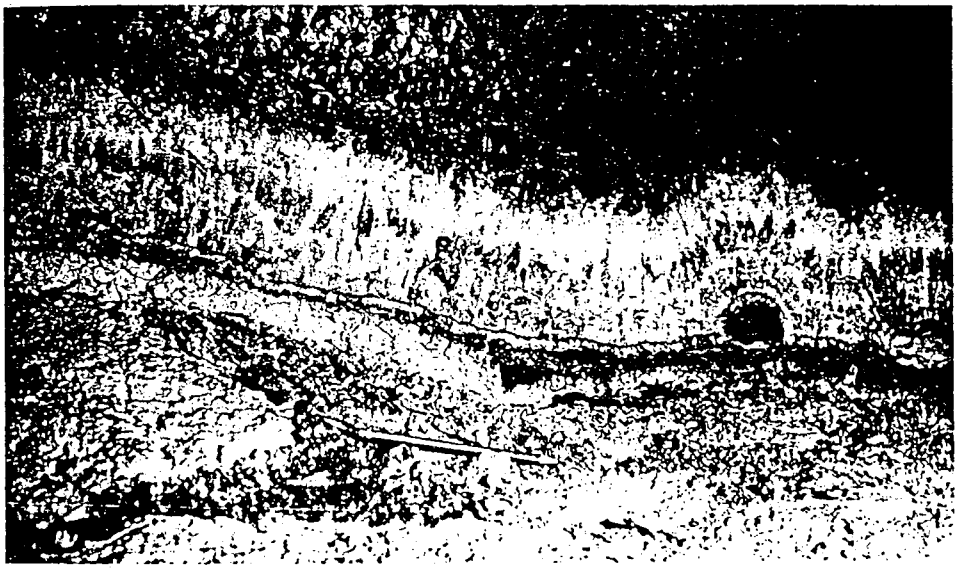

Figure 4. Cross section through calcite wall crust in Jewel Cave, in which it has broken away naturally from underlying silty textured dolomite. Thickness of crust is 15 cm.

calcite with later calcite overgrowths (Fig. 5). The vein calcite has resisted dissolution and protrudes as much as 1 m from walls and ceilings. The cave air temperature is 11 °C in the upper parts, rising to 14 °C in the lowest levels.

Local Hydrology

The climate in the vicinity of the caves is semi-arid, with a mean annual rainfall between 360 and 420 mm. Radially draining streams that rise in the wetter central Black Hills become influent where they cross the sedimentary rocks and lose much of their water to infiltration, particularly into the Pahasapa.

The Pahasapa and its equivalents form a major confined aquifer in the region. Most of the recharge in the vicinity of the caves appears to be through diffuse infiltration, except where perennial or ephemeral streams sink at rare swallow holes. Ground water flows outward from the Black Hills, much of it emerging at springs along anticlines and fault zones a few kilometres farther out from the limestone outcrops. Some ground water continues down gentle hydraulic gradients into the deep basinal areas beneath the surrounding plains.

The modern caves bear no apparent relation to the present surface topography or stream patterns. They are drained, relict features that have been intercepted locally by shallow canyons that carry seasonal runoff. Wind Cave, however, is located along a paleovalley of possible Tertiary age, now largely abandoned by flow (Palmer, 1981). Jewel Cave is beside Hell Canyon, the major drainage line in the southwestern Black Hills, which carries only intermittent flow. Despite their juxtaposition to major river valleys, there is no clear evidence that the caves were formed by water from these sources.

An important feature of Wind Cave is the presence of permanent lakes at its lowest (downdip) end. Similar lakes are not known in the other Black Hills caves, probably because explorers have not yet found them. The Wind Cave lakes are at 1,120 m above sea level, essentially the same as the static water level in wells that penetrate the Pahasapa Limestone farther downdip. This suggests the existence of a very flat piezometric surface within the limestone. That, in turn, implies that there is high permeability (probably in the form of solutional caves) extending farther downdip below the water level. This piezometric surface is 70 m higher than is Buffalo Gap Spring, 8 km east of the lakes in Wind Cave. The elevation difference represents the head required for ground water to flow upward to the spring through the overlying Minnelusa Formation.

The lakes appear to be stagnant backwaters fed from below. Their level has varied seasonally ~1 m since they were discovered 20 yr ago. Calcite rafts are forming upon them, the calcite precipitating onto dust particles settled on the water surface. Raft formation is indicative of great hydraulic stability and the renewal of a supersaturated solution. Raft debris is abundant as much as 30 m above the modern lakes. It is draped over some helictite bushes, which are fragile, subaerial branching calcite speleothems; very slow, steady rise and fall of the water level is implied by this phenomenon.

A final relevant feature is the occurrence of groups of warm springs at Buffalo Gap, Fall River, and Cascade River (Fig. 1). These springs

rise through the Minnelusa Formation where it is exposed as inliers along local anticlinal folds on the dip slope. Water temperatures at points of emergence are 17–26 °C and do not display seasonal variation. Cascade Spring has a mean discharge of 0.6 m^3s^{-1} and is the largest spring of any type in the Black Hills. Most of these springs deposit abundant travertine in their modern channels, which are incised into older alluvial deposits.

The lowest point in Jewel Cave lies 15 m above the local water table. The supposed resurgence for ground water of this area is 37 km to the southeast at Cascade Spring (Rahn and Gries, 1973).

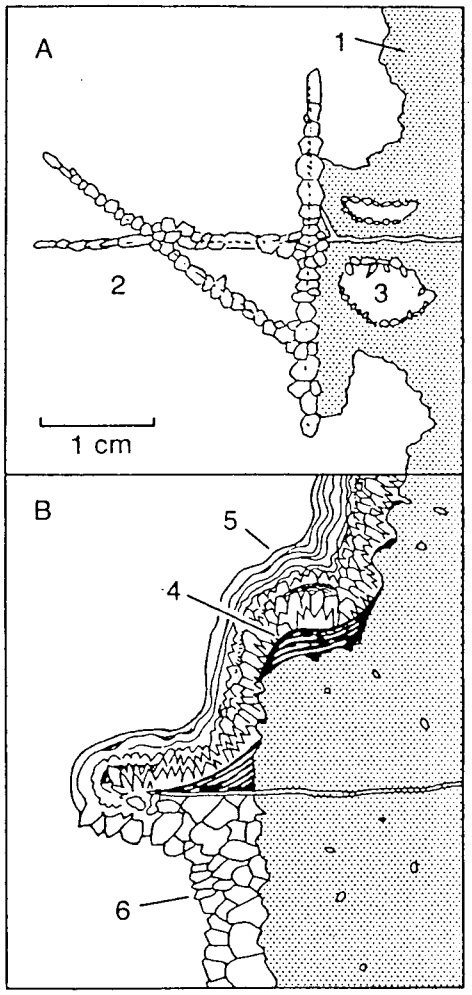

Figure 5. Idealized cross sections through boxwork in Wind Cave. (A) Typical boxwork exposed to weathering in upper passages above the level of calcite wall crust. Most boxwork fins project several tens of centimetres but are attenuated here for clarity. (B) Projecting veins coated with layered calcite wall crust in lower passages. 1 = bedrock (friable in A, competent in B); 2 = boxwork fins, consisting of recrystallized and overgrown pre-cave calcite veins, with ghosts of veins now represented by hematite crystals; 3 = pores in bedrock lined with calcite crystals; 4 = "internal sediment" of detrital carbonate weathered from higher walls; 5 = layered calcite wall crust; 6 = local recrystallized wall crust, the layers of which are faint or absent (in most cases, on undersides of projections).

Previous Work

Davis (1930) attributed the caves to deep-seated solution of the kind associated with hydrothermal ores on the basis of (a) their equidimensional maze characteristic, suggesting dissolution by slowly flowing waters in the phreatic zone, and (b) the spar coatings of Jewel Cave, which resemble the linings in hydrothermal veins.

Later authors accepted the morphologic evidence of dissolution by low-velocity phreatic water but turned away from the thermal interpretation. The earliest hydrogeologic studies (for example, Darton, 1918) established the existence of a regional, artesian aquifer in the Pahasapa Formation beneath the nearby plains in Wyoming and South Dakota. Tullis and Gries (1938) suggested that the caves were excavated by meteoric water circulating slowly through the aquifer during the Eocene-Oligocene, soon after the uplift of the Black Hills. Howard (1964) developed this concept into a more comprehensive model in which vadose and water-table feeder caves developed updip from the artesian mazes. Cave development was not necessarily tied to Eocene-Oligocene events and could have been much more recent. A problem with Howard's proposal is that only isolated and poorly integrated cave fragments survive in the putative feeder areas.

Deal (1962, 1968) published perceptive studies of the mineral suites in Jewel Cave, recognizing no fewer than seven distinct genetic phases: (1) solution by meteoric waters in a confined aquifer, as above; (2) partial or complete drainage of the caves; (3) return to phreatic conditions, with deposition of the principal deposits of nailhead spar; (4) drainage of certain cavities, as indicated by typical vadose deposits; (5) further complete inundation accompanied by widespread dissolution of nailhead spar in the higher parts of the caves; (6) progressive drainage, with deposition of travertine in the upper cave and of mud in the lower parts; and (7) the modern phase, in which minor travertine deposition continues. The final drainage of the known cave has long been complete; it is a hydrologic relict.

We agree with Deal that the history of dissolution with mineral deposition in these caves has been complicated rather than simple. Two of us (Bakalowicz and Ford) question the strength of the evidence for a vadose phase 2, and clearly, a great problem of Deal's sequence is the integration of an apparent hot-water inundation (stage 3) into what is treated otherwise as alternate filling and emptying of normal meteoric water. White and Deike (1962) used geochemical and mineralogical criteria to suggest pressures of 10–100 atm and temperatures of 150–200 °C during stage 3. White (1982, personal commun.), however, has since accepted that these criteria may be irrelevant and that the minerals in question may have been deposited at much lower temperatures.

Palmer (1975, 1981, 1984) pointed to problems of explaining the caves by a simple confined-flow model. He emphasized that the caves are located in a zone in which flow of water undersaturated with respect to calcite and dolomite is possible to and from the Pahasapa limestones through the overlying Minnelusa sandstone and via flood-water recharge from sinking streams. Examples elsewhere show that network mazes in limestone are commonly formed by either type of recharge.

Wind and Jewel Cave morphology, although unusual, is similar to certain caves of flood-water origin. Palmer (1984) suggested that the major episode of wall coating might be caused by ponding of water in the caves, resulting from the regional Oligocene aggradation. Petrographic evidence suggesting that the caves formed under hydrochemical conditions similar to those of hydrothermal ores, however, has turned Palmer's opinion away from a standard origin (cool water and soil CO_2) for the cave formation.

Presented below are evidences for cave origin and development by ascending thermal water. It is possible that some mixing with cool meteoric waters played a role that is not yet elucidated. The evidence derives from geomorphic features, from A. N. Palmer and M. V. Palmer's petrographic and mineralogical studies, and from isotopic measurements of wall rocks, secondary minerals, and waters by Bakalowicz, Ford, and Miller.

CAVE ORIGIN BY RISING THERMAL WATER

Morphological Evidence

In the Western literature, there is little discussion of modern and relict hydrothermal solution caves. They have been much studied in eastern Europe (Czechoslovakia, Hungary, Poland, and

the Soviet Union), however, where Kunsky (1950), Jakucs (1977), Rudnicki (1978), and Dublyansky (1980) have published English or French summaries. Dublyansky listed five criteria that strongly indicate a hydrothermal origin. One of them—composition of exotic precipitates—is considered later in our paper. The other four are morphological.

1. The cave systems lack a genetic relationship to the surface topography.
2. They are largely or entirely devoid of fluvial sediments.
3. The caves in most cases display a three-dimensional rectilinear maze form guided by major fracture systems and, more rarely, by bedding planes. This is indicative of excavation by slowly flowing ascending waters.
4. The highest parts of the caves may display cupola-form solutional pockets dissolved upward into the ceilings. These pockets appear to be convectional in origin. Their form is in most cases different from that of ceiling and wall solution pockets in meteoric-water caves, being more multi-faceted but lacking deep penetration into a guiding joint.

Jewel and Wind Caves meet all four of these criteria very clearly. The first three are noted in our introductory description. Cupola-form ceiling pockets as much as 10 m in height form "The Loft" and other highest places in Jewel Cave. They are best seen in "The Fairgrounds," stratigraphically the highest part of Wind Cave. They are not well developed lower in these caves.

Petrographic and Mineralogical Evidence

Samples of wall rocks and secondary minerals were taken from the caves and nearby outcrops for analysis with petrographic microscope, X-ray, and scanning electron microscope. Samples were obtained under permit from the National Park Service and consisted chiefly of small, detached fragments. A complex sequence of solution, alteration, deposition, and replacement is revealed in the walls of both caves and will be treated in detail in later papers. Only the main aspects pertinent to cave origin are described herein.

The Pahasapa Limestone was subjected to continental weathering late in the Mississippian Period, as noted. In addition to karst forms filled with Pennsylvanian clastics, the carbonate bedrock contains highly fractured and brecciated zones. Wedging features in the breccia indicate an origin due to crystallization and later solution of sulfates. Fractures and breccia interstices then were filled with hematite-rich calcite, as is common during dolomitization. This calcite comprises many of the boxwork veins; most of them are ~100 μm thick, although in breccia zones, some reach several centimetres. The veins show at least two orders of crosscutting that represent different episodes of fracturing. Most are truncated by the solutional paleokarst features, although some extend upward into Pennsylvanian rocks.

The major phase of cave development occurred after the Laramide uplift of the Black Hills. Limestone and dolomite were at first dissolved at approximately equal rates, as shown by somewhat uniform passage enlargement in different lithologies, but during the late stages, dolomite was selectively removed by water that was apparently close to saturation with respect to calcite. Solution of dolomite rhombs created porosity as high as 90% in the cave walls and exposed the calcite veins as resistant fins (Fig. 5). Although the calcite veins are much older than the cave, their exposure as boxwork is the result of cave origin by slow-moving, nearly saturated water quite different from that in karst areas fed by normal surface infiltration.

Iron-rich silica replaced much of the remaining calcite in the porous bedrock. X-ray analysis shows it to vary from opal to microcrystalline quartz. The latter is not uncommon in meteoric-water caves if there are sources of silica. It is an evaporite and thus is limited to frequently wetted patches of rock. In Jewel and Wind Caves, the silica is rather uniformly distributed. This suggests subaqueous deposition, which requires a decrease in either pH or temperature. A small amount of the silica forms meniscus cement, indicating vadose conditions. This may be reworked.

Precipitation of the great calcite spar coatings succeeded silica deposition in Jewel Cave. These crusts average 15 cm thick and contain as many as 20 distinct growth layers (Fig. 4). There are no hiatuses or erosion surfaces between layers. They appear to be cyclic phenomena.

In Wind Cave, the calcite crusts occur as overgrowths on the protruding boxwork veins in the dolomitic middle strata (Fig. 5) and as more general wall cover in the lower cave. They average only a few millimetres in thickness. There are also some pool rim deposits associated with them in the lower cave.

These crusts are subaqueous deposits from water brought to supersaturation either by degassing of CO_2 into air-filled upper caves or by heating. Degassing evidently occurred in Wind Cave. Warming (that was perhaps cyclical) appears necessary to account for the great extent and volume of the encrustation in Jewel Cave.

Modern Geothermal and Hydrochemical Features

Rahn and Gries (1973) studied present geothermal conditions in the Black Hills, including the chemical character of the hot-spring waters. In January 1982, we sampled the hot springs, artesian water, and the different types of water in the caves and obtained the results shown in Table 1. The hot-springs data are essentially identical to those of Rahn and Gries. Cave waters gave results very like those from a larger sampling in Wind Cave by Miller (1979).

TABLE 1. SUMMARY OF SAMPLE WATERS COLLECTED JANUARY 29 AND 31, 1982

	°C	pH	Ca^{2+} (mM/l)	Mg^{2+} (mM/l)	HCO_3^- (mM/l)	SO_4^{2-} (mM/l)	P_{CO_2} 10^{-2}atm	SI calcite	SI dol.	SI gypsum	$\delta^{18}O$ SMOW
Hot Spring	26.7	7.01	2.75	1.03	4.12	6.66	2.5	-0.14	-0.50	-1.28	-16.0
Hot Brook Spring*	17.2	7.80	1.76	0.94	4.20	0.36	0.35	0.39	0.65	-2.04	-14.8
Higher Cascade Spring	20.9	6.75	14.70	2.75	3.80	13.12	3.1	0.04	-0.52	-0.08	-15.4
Lower Cascade Spring	20.4	6.78	14.15	3.03	4.04	20.84	4.1	-0.14	-0.83	0.04	-15.1
Buffalo Gap Spring*	16.9	7.18	13.25	2.10	3.90	11.25	1.5	0.28	-0.15	-0.14	-14.3
Artesian well near Buffalo Gap†	19.0	7.60	1.09	0.49	3.20	0.03	0.56	-0.12	-0.49	-3.19	-11.6
Drip waters in Jewel Cave	9.2	8.30	0.83	1.42	3.86	0.52	0.09	0.45	1.11	-2.13	-12.7
from modern	8.9	8.30	0.85	1.73	4.52	0.13	0.10	0.53	1.34	-2.71	-12.6
speleothems	8.3	8.40	0.65	2.15	5.00	0.32	0.08	0.53	1.53	-2.47	-12.3
Drip water in Wind Cave	9.6	8.07	0.94	0.69	2.92	0.24	0.13	0.14	0.14	-2.36	-11.8
"Calcite Lake" in Wind Cave	12.9	7.93	0.92	0.77	3.49	. .	0.22	0.17	0.29	. .	-12.1
"Windy City Lake" in Wind Cave	14.1	7.90	0.82	0.67	3.00	. .	0.19	0.07	0.08	. .	12.3

*Sampled downstream of spring point.
†Well penetrates Pahasapa Formation.

Rahn and Gries (1973) showed that the Black Hills are characterized by two thermal anomalies. The first is regional, a slightly higher ground-water temperature around the perimeter of the Black Hills uplift. The second is the more sharply defined high-temperature zone around the hot springs. In our data, an artesian well penetrating through the Minnelusa Formation into the Pahasapa Limestone at 2 km from Buffalo Gap hot spring yields a geothermal gradient of 5 °C/100 m. Air and water temperatures at Wind Cave, 7 km from these springs, show a gradient of 3.7 °C/100 m, also rather high. The thermal anomaly observed at the hot springs thus extends beneath modern Wind Cave. There are no comparable data for Jewel Cave.

Schoeller (1962) defined a thermal water as one for which the mean temperature is at least 4° (Celcius) higher than the mean annual surface temperature at the spring. In the southern Black Hills, this implies temperatures above 15–16 °C. Table 1 shows that the hot springs are only feebly thermal but highly mineralized, especially in SO_4^{2-}. Hot Brook Spring (a tributary to Fall River) and Buffalo Gap Spring appear anomalous because they could not be sampled at their bedrock outlets but were sampled downstream after some chemical evolution in the open air in cold weather. The other springs are weakly undersaturated or at equilibrium with respect to calcite and are undersaturated with respect to dolomite. Their calculated P_{CO_2} shows high values, $1.5–4.1 \times 10^{-2}$ atm.

Drip waters in the caves unquestionably represent meteoric infiltration. They are marked by high pH and high Mg^{2+} but very little SO_4^{2-}. They are clearly supersaturated with respect to calcite and are presently depositing stalactites, but they have a relatively low P_{CO_2} ($1–2 \times 10^{-3}$ atm). The artesian well water from the Pahasapa Limestone, at a depth of 200 m, is chemically most like the drip waters ($P_{CO_2} = 6 \times 10^{-3}$ atm) but is warmed to within the thermal range of the hot-springs anomaly (19 °C). The lake waters of Wind Cave display characteristics intermediate between the drips and the artesian sample. They are best interpreted as local artesian waters that have cooled and degassed in the cave.

Stable Isotope Evidence

Water. Meteoric waters dripping into Jewel and Wind Caves have average $\delta^{18}O$ values of $-12.5‰$ and $-12.1‰$, respectively, with respect to SMOW. Yonge and others (1986) have shown that $\delta^{18}O$ of cave drip waters is equal to that of the average annual precipitation in the recharge area. The Jewel and Wind values agree well with the local precipitation values given in Yurtsever and Gat (1981).

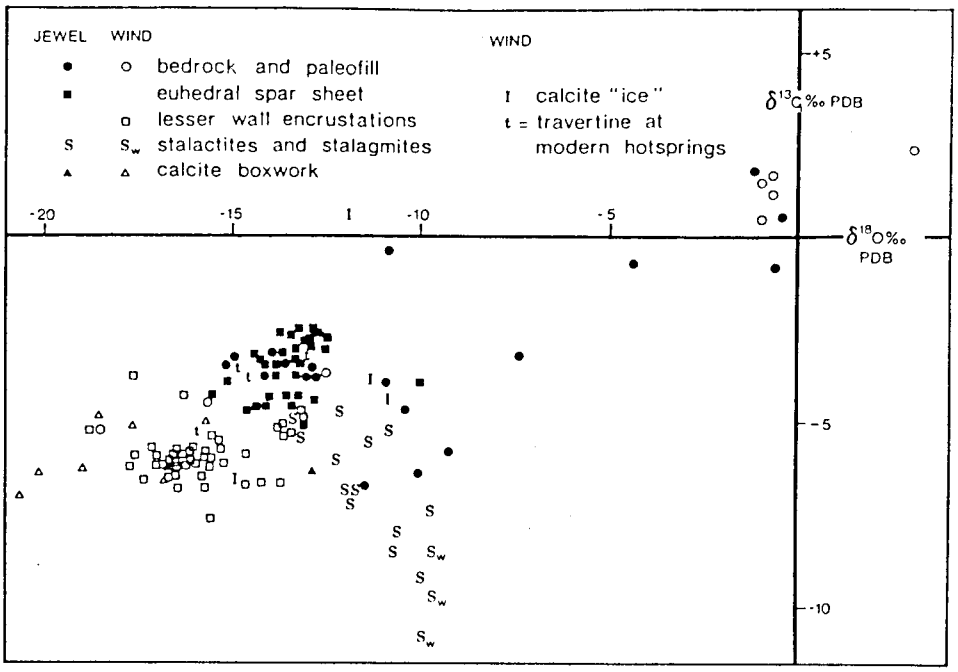

Figure 6. $\delta^{13}C$ and $\delta^{18}O$ per mil values wrt PDB for wall rock, suspected thermal calcites, and normal (meteoric water) speleothems from Jewel and Wind Caves, plus recent and modern hot-springs travertines for Hot Brook and Cascade River, South Dakota.

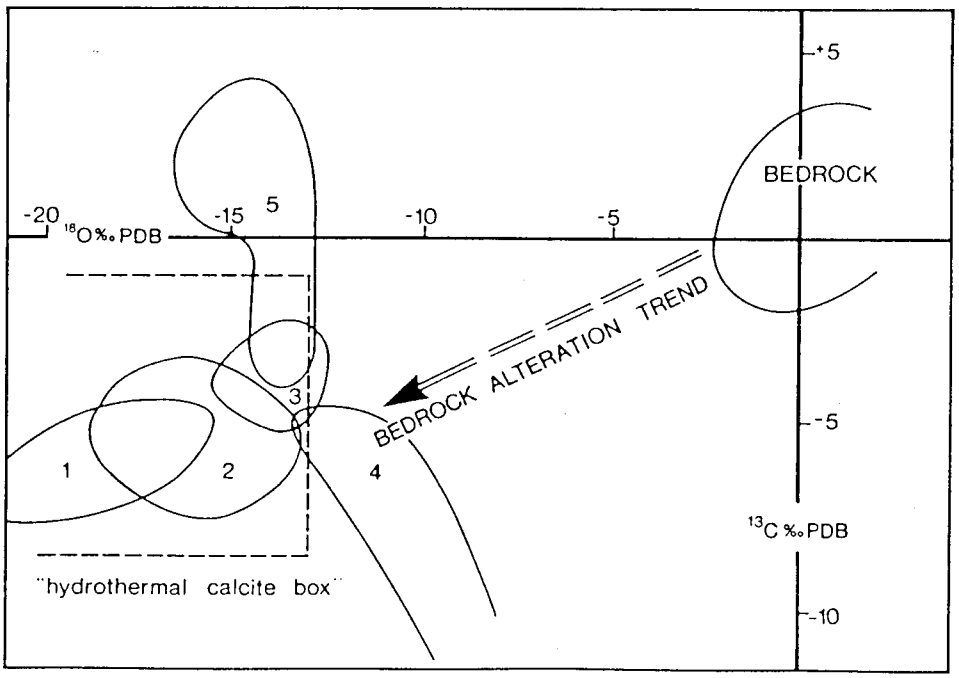

Figure 7. Interpretation of the data plotted in Figure 6, plus data from paleo-hotsprings caves at Budapest, Hungary. Envelope 1 contains all boxwork samples from Wind Cave; 2, all suspected thermal calcite crusts in lower Wind Cave; 3, all euhedral spars from Jewel Cave; 4, normal stalactites and stalagmites from both caves; 5 is the envelope for subaqueous and pool-rim deposits sampled in relict hot-springs caves of Budapest.

TABLE 2. TESTING FOR EQUILIBRIUM OR KINETIC ISOTOPE FRACTIONATION IN JEWEL AND WIND CAVE CALCITES

Site; type	Sample no.	‰ versus PDB $\delta^{13}C$	‰ versus PDB $\delta^{18}O$	Equilibrium precipitation?
Jewel Cave; calcite crusts of supposed hydrothermal origin	JC 0-4	-2.55	-13.52	
	0-5	-2.94	-13.83	Yes
	JC 2-1	-2.60	-13.91	
	2-2	-2.80	-13.92	
	2-5	-2.67	-13.91	Yes
Wind Cave; calcite crusts of supposed hydrothermal origin	WC 0-1	-6.82	-16.54	
	0-2	-6.14	-16.04	
	0-3	-6.55	-16.24	Yes
	WC 8-6	-4.72	-13.76	
	8-7	-4.89	-13.33	Yes
	WC 36-1	-5.88	-16.17	
	36-2	-5.90	-15.92	
	36-4	-5.94	-15.66	Yes
	WC 37-2	-6.52	-17.29	
	37-4	-6.15	-16.95	Yes
	WC 38-1	-6.12	-16.71	
	38-2	-6.36	-16.48	
	38-3	-6.08	-16.23	Yes
Wind Cave; modern calcite "ice"	WC 21	-4.45	-10.94	^{13}C, no ^{18}O, yes
Jewel Cave; normal stalactite	JC 11/3-1	-8.13	-10.58	
	11/3-2	-6.12	-12.19	No

Note: JC 2-1, 2-2, 2-5, and so on = samples measured at fixed intervals along one travertine growth layer.

The hot-springs waters display average $\delta^{18}O$ of $-15.1‰$, that is, $3‰$ lighter than the drip waters. The possibility that this depletion is due to discharge of "fossil" Pleistocene water from cooler climatic phases may be ruled out on quantitative grounds. The depletion must be attributed to isotopic exchange with depleted rocks.

Calcite. To investigate carbon and oxygen isotopic characteristics, 75 samples were collected and 150 analyses made by mass spectrometry. Samples included cave wall rock (ranging from fresh to highly weathered), paleokarst fills, euhedral spars, boxwork and other exotic coatings and modern (meteoric water) speleothems in the caves, plus travertine and water from the hot springs. Results are displayed in Figure 6 and interpreted in Figure 7. Seven samples of the calcites believed to be of hydrothermal origin were selected at random to test for isotopic fractionation. In terms of the criterion of Hendy (1971), all of these were deposited in isotopic equilibrium with the source water (Table 2). One ordinary stalactite that was tested was not in equilibrium.

In ordinary speleothems (that is, deposited from meteoric waters), $\delta^{13}C$ values are about $-11 \pm 2‰$ PDB. In effect, the carbon is a one-to-one mixture between the carbonate rock ($\delta^{13}C = 0 \pm 3‰$ PDB) and soil CO_2 ($\delta^{13}C = -22 \pm 5‰$ PDB). More positive values of $\delta^{13}C$ (as in all suspected hydrothermal precipitates shown in Fig. 6) are explained by precipitation from HCO_3^- which is enriched in ^{13}C with respect to such a mixture. Such CO_2 could be derived from a magmatic source or by high-temperature decarbonation of limestone (Truesdell and Hulston, 1980).

The equilibrium ^{18}O concentration in speleothems is determined principally by the concentration in the source water and the water temperature at time of calcite precipitation. In Figure 6, it is shown that with few exceptions, all suspected hydrothermal calcite deposits in Jewel and Wind Caves plot within the domain of hydrothermal calcites (Fig. 7), as shown by Friedman (1970), Robinson (1975), Barnes (1979), and Hoefs (1980). Travertines from the Black Hills hot springs are also in this domain. Three samples of modern "calcite ice" precipitating on the surface of a lake in the bottom of Wind Cave have $\delta^{18}O$ values in equilibrium with the modern temperature there but are enriched in ^{13}C if compared to normal speleothems in the caves. By contrast, an ancient "ice" sample 30 m above the lakes has more characteristically hydrothermal isotope ratios ($-6.64‰$ $\delta^{13}C$ and $-14.9‰$ $\delta^{18}O$ PDB). One small and aberrant sample of spar from Jewel Cave ($-10.0‰$ $\delta^{18}O$) is now believed to have derived from the paleokarst fill.

The great encrustations of nailhead spar in Jewel Cave display little isotopic variation; their mean value is $\delta^{18}O = -13.8 \pm 0.7$ (1 σ) PDB. If the temperature relationship proposed by O'Neil and others (1969) is used, their depositional temperatures were probably in the range 15 to 35 °C. The mean value of boxwork in Wind Cave is $\delta^{18}O = -18.1 \pm 1.6$ PDB. This corresponds to a temperature range of 30 to 60 °C. Wind Cave boxwork has precisely the same isotopic range as does the hot-springs calcite in Yellowstone Park reported by Truesdell and Hulston (1980).

Modern dripstone and flowstone speleothems that have been deposited by meteoric waters infiltrating into the caves are generally quite distinct. Some Jewel Cave samples are unusually enriched in ^{13}C and depleted in ^{18}O. These may be disequilibrium deposits, as in the example given in Table 2. Alternatively, their feed waters may flow over or through the spar sheets as well as through isotopically depleted country rock, exchanging with these depleted rocks. Wind Cave speleothems are not depleted in ^{18}O; there are no great barriers of spar along the courses of their feed waters. Samples of the wall rocks are also shifted to lower $\delta^{18}O$ and $\delta^{13}C$ values, presumably by exchange and some recrystallization in the same hydrothermal waters.

For comparison, Ford collected samples of spar, lesser crusts, and pool rim deposits from caves at Budapest that undoubtedly are of hydrothermal origin. They display the same depletion in ^{18}O as do the suspected hydrothermal calcites in Jewel and Wind Caves (Fig. 7). Some Budapest samples are notably enriched in ^{13}C. This is probably due to local, high-temperature metamorphism of limestone along some master joints during a Miocene volcanic phase that preceded cave genesis there (Muller, 1987). Budapest cave wall-rock samples also display a strong complementary alteration trend.

In summary, conditions similar to those measured at the Black Hills hot springs today ($\delta^{18}O$ of waters in the range $-14‰$ to $-16‰$ SMOW and temperatures of 20 to 40 °C) will readily explain the isotopic composition of most of the exotic precipitates sampled in the caves.

Uranium Series Dating of Cave Calcites

The caves are devoid of flowing water today and are therefore hydrologic relicts. We thus cannot measure directly the conditions that created them. It is possible that they were formed as early as the late Eocene–Oligocene or in the mid-Tertiary and thus in origin, might be fully divorced from any current geohydrologic or geothermal conditions, although the contrary is implied by some of the hydrochemical and stable-isotope evidence already discussed.

Tables 3 and 4 present U-series dates for the cave and hot-spring deposits. The modern rate of deposition of travertine at the hot springs appears to be very rapid; samples are difficult to date with great precision because of detrital thorium contamination, a problem that is encountered in many subaerial tufas. Sample CRO 1 (which was taken from an extensive terrace that is now distant from the modern hot-spring channel of Cascade River), however, can be only a few thousand years in age.

Passages in Wind Cave rise to a maximum height of ~145 m above the basal lakes. Sample WC+ was a small nailhead spar encrustation

TABLE 3. $^{230}Th/^{234}U$ AGES OF SECONDARY CALCITE SAMPLES IN WIND CAVE AND AT CASCADE RIVER

Sample no.	Description	U (ppm)	$^{234}U/^{238}U$	$^{234}U/^{238}U_0$	$^{230}Th/^{232}Th$	$^{230}Th/^{234}U$	Age (Ka)
WC +	Nailhead spar, central cave; ~1,200 m asl	3.57	1.568	. .	33	1.099	>350
WC 7	Wall crust, Boxwork Pit; ~1,180 m asl	2.17	1.587	2.157	540	0.982	242 ± 16
WC 9	Vadose flowstone floor, Boxwork Pit; ~1,180 m asl	3.17	1.580	2.035	134	0.923	205 ± 14
WC 5	Rafts of "calcite ice" draped upon WC 7 and WC 9	3.579	1.663	2.114	60	0.890	185 ± 12
WC 6	Cornice grown upon WC 7; probable pool surface deposit	7.86	1.122	1.189	110	0.775	154 ± 2
WC 10	Wall crust, Rescue Pit; ~1,155 m asl	2.67	1.670	2.284	316	0.974	230 ± 20
WC 12	Cornice, Rescue Pit	0.532	1.538	1.665	7	0.571	76 ± 5*
WC 20	Wall crust; ~1,145 m asl	2.62	1.585	2.101	210	0.957	225 ± 15
WC 14	Wall crust; ~1,135 m asl	1.71	1.736	2.421	47	0.984	234 ± 20
WC 19	Fine-grained calcite crust 10 m above Calcite Lake; shows re-solution	1.51	1.514	1.922	24	0.931	208 ± 23
WC 18	Coarse-grained calcite crust 10 m above Calcite Lake; shows re-solution	1.14	1.459	1.706	350	0.802	153 ± 18
WC 16	"Calcite ice" draped 25 m above lakes	1.62	1.600	1.600	1.3	0.014	1.6 ± 0.4*
WC 17	"Calcite ice" draped 14 m above lakes	1.29	1.773	1.773	5	0.002	0 ± 0.1*
81062-2	"Calcite ice" draped 5 m above lakes	2.19	1.627	1.627	1.5	0.016	0 ± 0.3*
WC 21	"Calcite ice" floating on Calcite Lake; 1,120 m asl	1.41	1.810	1.817	10	0.032	3 ± 0.2*
CRO 1	Oldest tufa terrace below Cascade River hot springs	1.04	1.903	1.906	2	0.029	3 ± 0.2*

Note: $^{234}U/^{238}U_0$ = calculated ratio of these two species at time of co-precipitation in the calcite.

*Age calculated assuming an initial $^{230}Th/^{232}Th$ ratio of 1.25.

TABLE 4. $^{230}Th/^{234}U$ AGES OF SECONDARY CALCITE SAMPLES IN JEWEL CAVE

Sample no.	Description	U (ppm)	$^{234}U/^{238}U$	$^{234}U/^{238}U_0$	$^{230}Th/^{232}Th$	$^{230}Th/^{234}U$	Age (Ka)
JC 1*	Nailhead spar crust, 12 cm thick; whole-rock age	0.37	1.010 ± 0.03	. .	11	1.07	>350
JC 1T	Top 2.5 cm of JC 1	0.36	0.922 ± 0.02	. .	158	1.08	>350
JC 1B	Basal 2.5 cm of JC 1	0.45	1.001 ± 0.01	1.001	186	0.912	265 ± 30
JC 3	Nailhead spar 6 cm thick	0.30	0.998 ± 0.02	. .	107	1.62	>350
JC 7A/T	Nailhead spar crust 6 cm thick; top 1.0 cm	0.17	0.922 ± 0.11	. .	86	1.04	>350
JC 7A/M	As above, central 1 cm	0.33	0.972 ± 0.02	. .	170	1.02	>350
JC 7A/B	As above, basal 1 cm	0.49	0.980 ± 0.02	. .	780	1.02	>350
JC 7B	Nailhead spar sheet 2-5 cm thick, below JC 7A and separated from it by a thin red and black silty layer	0.18	0.986 ± 0.04	. .	130	1.64	>350
RL 18A	Nailhead spar from lowest part of cave; A = base of spar crust	0.28	1.011 ± 0.12	. .	123	0.99	>350
RL 18B	As above; B = top of spar crust	0.18	1.047 ± 0.10	. .	15	1.04	>350
JC 11M	Stalactite drapery (vadose) growing over spar, now fallen from wall; 4-5 cm above base	7.47	1.078 ± 0.01	1.119 ± 0.02	690	0.766	153 ± 18
JC 11T	Stratigraphic top of JC 11; 9-12 cm above base	2.97	1.173 ± 0.02	1.233 ± 0.02	76	0.64	106 ± 6
JC 11TR	Replicate of JC 11T	1.72	1.170 ± 0.02	1.226 ± 0.02	87	0.62	102 ± 7

*This sample displayed normal magnetic polarity; signal very weak.

taken ~80 m above the lakes (Table 3). It indicates that the cave was filled with water to or above this level before 350 Ka. ^{234}U and ^{238}U are far from equilibrium with each other (compare with the Jewel Cave spar samples, Table 4), indicating that the sample is probably much younger than 1.25 Ma.

The remaining samples were collected between the lakes (water table) and Boxwork Pit, 60 m above. Samples WC 7, 10, 20, 14, and 19 are of thin, discontinuous calcite wall crusts occurring throughout this height range. They indicate that between, broadly, 200 and 250 Ka, this lower region of the cave was water filled and experienced slow deposition of calcite everywhere. Crust deposition continued, in the lowest places at least, until ~150 Ka (WC 18).

Between ~200 and 150 Ka, the water table appears to have stood close to the bottom of Boxwork Pit, with some oscillation through a range of several metres or more. This is illustrated by the excellent stratigraphic sequence of dates for phreatic crusts, pool rimstones, and "calcite ice" shown in Figure 8. The dynamic hydrologic conditions implied by this figure appear to be the same as those that now occur at the modern lakes 60 m below.

Abundant deposits of "calcite ice" accreted to dust particles on the lake surfaces and became stranded as the lakes withdrew. The dust nuclei create serious detrital thorium problems for dating, but preliminary results suggest that the pattern of slow lake-level fluctuations superim-

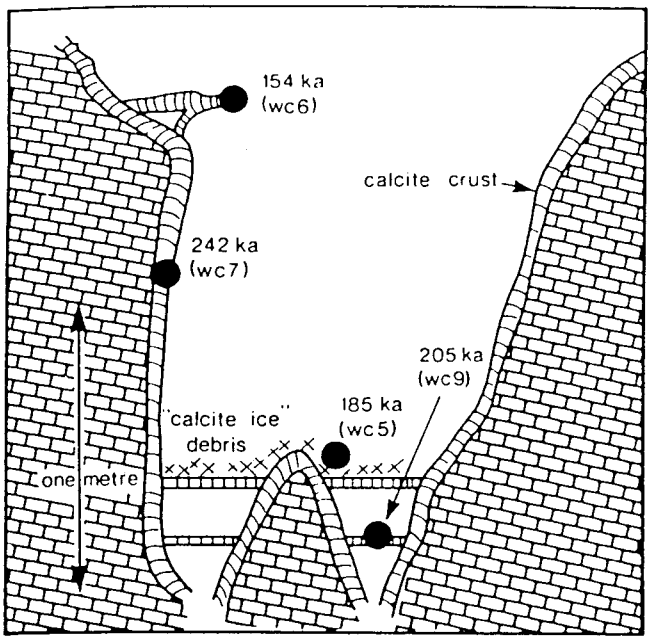

Figure 8. $^{230}Th/^{234}U$ ages of the general wall crust, a former flowstone floor, a waterline fin, and lake "ice" of calcite at the bottom of Boxwork Pit, Wind Cave. This is from a field sketch; the scale is only approximate. See the text for discussion.

posed onto a longer term lowering are continuing (samples WC 16, 17, 21, and 81062-2, Table 3).

These U-series results for Wind Cave are highly consistent. They show that at least the lower half of the cave has been in a phreatic state with warm-water calcite precipitation during the past 250,000 yr or less. Most or all of its solutional excavation could have occurred immediately before crust deposition. The lowest one-third of the explored cave (lying below the main boxwork zone) has drained as a backwater since ~200 Ka and continues to do so today. A more comprehensive dating study is now in progress to elucidate the details there.

Jewel Cave stands 400 m higher in elevation than does Wind Cave and is much farther from the modern hot springs. It is to be expected that it has been relict for a longer period, even in its lowest parts. This is clearly borne out by U-series analyses of the great spar sheets. Our sampling included spars from high, intermediate, and low sites in the cave. If sample JC1B is set aside as aberrant (Table 4), all specimens are older than 350 Ka in age, the limit of the $^{230}Th/^{234}U$ method. The ratio $^{234}U/^{238}U$ is less than or equal to unity. This is in marked contrast to Wind Cave. We have no means of estimating the initial $^{234}U/^{238}U$ ratio in spar sheets at Jewel Cave. The lowest modern ratio is 0.922, recorded in the stratigraphic tops of samples JC1 and JC7A. If it is assumed that this ratio corresponds to an age of 350 Ka (it cannot be younger), then a ratio of 1.00 ± .01 would represent a minimum age of 830 Ka. The bases of the Jewel Cave spar sheets are most probably older than 1.0 Ma in age.

The remanent magnetism of sample JC1 was measured. The polarity was normal, but the signals were very weak; no age inferences can be drawn from the data.

As noted, there are very few normal (meteoric water) calcite speleothems in these caves. At Jewel Cave, one of the most massive (and presumably older?) examples had fallen and shattered. Samples taken from its stratigraphic middle and upper parts are 150–100 Ka.

At Jewel Cave, the spar sheets are ancient, and their deposition could have been completed before 1.0 Ma. Much of the cave may have been drained at that time. If that is the case, there has been remarkably little development of overhead infiltration routes into these empty voids during the past 1 m.y. because even comparatively massive stalactites are quite young.

DISCUSSION AND CONCLUSIONS

The evidence we have presented suggests that the large network caves of the Black Hills were formed by CO_2-rich waters that were heated and that ascended through the Pahasapa Formation. Most or all of the thermal precipitates in Wind Cave are of Quaternary age. It is possible that the dissolutional enlargement of this cave was also limited to the Quaternary, but we suspect that it probably began during the Miocene or Pliocene. Formation of Wind Cave is compatible with the geothermal and hydrogeological conditions that exist in its local region today. Waters rose and converged through what is now the cave zone en route to spring points in an adjoining valley. As a consequence of continued surface entrenchment, spring positions have shifted down dip. The valley is now a dissected relict, and the known cave has become a drained backwater. It retains a strong thermal gradient, and final-stage warm-water precipitates (the "calcite ice") are still forming at the modern water table in it, where exploration is terminated.

Jewel Cave is significantly older and more completely relict. It is not related to the hydrogeological conditions prevailing in its region today. Nevertheless, we suggest that it was formed in the same mode as Wind Cave, by thermal waters rising and converging through it toward spring positions in an earlier level of Hell Canyon. The culminating morphological event in Jewel Cave was the deposition of the calcite spar sheets. They are among the greatest known in any cave. We have shown that they are deposited from thermal waters probably at some time before about 1.0 Ma.

The caves display a broad tendency to descend in stratigraphic elevation in the direction of stratal dip. This suggests that the thermal plumes were rising with an updip component. An igneous heat source within the Precambrian rocks is envisioned, related to the igneous activity occurring elsewhere in the Black Hills throughout much of the Tertiary. Recharge to the systems was probably from infiltration of meteoric waters over wide areas, a pattern of circulation that has been documented in many geothermal systems (Dublyansky, 1980; Ellis and Mahon, 1977). The position of the Pahasapa Formation close to the base of the sedimentary sequence plus the greater elevation of the Black Hills make it unlikely that any significant part of the cave discharge consisted of basinal fluids from strata beneath the surrounding plains.

There were three principal modes of cave development: (1) solution of limestone and dolomite at nearly equal rates by water considerably undersaturated with respect to both carbonates, which mode was quantitatively predominant, (2) selective solution of dolomite only, by water near to saturation with calcite, and (3) deposition of calcite from supersaturated water. In an ideal thermal model, all three stages will occur simultaneously in a vertical sequence. At any fixed P_{CO_2}, the saturation concentration of dissolved carbonate in water increases as the temperature decreases. As the thermal water rises and cools, it acquires or retains solutional aggressiveness with respect to both calcite and dolomite, regardless of the initial dissolved carbonate content. If cooling of water is very gradual, however, the system can hover near the saturation value of calcite and dolomite, preferentially dissolving the species that is more soluble under prevailing geochemical conditions.

Decreasing hydrostatic pressure in the rising water may allow partial degassing of CO_2, which sharply reduces the saturation concentrations and causes precipitation of the secondary carbonates. At most sites observed by Ford in the Budapest thermal caves, precipitation was intense down to 2 m below the paleo–water tables and reduced to zero at depths greater than ~10 m. Rapid degassing in well-ventilated hills best explains such sharp zonation. Wind Cave is not so well ventilated (despite its name); slower degassing probably explains its poorer zonation of precipitates. Simultaneous deposition of the spar crust throughout Jewel Cave can be explained by a phase of warming, inducing degassing within the cave zone, or by a protracted backwater phase marked by very slow circulation and degassing.

This model is highly simplified and must be modified to account for details in the history of the individual caves. Water levels appear to have fluctuated in response to local aggradation at springs or to wetter spells during the later Tertiary and Quaternary, within the over-all lowering induced by regional erosion. As higher springs were abandoned, meteoric flood waters may have penetrated by way of them, contributing to cave enlargement.

Finally, the three-dimensional network pattern of the caves must be explained. Their origin requires a way of distributing the solutional capacity of the water rather uniformly between major joints over particular areas of several square kilometres or more. We suggest that regional and diffuse, heated discharge converged upon what became the cave zones, flowing up dip in the Pahasapa and ascending through the lower formations. Cooling of these waters simultaneously throughout the joint nets produced the crucial solutional aggressiveness. Fluctuating head within the evolving cave zones (in response to varying recharge) and mixing corrosion probably played subordinate roles.

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