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NATIONAL SPELEOLOGICAL SOCIETY

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

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

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

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Biology of the Idaho Lava Tube Beetle, *Glacicavicola*

Stewart B. Peck *

ABSTRACT

Glacicavicola bathyscioides Westcott beetles were kept in laboratory culture. Indirect evidence suggests that the beetles in culture fed on arthropod remains. Of 52 individuals, eight still were alive two years after capture. They may live, in nature, at least 3 years as adults. Immature stages are not known. Some of the caves in which the beetle is presently known to occur probably were occupied less than 2,000 years ago. Their present range, 186 km maximum diameter, was attained during the Wisconsinan glaciation by overland dispersal.

Introduction

The known fauna of lava tube caves consists only of a few specialized animals, contrasting sharply with the rich faunas which have been discovered in limestone caves. Of these few, virtually nothing is known about the general biology and ecology. This note reports aspects of the biology of the unique eyeless beetle *Glacicavicola bathyscioides* Westcott (Coleoptera, Leiodidae). *G. bathyscioides* is known only from cold lava tube caves in Idaho, where it usually is found in association with permanent ice (Fig. 1). The name *bathyscioides* refers to the fact that this beetle bears a striking resemblance to the Bathysciine cave beetles of Europe, even though there is no evidence that these two groups are closely related.

In 1969, I observed the beetles in their natural habitat (Peck, 1970). A total of 52 beetles were captured alive in caves in Craters of the Moon National Monument, Idaho. These were transported to and kept alive in laboratory culture at Harvard University and, later, at Carleton University.

Culture Methods

The beetles were kept in a clear polystyrene plastic box 26 cm wide, 32 cm long, and 11 cm deep with a tight-fitting lid. The bottom of the box was covered with a hardened mixture of equal parts of plaster of paris and powdered wood charcoal. Flat pieces of basaltic lava from the beetles' cave were embedded in the plaster-charcoal substrate; other pieces rested free on the surface. The substrate and the atmosphere above it were kept near saturation by periodic additions of distilled water. The culture box was kept in a refrigerator at 40°F (4.5°C).

Feeding

In nature, the beetles have been observed feeding only on portions of a dead individual of the same species (Westcott, 1968). They have been found aggregating around human dung and decayed meat, possibly for feeding (Peck, 1970). In culture, the beetles were never observed to feed on or to pay any attention to potential food items. Items offered as food were live bakers' yeast, fresh and decayed chicken liver, human dung, freshly killed Phorid and Drosophilid flies, portions of mealworm larvae and pupae, limburger cheese, freshly removed roach heads, Tetramin tropical fish food, and freshly killed terrestrial isopods.

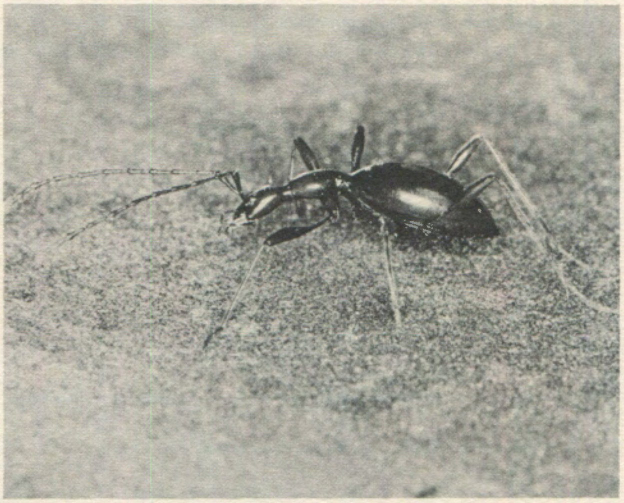

Fig. 1. *Glacicavicola bathyscioides* Westcott from Beauty Cave, Craters of the Moon National Monument, Idaho.

Behavior that could be called exploratory feeding activity was limited to two individual beetles, which were observed to randomly approach a freshly cut fragment of a mealworm larva and place their mouthparts in the haemolymph fluid for four seconds. They then withdrew without palpation or other indications of excitement. Three other beetles walked over the same fragment without stopping or seeming to notice it. There is no other direct evidence of feeding activity. However, indirect evidence of feeding exists. Of the offered potential food, some was moved about the box and dragged under rocks. This happened to the Phorids, Drosophila, roach heads, and isopods. The soft foods (dung, cheese, etc.) were examined for marks of mouthparts, but none were observed. Although not conclusive proof, the above data indicate that the beetles are scavengers on arthropod remains and do not favor other decomposing organic material, such as dung or carrion, for feeding.

Other evidence besides the movement of arthropod remains suggests feeding. Six months after capture, the living beetles (32) were anesthetized with CO_2 and the conditions of their abdomens were observed. Full abdomens, with the tergites fully pressed against the elytra, were found in 14 beetles; half-full abdomens, with tergites only about one half in contact with the elytra, were found in 13 beetles; reduced abdomens, with the tergites entirely free from the elytra, were found in five beetles. Full or partially full

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abdomens in 27 out of 32 beetles a half-year after capture suggests that feeding was maintaining the population at or close to a replete condition. The same conclusion was reached with an examination of the beetles which survived to 24 months after capture. Of these, six had maintained a full abdomen.

Longevity

Figure two is a plot of the survival of the original 52 beetles. Seventeen died in the first month, during transport in an ice chest from the cave to the laboratory in Massachusetts. Thereafter, a fairly uniform death rate is encountered. Of the 35 that survived transport to the laboratory, eight (25%) were still alive 23 months later, at which time they were killed. Since the age structure of the initial sample was unknown, a maximum and a mean longevity cannot be determined. An extrapolation of the death rate curve into 1973 suggests that the beetles may live three years or more at their natural cave temperature (which ranges from at least 0° to 4°C). A longer life is indicated if the winter temperatures of the caves fall lower and the beetles' metabolism is thereby further decreased.

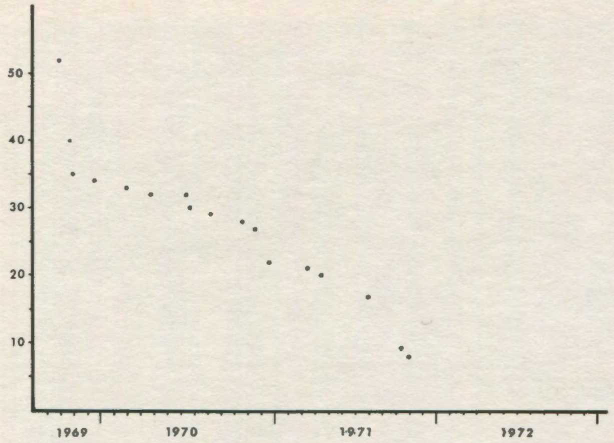

Fig. 2. Population decline in *G. bathyscioides* in culture.

Reproduction

The beetles originally were kept in culture for the principal reason that, if life cycle and larval characteristics could be obtained, these might help to solve the problems of the taxonomic relationship of the beetle. During the 24 months that the beetles were held in captivity, no reproductive behavior and no immature life cycle stages were observed. The culture contained a mixture of both sexes, although females were three times more abundant than males. Successful culture of the beetles may be possible only in one of their native caves.

Due to the lack of larvae, their characteristics cannot be used in solving the problem of relationship. Westcott (1968) erected a new subfamily to hold the genus within the family Leiodidae (sensu latu). This classification still seems best. Although Barr (1968, p. 81) placed this beetle in the family Brathinidae, this possibility now seems very slight as a result of the review of *Brathinus* by Hammond (1971) and of his placement of *Brathinus* in the staphylinid subfamily Omaliinae.

Evolution and Dispersal

Prinz (1970) reported a radio-carbon date of 2,130±130 years BP based on sagebrush charcoal found under the Kings Bowl lava flow. This flow contains Crystal Ice Cave, in which *G. bathyscioides* occurs. The flows (hence, the caves) (Boy Scout and Beauty caves) to the north in Craters of the Moon National Monument are comparable in age to the Kings Bowl flow, judging from their similar degrees of weathering and stages of vegetational succession. A tree ring date of 1,650 years BP in the Monument provided a minimum date for one of the flows. It can be concluded from these dates that the beetles probably have occupied these caves for less than 2,000 years. It is interesting to ask, where they were before that? Their high degree of troglobitic modification indicates that the beetles have had a very long history of cave occupation. Undoubtedly, the immediate sources of colonizers for the young lava tubes were older tubes in the immediate vicinity. New caves probably can be occupied via cracks from older caves as early as a score of years after their formation. New caves have to reach temperature equilibrium with their surroundings, and suitable moisture and food input mechanisms have to be established. This dynamic process of faunal migration from older into younger lava tubes has potentially been going on for a long time in many suitable sets of caves in the western United States. It may have begun as early as the Pliocene, as indicated by the diverse fauna of terrestrial and aquatic troglobites and troglophiles found in these caves. The list is too extensive to present here.

But, what was the original source of the beetles before they colonized lava tubes? It seems most likely that the lava tubes were colonized originally by preadapted *Glacicavicola* ancestors from nearby montane localities. Westcott (1968) suggests the mountainous vicinity where Idaho, Montana, and Wyoming come together. Once the beetles entered the caves, dispersal overland became less likely than dispersal through abundant cracks in the basaltic bedrock.

There are arguments against this interpretation, however. The present range of the beetle is not connected by continuous and suitable basalts, there are intermediate regions which are unsuitable for subterranean dispersal, the maximum diameter of the range (184 km) is very large for a beetle that ecologically and physiologically seems to be restricted to caves and there is a lack of substantial variation between the populations. These considerations suggest a fairly recent overland dispersal, probably during the cooler and wetter Wisconsinan glaciation, from either a montane or a cave habitat.

What would the climate have been like at this time? Unfortunately, there are virtually no data that can be used to answer this paleoclimatological question for the unglaciated Snake River Lava Plain during the late Pleistocene and Recent. Hence, it is very difficult to make any firmly based speculations upon the possible times and conditions of overland dispersal. Conditions may have been somewhat similar to those in western Oregon and western Washington (reviewed by Heusser [1965] for Liberty Lake, Washington in the late Wisconsian glacial and postglacial). There, a lodgepole pine parkland or, perhaps, a treeless tundra-like vegetation existed during the Wisconsinan glacial stage. With glacial retreat and the onset of warmer and dryer climates about 8,000 years BP, the vegetation changed to grasses, composites, and chenopods. This vegetation has persisted until today in an assemblage dominated by *Artemisia* sagebrush and *Agropyron* grass.

Such a vegetation would not seem to be suitable for overland dispersal of the beetle, but the size of the range and the lack of differentiated local populations favor overland acquisition of the present range during the Wisconsin. If the dispersal of *G. bathyscioides* was this recent, we may yet expect to find the beetles in other caves over even a greater area and, perhaps, in montane epigean habitats such as deep, wet talus slopes.

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Reconnaissance Geology of Timpanogos Cave, Wasatch County, Utah

William B. White * and James J. Van Gundy **

ABSTRACT

Timpanogos Cave is located in American Fork Canyon of the Wasatch Mountains. The cave consists of high, narrow passages developed along minor faults. These faults are parts of the complex fault structure near the intersection of the north-south trending Wasatch thrust system and the east-west trending Uinta fold system. Complex solution sculpturing indicates slow solution of the dolomitic Deseret limestone by percolating waters. Tilting of clastic sediments on the cave floor suggests fault movement after the formation of the cave.

The cave contains spectacular carbonate speleothems. Calcite occurs as dripstone and also as helictites and other erratic forms. Aragonite is common and occurs as needles and anthoditic forms. Moonmilk occurs sparsely as tufts as hydromagnesite on tips of dripstone deposits. Unusual yellow calcite and green aragonite are colored by Ni++, which locally may reach concentrations of several percent.

Introduction

Timpanogos Cave National Monument is located in American Fork Canyon, in the Central Wasatch Mountains about ten miles east of the town of American Fork and fifteen miles north of Provo. The cave entrances are high on the south wall of the canyon, at an elevation of 6700 ft.

Modern Timpanogos Cave consists of three formerly separate components: Hansen Cave, Middle Cave, and Timpanogos Cave, each with its own entrance but connected internally by tunnels. Hansen Cave was discovered in 1887, the other caves somewhat later. The entire area was set aside as a National Monument in 1922. Accounts of the discovery of the caves and associated history were published recently (Halliday, 1970).

Geological studies of the cave have been few. Those of Bullock (1942, 1954) are the primary references. Although the bedrock geology of this portion of the Wasatch is moderately well known, little has been written about the geomorphology of the area.

Field work for the present study was carried out by the writers during late June, 1970, with the coöperation of the National Park Service. The cave was re-mapped, observations were made of cavern features, minerals were collected for later analysis, and a brief and rather superficial reconnaissance was made of the surface landforms. The present paper emphasizes the results (primarily mineralogical) obtained from the study of the cave.

Geologic Setting

Geomorphology

The Wasatch Mountains were formed by a north-south trending fault block. The western margin of the block is a prominent fault scarp, which forms a sharp boundary between the mountain masses to the east and the Salt Lake basin to the west. Westward-flowing, sub-parallel, high-gradient streams have dissected the scarp into isolated peaks separated by deep canyons. One of these peaks is Mount Timpanogos, which rises to an elevation of 11,750 ft. The land slopes precipitously westward down to the elevation of the basin floor, at roughly 4800 ft. To the north and east of Mount Timpanogos, a major surface at 8000 ft., known as Sagebrush Flats, interrupts the descent of the land into American Fork Canyon.

The upper slopes of Mount Timpanogos are dominated by glacial landforms. Large cirques occur on the north and east and great patches of quaternary morainic material on the north flanks of the mountain. American Fork Canyon, however, is a deep, V-shaped valley containing little evidence for a previous glacial history. The canyon walls have a uniform slope from the level of Sagebrush Flats to that of the stream at 5500 ft. The location of Timpanogos Cave is shown in Fig. 1. At roughly 6700 ft, the cave lies halfway up the canyon wall between American Fork Creek and the Flats.

The Cenozoic history of the Wasatch landscape has been summarized by Threet (1959). The Pleistocene history of the canyons and, particularly, the relative roles of glacial and fluvial erosion do not seem to have been worked out.

Stratigraphy

This part of the Wasatch Mountains is made up of thousands of feet of sediments, mainly carbonates (Fig. 2) (Crittenden, 1959; Baker and Crittenden, 1961). The middle slopes of American Fork Canyon are underlain by the Fitchville formation (mainly a dolomite) and by the Gardison limestone (mainly a dolomite), and by the Deseret limestone (highly dolomitic) in which the cave is developed. On the higher slopes of the canyon is exposed the Humbug formation, a quartzitic sandstone with interbedded limestone and dolomite. Near the top of the canyon and underlying parts of Sagebrush Flats and Mahogany Mountain is the Great Blue Limestone, more than 3000 ft thick. Sagebrush Flats is partly underlain by the Manning Canyon Shale, which also contains some interbedded carbonates. Finally, the Oquirrh formation on the lower slopes of Mount Timpanogos contains the Bridal Veil limestone member of Pennsylvanian age. The Fitchville and Gardison would be considered equivalent to the Madison limestone, a major cave former in many of the western states. The Great Blue is considered to be of Upper Mississippian (Chesterian) age.

Timpanogos Cave lies entirely within the Deseret limestone. At the surface, this rock is a massive to medium-bedded, dark blue-black limestone that weathers to a dark grey. Of two bedrock samples, collected at the top and at

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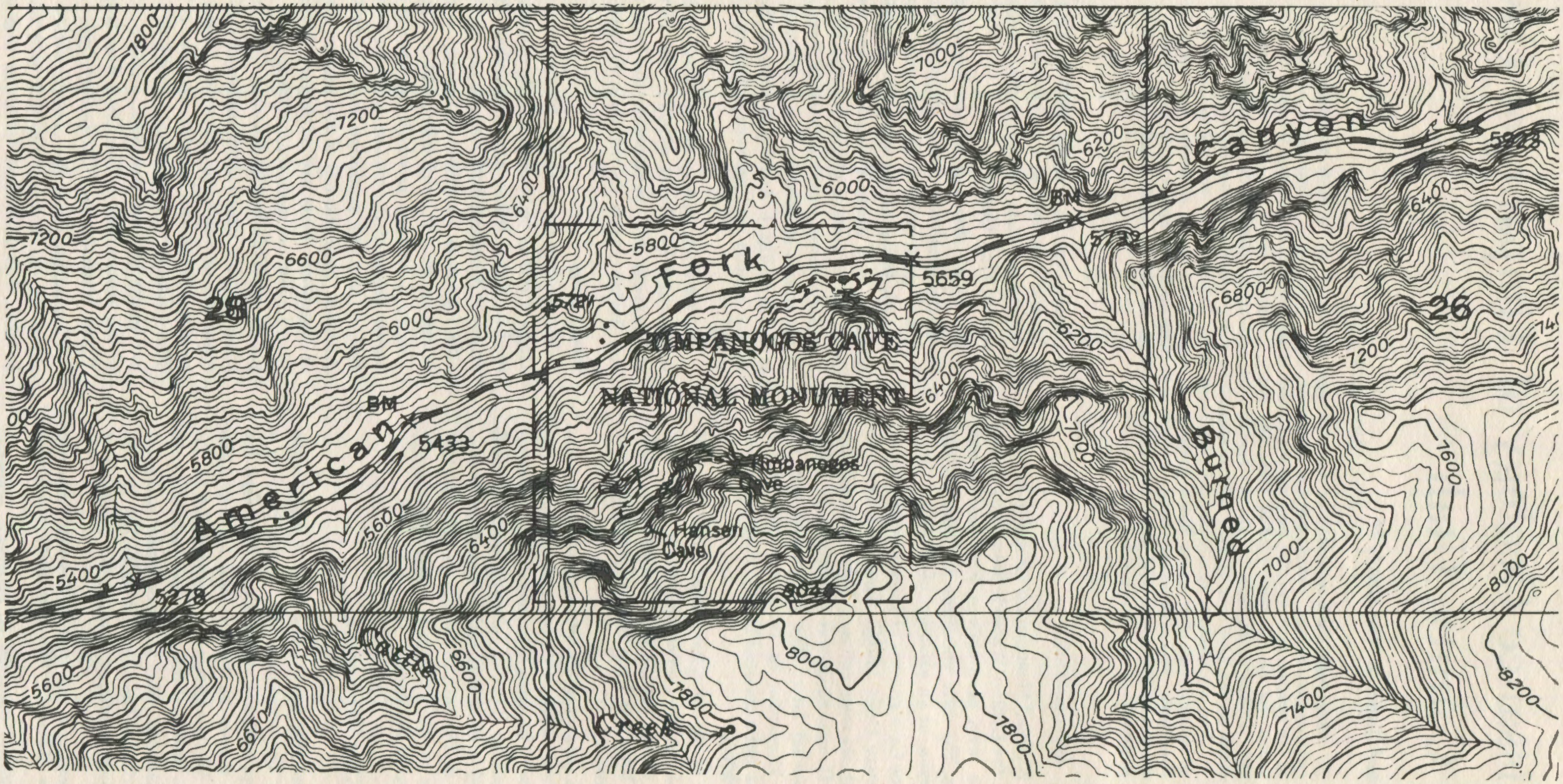

Fig. 1. Segment of USGS Timpanogos 7.5 minute quadrangle, showing locations of cave and monument on south wall of American Fork Canyon. The surface shown on the south-central edge of the figure is a portion of Sagebrush Flats.

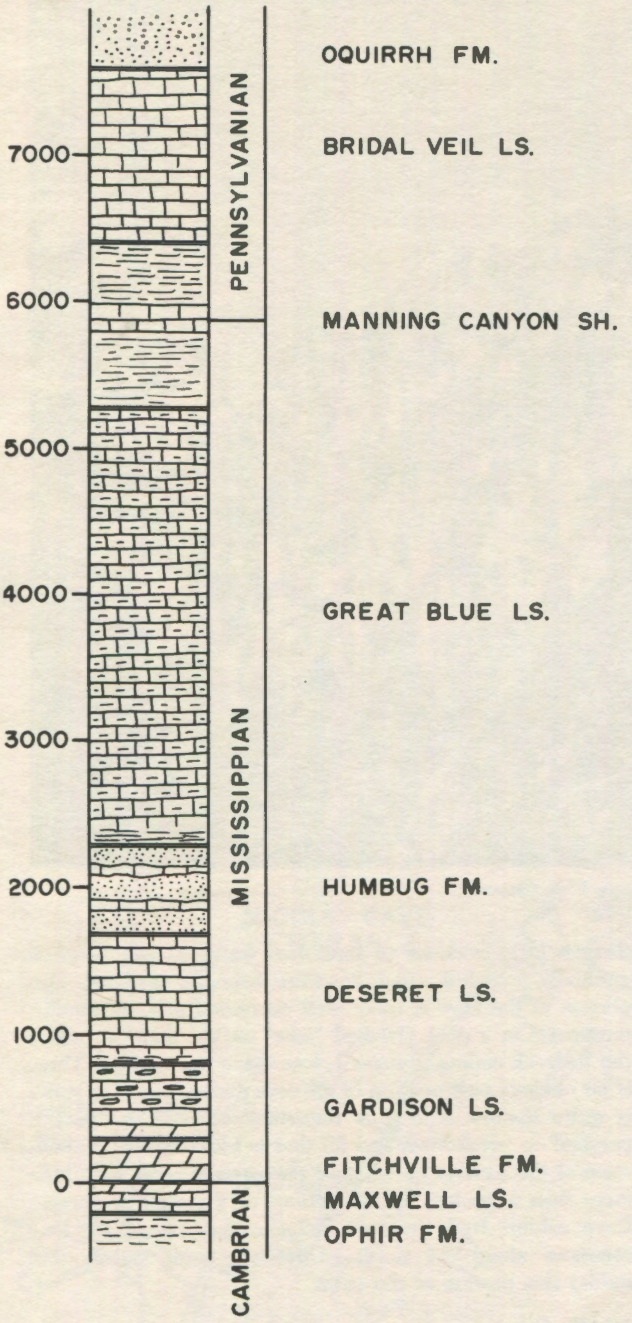

Fig. 2. Stratigraphic column for the Timpanogos area (adapted from Crittenden, 1959). Thicknesses are given in feet.

the bottom of the tunnel connecting Middle Cave with Timpanogos Cave, one x-rayed as dolomite and the other as a mixture of dolomite and calcite. The high proportion of dolomite throughout the carbonate section may have caused the paucity of cave and karst development in the cave area. Dolomite dissolves much less rapidly than limestone, and the amount of solution required to form caves rarely occurs in dolomite.

Structure

American Fork Canyon lies near the intersection of the north-south trending Wasatch fault and the east-west trending Uinta fold axis. The local structure is extremely complex. The cave lies in an area of block faulting. Some blocks have displacements on the order of 1000 feet. In the vicinity of the cave, dips average 20 to 30° toward the south. Many of the faults appear on Baker and Crittenden's (1961) geologic map, but the frequency of minor faults is too high to map adequately on a 7.5′ base.

The cave appears to lie within a single fault block, in which the beds dip 15° to 20° to the south. Cracked flowstone along the Middle Cave rift provides evidence of continued motion along the fault. Dipping sediments in Timpanogos Cave provide at least the suggestion that the entire block may have been tilted since the cave was formed.

Karst Landforms

There is little soil cover on the precipitous slopes of American Fork Canyon and bare rock ledges are common. In spite of the exposed limestones, however, characteristic karst landforms are only poorly developed. These are limited to minor solutional sculpturing and to a pinnacle karst on the Deseret limestone.

Solutional etching of joints and development of shallow solution pans occur in the Bridal Veil Falls limestone on the lower slopes of Mount Timpanogos, above Scout Falls. Patches of exposed limestone also occur, although they are not extensive enough to be called a limestone pavement. Small pinnacles a foot or so high occur on some of the ledges.

In the canyon itself, solutional sculpturing is not extensive in the dolomitic Maxwell, Fitchville, and Gardison formations. The Deseret, however, is carved into extensive bands of pinnacles that are clearly visible along the entire north wall of the canyon (Fig. 3). The pinnacles stand a few feet to tens of feet in height. A few similar pinnacles also occur in Little Cottonwood Canyon, a few miles to the north. On close examination, the pinnacles are seen to be rough and highly fractured, as though they gradually are being destroyed under contemporary weathering conditions. The smooth, sculptured surfaces that one usually associates with karst landforms are not present. This suggests, albeit without any supporting evidence, that the pinnacles are relict features out of equilibrium with present climatic conditions.

Description of Timpanogos Cave

The pattern of (modern) Timpanogos Cave is shown in Fig. 4. The three constituent caves are strongly linear and apparently are controlled by fractures or fault traces. The structural orientations of Hansen and Middle Caves are the same (N55°E, measured from true north), but the orientation of Timpanogos Cave (N45°E) is 10° northward from those of the other two. Hansen and Middle caves quite obviously are fragments of the same cave and are separated only by a flowstone plug. Timpanogos Cave appears to have been formed independently, along a different fracture system and, perhaps, in a different sub-unit of the fault block.

The (modern) cave exhibits little variation in elevation. The Hansen entrance is at 6730 ft and the Timpanogos entrance is at 6717, according to U. S. Geological Survey figures. Establishing the chamber at the end of the Grotto at the Hansen entrance as a reference zero, most of Hansen and Middle caves are at –13 feet. The lake at the southwest end of Hansen Cave is at –33 feet. The southwest end of Timpanogos Cave is distinctly lower than the rest of the

Fig. 3. Pinnacle karst on ledges of Deseret limestone, north wall of American Fork Canyon.

cave, with the lowest point (Father Time's Jewel Box) at –61 feet. The lowest point in the Chimes Chamber is –55 feet. The remainder of Timpanogos Cave is at about the same elevation as the Hansen entrance, although it drops slightly in elevation near the Timpanogos entrance. The (modern) cave, therefore, can be described as essentially horizontal, at a mean elevation of about 6720 ft, with several lower chambers.

Hansen Cave

The trail leading up from the Visitor Center to the Hansen entrance enters a small half-chamber, called the "Grotto". This must at one time have been an actual room in Hansen Cave, one side of which has now been stripped away by erosion of the canyon wall. An arch in the side of the Grotto leads to two superimposed tunnels that extend to the southeast. The lower of these tunnels has been enlarged by blasting. The higher is an irregular tube elongated along an oblique fracture. Both lead into the entrance chamber of Hansen Cave, a moderately large room 25 to 30 ft across. The original walls here are mostly fracture surfaces, although they have been obscured by flowstone to some extent.

The main portion of Hansen Cave extends to the southwest. From the entrance chamber, one clambers up a rubble slope, through a hole in the breakdown, and enters a segment of wide, moderately high passage about 200 ft in length. At the entrance end, the walls of this passage exhibit fresh fracture surfaces and great quantities of relatively fresh breakdown are apparent. Proceeding inward, again over a breakdown of massive blocks bounded by fresh fractures, there is little evidence of solutional walls. As one proceeds gradually downhill, much flowstone becomes apparent. This portion of the cave is quite well decorated and, eventually, terminates in a pool (labeled "lake" on the map) in which the bedrock ceiling plunges below the water surface. There is no obvious continuation of the cave underwater. The water is quite shallow, and it is uncertain whether the pool is perched on breakdown and fill that is blocking the continuation of the passage or whether the passage is, in fact, little more than a pocket. Many sections of the wall of Hansen Cave exhibit fresh fracture surfaces that extend for long distances along the northeast-trending fault system that guides this portion of the cave.

Middle Cave

Middle Cave consists, essentially, of a single high, narrow passage oriented along the N55°E guiding fracture. The natural entrance to the cave is a small hole 50 feet above the floor of the cave. To the southwest, the Middle Cave ceiling becomes lower, the passage widens somewhat, and the cave becomes decorated with a considerable amount of flowstone. It is here that the cave is connected to the entrance room of Hansen Cave by means of a 100-foot-long tunnel. The solutionally widened fracture continues to the southwest and comes very close to Hansen Cave.

To the northeast, the passage opens into a single large room ("Big Room", on the map) and, eventually, is terminated by a high, flowstone choke. The triangularly shaped block of limestone making up the southwest wall of the Big Room has slumped several feet and has separated from

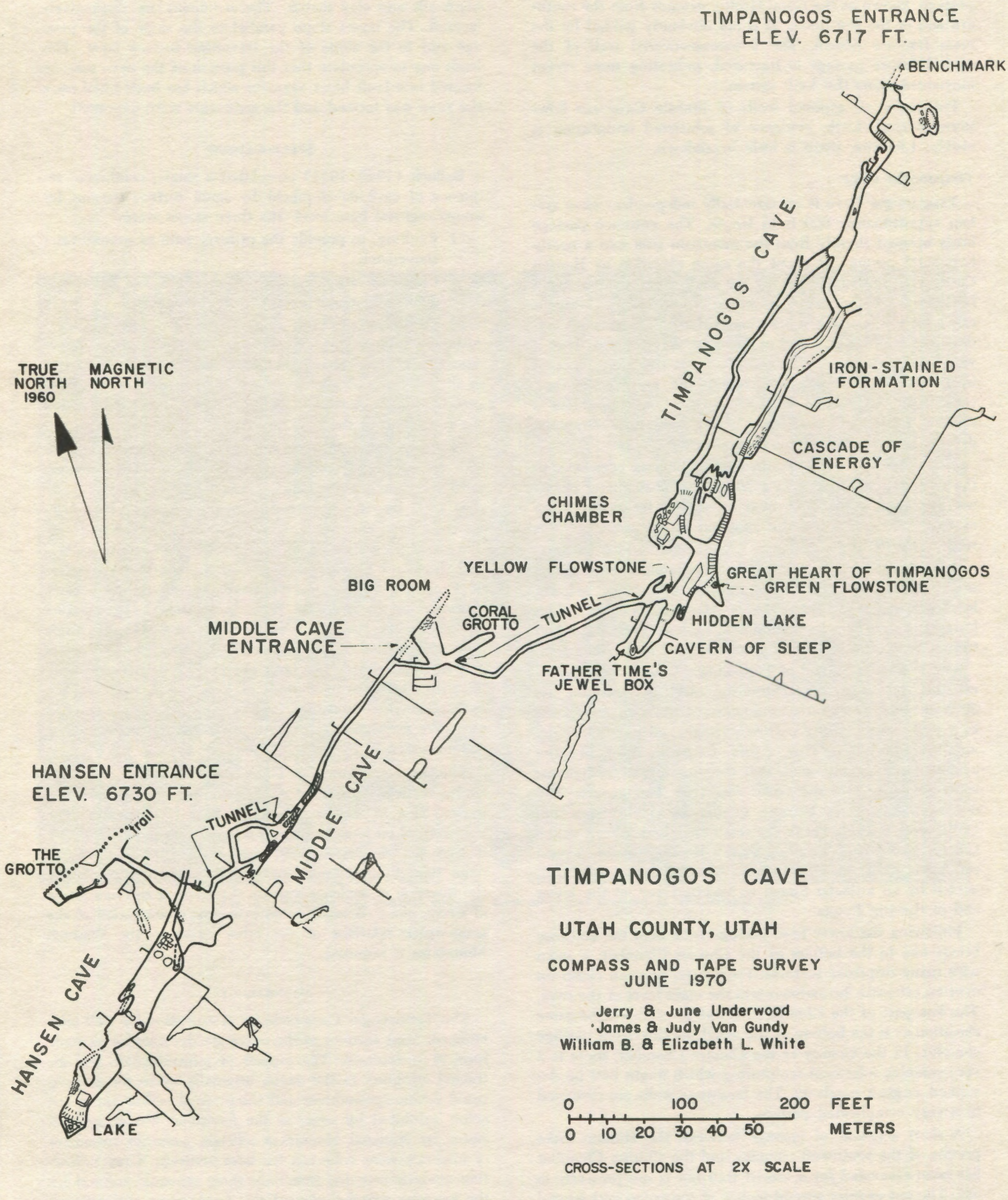

Fig. 4. Map of Timpanogos Cave, based on a new compass and tape survey.

the ceiling by several feet along a bedding plane. It is completely surrounded by fissure passages. A short, tubular passage, known as the Coral Grotto, extends from the southern end of the Big Room. It is not obviously guided by the main fracture system. The flowstone-covered wall of the central fissure passage is fractured, indicating some recent movement along this fault system.

Practically all exposed walls of Middle Cave are flowstone-covered. Little evidence of solutional sculpturing is visible. Likewise, there is little breakdown.

Timpanogos Cave

Timpanogos Cave is an essentially independent cave system approximately 600 ft in length. The entrance passage leads upward slightly from the mountain side into a nearly horizontal passage at about the same elevation as Hansen Cave. This passage trends to the southwest, maintaining a fairly large, though rather irregular, cross-section. The walls are smoothly sculptured solution surfaces. There is no evidence of scalloping in any of these passages, nor is there a well-defined gradient. The passage floors rise and fall in an irregular way. The passages follow the guiding fracture, which here is inclined from the vertical and produces a slanting passage cross-section. There are extensive flowstone deposits, particularly along the eastern sides.

Just inside the entrance, an upward-sloping crawlway to the southeast brings one to a 20 to 30 foot-diameter chamber containing a pit. The pit is reputed to be on the order of 100 ft deep but was not explored in the course of this work.

Approximately 150 ft from the Timpanogos entrance, a side passage descends to the right and becomes a lower level passage about 40 ft below the main channel. This lower passage likewise contains complex solutional sculpturing, lacks a well-defined gradient, and lacks scalloping.

The southwestern end of Timpanogos Cave is rather complicated. It becomes precipitously lower in elevation, by 30 ft or more, and continues downward along parallel joints as a sequence of rather narrow passages, all paralleling the main fracture system. One of these forms the small off-shoot against the structural grain that contains the large, bulbous stalactite known as the Great Heart of Timpanogos. The down-slope passages become smaller in cross-section and end in dripstone chokes in the small chambers known as the Cavern of Sleep and Father Time's Jewel Box, the deepest points in the cave. This complex of small passages is connected by an artificial tunnel to Middle Cave, at the lower end of the Big Room.

Following the lower level passage from near the entrance brings one to the bottom of the Chimes Chamber, a room with many dripstone features. From here, one can climb up over an extensive breakdown into the main level of the cave. The low part of the Chimes Chamber is at about the same elevation as is the bottom of the small, southwestern passage complex. In the vicinity of the Chimes Chamber, there is a very complex solutional sculpturing which might best be described as giant pendants. The limestone walls are dissolved in a very complicated pattern.

A short segment of passage between the Hidden Lake portion of the southwest complex and the Chimes Chamber has been excavated for a tourist trail but is not presently in use. The cut exposes several inches of travertine and several feet of underlying clastic sediment. This is one of the few places in the cave where the clastic sediments are well exposed. Here, they consist of clear grains of quartz sand a few tenths of a mm to one mm in size and have a brown-black silt and clay matrix. The sediments are distinctively layered. The layers slope parallel to the slope of the passage and to the slope of the travertine surface layer. This leads one to speculate that this portion of the cave may be located in a fault block segment which has been tilted since the cave was formed and the sediments were deposited.

Speleogenesis

Bullock (1942, 1954) postulated a single, continuous sequence of cave development by small surface streams far above regional base level. His three stages were:

i. Faulting, to provide the primary path of groundwater movement.
ii. Cave excavation by solution, corrasion, and removal of fault gouge and breccia.
iii. Deposition of clastic sediments and speleothems.

Although Bullock did not assign an age to the Mt. Timpanogos caves, his mechanism of development places the caves in the very recent geologic past. As evidence, he cites the presence of clastic sediments, including stream pebbles, and the restriction of the cave to active fault traces.

The present investigation supports Bullock's structural controls completely. Examination of the solutional sculpturing, however, provides no evidence that the cave ever was occupied by a free surface stream or that it was excavated by water in open channel flow. The walls are smooth and regular, with no scalloping. The large, joint-controlled pockets and dead-end passage segments in the southwest complex of Timpanogos Cave and the pendants and ceiling pockets elsewhere in the cave all point to solution by percolating water. Because the cave is located on a mountainside, these lines of evidence do not necessarily imply a phreatic origin in the classic sense. The fault openings, for example, could fill with water during periods of high runoff such as from spring snow melt and then drain during drier seasons of the year. We merely argue that the cave was not excavated by stream action in the usual sense.

The caves are, however, suspiciously horizontal, despite a strong vertical component in their controlling structural elements. This implies some sort of base level control. The only control on local base level is American Fork Creek and, if they are associated with it, the caves must be much older than Bullock envisioned. It is not possible to answer the question of speleogenesis on the basis of this one group of caves, alone. A much more extensive investigation of the geomorphic relations of the other caves in the Wasatch Mountains is required.

Mineralogy

The Timpanogos Cave complex is exceptionally well mineralized. Most sections of the caves are decorated with some form of speleothem. The aspects of mineralization that attracted attention to the caves, originally, were the complicated erratic speleothems and the unusual yellow and green colors exhibited by some of the flowstone. Under authorization by National Monument officials, some 20 specimens of minerals were collected for later analysis. X-ray diffraction analysis revealed that only three minerals account for the complex morphologies of the Timpanogos speleothems: calcite, aragonite, and hydromagnesite. Emission spectrographic analyses of selected specimens are listed in Table 1.

Fig. 5. Helictites and spicular aragonite from southwest complex of Timpanogos Cave.

TABLE 1. Analyses of Carbonate Minerals

Ref. No.	Specimen	MgO(%)	SrO(%)	Ti(ppm)	Mn(ppm)	Ni(ppm)	Fe(ppm)	Si(ppm)	Al(ppm)
833	Aragonite spicules	0.05	0.158	50-80	5-20	80-120	20-50	>7000	20-40
834	Calcite crystals	1.68	0.021	n. d.	5-20	100-150	20-50	600-800	80-120
836	Calcite flowstone	1.61	0.023	n. d.	5-20	120-170	20-50	400-600	25-50
837	Pale yellow stalactite: calcite core with aragonite exterior	1.32	0.198	n. d.	5-20	250-300	50-100	600-800	150-200
838	Yellow calcite flowstone	1.75	0.035	70-100	20-40	400-600	400-600	>1000	>1000
839	Green aragonite coating	1.45	0.137	n. d.	5-20	400-600	50-100	>1000	25-50

Calcite makes up the bulk of the flowstone and dripstone deposits. Aragonite is plentiful, but is restricted mainly to the erratic speleothems and to other, smaller, forms. Hydromagnesite occurs as moonmilk.

Dripstone and Flowstone

The flowstone and dripstone that occurs most commonly is coarse-grained and white or light brown in color. Those broken speleothems observed in cross-section appeared to be calcite throughout, with little evidence for aragonite banding. One specimen did show a mixed calcite-aragonite assemblage as a white, rather powdery layer between the more massive, calcitic outer portion of the speleothem and the dolomitic bedrock. Fracture surfaces showed either grains with radially oriented c-axes, or massive, polycrystalline forms that break with an irregular fracture. Monocrystalline dripstone or soda straw varieties were not in evidence.

Exceptions to the normal white and pale brown or tan colors occur in the Timpanogos Cave section. In the upper level, the large flowstone cascade known as the Cascade of Energy is a deep chocolate brown. Its composition was not investigated. In the southwest complex of Timpanogos Cave, many flowstone wall coatings are of a pale lemon yellow color which, in places, ranges to a deep clear yellow. The wall of the small chamber containing the Great Heart of Timpanogos is covered with an irregular layer of pale green flowstone. The origin of these colors was a special subject of investigation and is treated in a later section.

Five flowstone and dripstone specimens were tested for luminescence by placing chips in an evacuated tube and inducing a plasma discharge with a small Tesla Coil. This technique is a quick test for cathodoluminescence. In addition, more minerals are likely to fluoresce in this way than under ultraviolet light. A faint pink luminescence was produced on the powdery outside surface of two samples; there was no luminescence from a fresh, fracture surface. Three specimens did not luminesce at all. This negative result is itself unusual, for most cave calcite specimens are luminescent to some extent.

Erratic Speleothems

Bullock (1942, 1954) lists 16 types of speleothems in Timpanogos Cave. Most of these are different morphologies of dripstone and flowstone; the erratic forms are helictites, club-like, wart-like, and nodular forms, and aragonite crystals. We distinguish essentially these same forms under the names of calcitic helictites, aragonitic helictites, globulites, and spicular aragonite. The forms occur together in the Coral Grotto, in the southwest complex of (original) Timpanogos Cave, and in the Chimes Chamber. Small erratic forms are uncommon elsewhere in the caves. Some idea of the complexity of these groupings can be obtained from Fig. 5, which shows a variety of helictites and spicular aragonite. Nodular or globulite forms are shown in Fig. 6, which also illustrates later flowstone (in this case, a lemon-yellow flowstone) covering erratic speleothems deposited earlier.

Fig. 6. Yellow flowstone covering nodular speleothems, southwest complex, Timpanogos Cave.

Two distinct morphologies of helictite can be recognized. Calcitic helictites have smooth exteriors, are curved and twisted in complex ways, and are composed primarily of calcite. Fig. 7 shows a typical example. The speleothem is polycrystalline and appears to have (or to have had) a

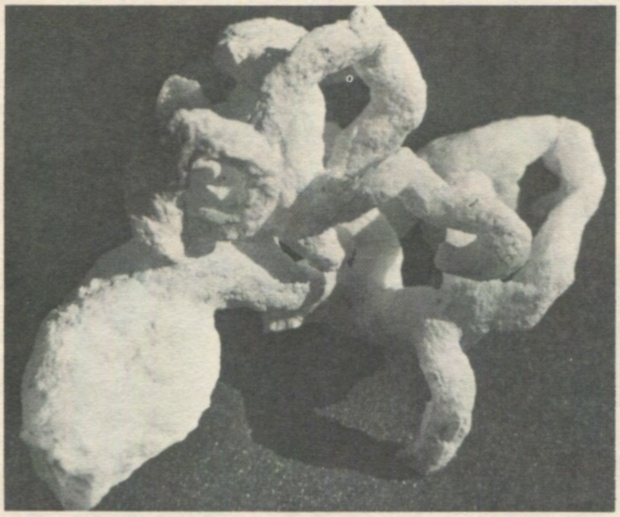

Fig. 7. Calcite helictite.

central canal. Some calcitic helictites occur with a coating of acicular aragonite crystals a few millimeters long radiating outward from the surface. Such cases appear to have resulted from secondary growth after the precipitation of the helictite core.

Aragonitic helictites tend to be linear, although they jut out from the walls at all angles. They have rough or jagged surfaces and are composed primarily of aragonite. Microscopic examination shows that many of these have central cores of finely crystalline aragonite, around which are arranged acicular, clear crystals of aragonite a few millimeters in length. The axes of the acicular crystals make an angle of 10° to 20° with the axis of the helictite (Fig. 8). A sequence of forms occurs in which this basic aragonitic helictite structure is modified by calcite overgrowths. In many specimens, small, transparent calcite crystals are imbedded in the aragonite needles near the tip of the helictite. These appear to be randomly oriented. No obvious topotactic relationship between the calcite and the underlying aragonite is apparent. As a further modification, the exterior of the helictite becomes completely covered with small calcite crystals, but the core materials retain the same structure as the calcite-free specimens.

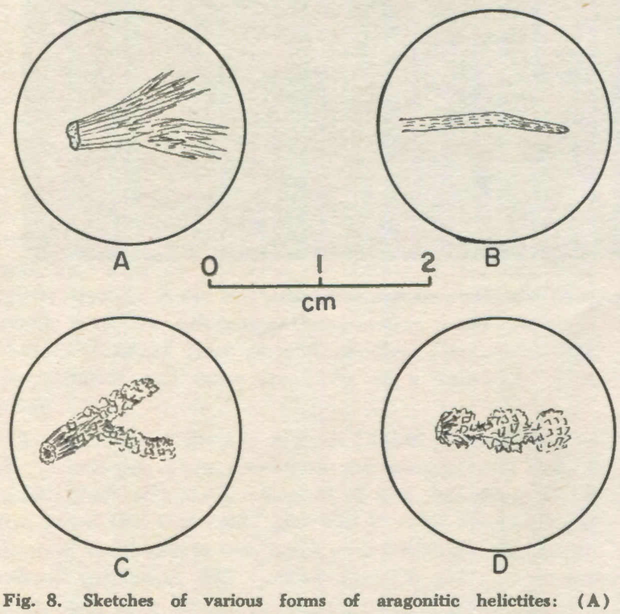

Fig. 8. Sketches of various forms of aragonitic helictites: (A) Spicular form that occurs in small tufts and coatings on surfaces of wall deposits and other speleothems. These are, in effect, protoanthodites. (B) Typical long, linear forms composed completely of aragonite. The orientations of the individual crystals diverge only slightly from the helictite axis. (C) A branching form composed mainly of aragonite, with small rhombs of calcite near the tips. (D) A nodular form, with an inner core of aragonite, entirely coated with small, randomly oriented rhombs of calcite.

Associated with other small, erratic forms are small clusters of acicular aragonite crystals that appear as bush-like clusters of crystals radiating from a common "stalk." These might be considered small anthodites but are here termed spicular aragonite. Their appearance in the cave is indicated in Fig. 5 and a close-up of a single stalk is shown in Fig. 9.

A variety of nodular, or globulitic, forms also occur. Some specimens appear as small spherical lumps on the tips of spicular aragonite clusters. The beginnings of such growth can be seen in Fig. 9. The spherical portions of these speleothems are calcite, while the supporting stems are aragonite. There are many variations in morphology, mostly not well defined, and no further classification is attempted here. Some indication of the variation can be seen in Fig. 6. In general, these appear to be similar in morphology to the speleothems in Carlsbad Caverns described by Thrailkill (1971).

Moonmilk

The spicular aragonite that occurs in the Coral Grotto and in several other highly mineralized parts of the cave is tipped with a loose, white, powdery moonmilk. The material seems to be associated only with aragonite and was not observed as coatings or as massive deposits. Three samples were collected:

Number	*Locality*	*Mineral Assemblage*
826	Tips of helictites, Coral Gardens	Hydromagnesite
827	Formed over surface of white calcite dripstone	Calcite and hydromagnesite, plus aragonite
840	From tips of aragonite spicules, Great Heart of Timpanogos	Hydromagnesite

The identification of hydromagnesite, $4MgCO_3 \cdot Mg(OH)_2 \cdot 4H_2O$, as the principal constituent of the moonmilk is consistent with results from other caves in temperate climates. However, Timpanogos Cave is sufficiently high and cold that one might have expected a calcite moonmilk, such as occurs in high altitude caves in the Rockies.

Mechanism of Aragonite Deposition

Aragonite is more soluble than calcite at atmospheric pressure and at cavern temperatures of about 10°C. It is not, therefore, thermodynamically stable under cave conditions. However, the solubility of aragonite exceeds the solubility of calcite only by about 11%. If the speleothem-form-

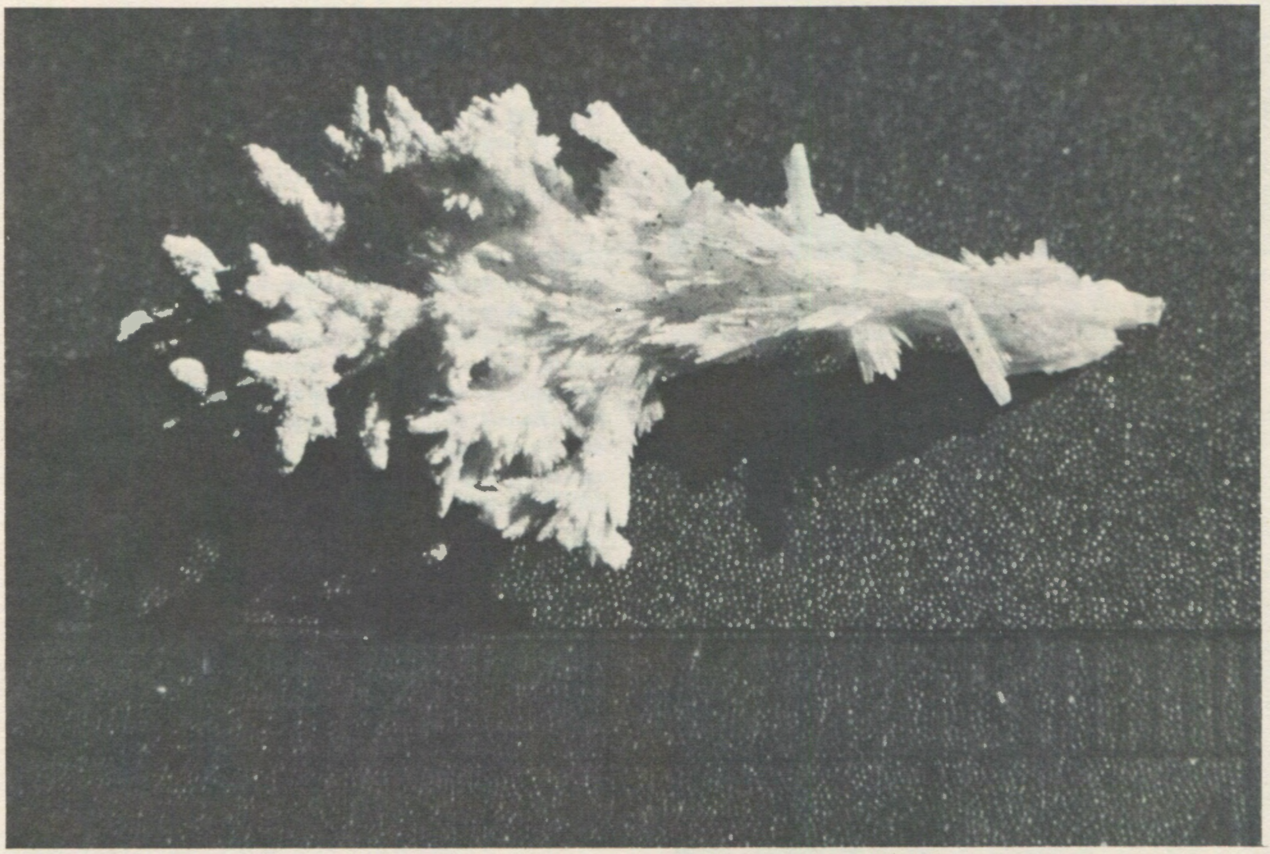

Fig. 9. Close-up of a clump of spicular aragonite crystals. The scale is in millimeters.

ing solutions were to become supersaturated by more than 11%, either calcite or aragonite could be precipitated and the mechanism would become one of kinetics. The kinetics, in turn, are controlled by factors affecting the rate of nucleation of calcite and aragonite, because this is the rate-controlling step. It has long been known that Mg++ enhances the growth of aragonite and that it acts by inhibiting the nucleation of calcite, thus permitting the supersaturation necessary for aragonite precipitation to build up. Because the bedrock at Timpanogos Cave is highly dolomitic, percolating waters should be magnesium-rich. The speleothems analyzed range on the order of 1 to 2% $MgCO_3$, with no obvious differences between those of calcite and those of aragonite.

The aragonitic helictites sometimes have calcite overgrowths. The calcite always occurs near the tip. It seems likely that the solutions which bathe the speleothem first deposit aragonite at high supersaturation. As the level of supersaturation falls, calcite begins to be precipitated. The spicular aragonite may be formed either from more highly supersaturated solutions or from more magnesium-rich solutions. Because the magnesium in the original water remains in solution and does not become incorporated into the speleothem, the final solution is very magnesium-rich and it is by the evaporation of this that the tufts of hydromagnesite on the tips of the spicular speleothems are formed.

Explanation for the Yellow and Green Flowstone

Much of the flowstone in the southwest complex of Timpanogos Cave is colored in varied shades of yellow. Some is a very deep lemon-yellow. Pale-green flowstone is confined to the wall of the chamber that houses the Great Heart of Timpanogos. The results that follow were obtained from two small specimens removed from the walls with Park Service authorization. Because the colored flowstone areas lie close to the visitor trail, exceptional care was used to remove only the least amount of material sufficient for analysis. We will show that both colors are due to substantial concentrations of nickel: the yellow, from nickel in solid solution in calcite and the green, from traces of a nickel-rich phase distributed in aragonite.

X-ray analysis of the yellow flowstone reveals only calcite. Microscopic examination shows that the yellow color is homogeneously distributed and that the material is clear and coarse-grained. The material breaks with an irregular fracture, indicating no preferred orientation of the individual calcite crystals. X-ray analysis of the green material shows mainly aragonite. Microscopic examination reveals radiating clusters of nearly clear acicular needles. The color is not homogeneous and there are blebs of a yellowish, waxy phase.

Emission spectrographic analysis (Table 1) indicates that all Timpanogos Cave specimens are exceptionally rich in nickel: there is a smaller amount of iron, but other transition metal ions that could give rise to color are either not detected or present in very small concentrations. The deeply colored specimens, 838 (yellow calcite) and 839 (green aragonite), both contain about 500 ppm (0.05 weight percent) nickel.

Proof that the Ni^{2+} ion is responsible for both yellow and green colors is provided by the reflectance spectra of the chips of speleothem (Fig. 10). These were obtained by mounting fresh fracture surfaces in the diffuse reflectance attachment of a Beckman DK-2A spectrophotometer. The

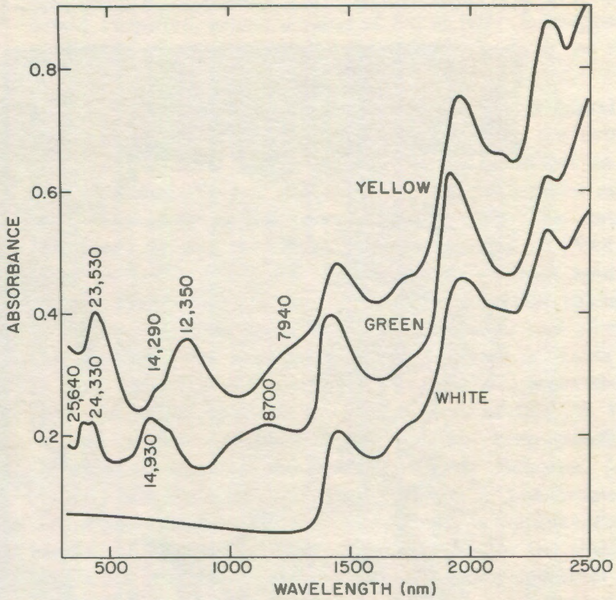

Fig. 10. Diffuse reflectance spectra of colored flowstones compared with that of a white specimen. The frequency of each absorption feature is indicated in cm–1

spectra show the distribution of absorbed light as a function of wavelength through both the visible and near infrared. Absorption bands at wavelengths greater than 1500 nm are common to all specimens, including the white one. These absorptions have been shown in other work to be due to vibrations of the carbonate anion and are not relevant to this study. Below 1500 nm, the spectrum of the white specimen is featureless, but three absorption bands appear in the spectra of both the green and yellow specimens. The main difference between the two is a small frequency shift. These bands are characteristic of the Ni++ ion and arise from electronic transitions in the unfilled d-orbitals through their interaction with the electrostatic field produced by the coordinating oxygen ions in the carbonate crystal structure. An analysis of the spectra by crystal field theory* allows an assignment of the bands to be made and, also, a calculation of the crystal field parameters. These confirm the hypothesis that the color of both specimens is due to nickel (Table 2). The value of the B-parameter is typical for Ni++ in an oxide host. The parameter Dq requires more comment. Dq measures the strength of the interaction between the Ni^{2+} ion and its surroundings. It is a function of the metal-anion distance and the coordination number.

The value for the calcite specimen is 794 cm^{-1}, lower than the value for NiO itself (Newman and Chrenko, 1959) or for Ni^{2+} substituting for the similar-sized Mg^{2+} ion in MgO (Pappalardo, et al. 1961). This would be expected if Ni^{2+} were substituting for the larger Ca^{2+} ion in the 6-coordinated site of the calcite structure. Occupancy of the relatively larger Ca^{2+} site by Ni^{2+} results in a weakening of the crystal field and a corresponding shift of the spectral bands to the red. The result is that the absorbance

* Crystal field theory and the optical absorption spectroscopy of insulating transition metal compounds is outside the scope of the present investigation. Readers are referred to textbooks by Figgis (1966) or Cotton (1971) for a discussion of the theory and to Burns (1970) for its application to mineralogy.

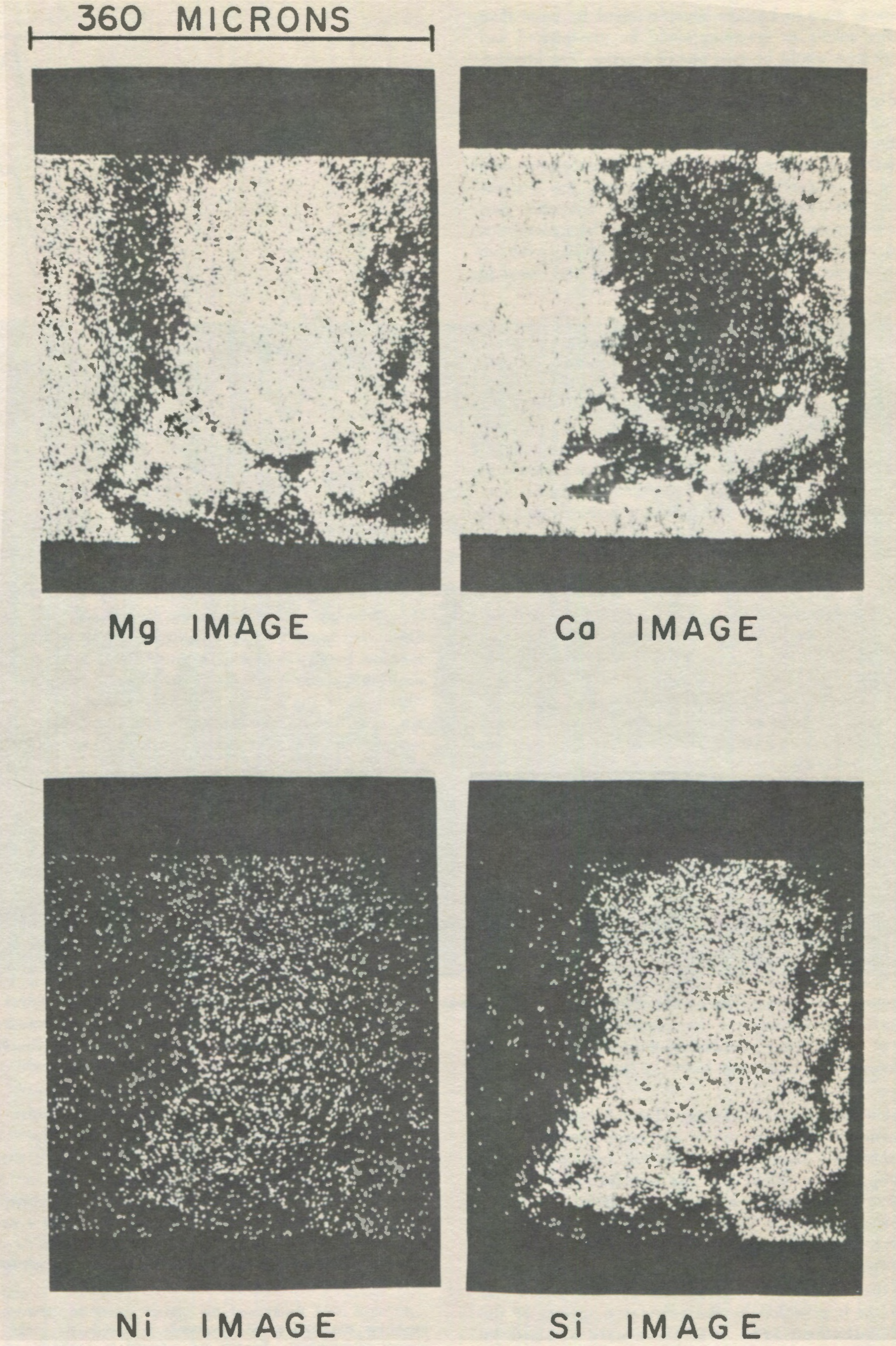

Fig. 11. Electron microprobe images of the green, aragonitic flowstone, showing distribution of Mg, Ca, Ni, and Si in one of the second phase blebs.

TABLE 2. Frequencies and Crystal Field Parameters for Ni^{2+} in Speleothems

Ref. No.	838 (yellow calcite)	839 (green aragonite complex)
Band Assignments	Band Frequencies (cm^{-1})	
$^4A_{2g} \rightarrow {}^4T_{2g}$	7,940	8,700
$^4A_{2g} \rightarrow {}^4T_{1g}$ (F)	12,350	14,930
$^4A_{2g} \rightarrow {}^4T_{1g}$ (P)	23,530	25,640 24,330
Crystal Field Parameters (cm^{-1})		
Dq	794	870
B	860	883

minimum between the 12,350 cm^{-1} band and the 23,530 cm^{-1} band is shifted to 620 nm, the crystals transmit red and yellow light preferentially, and the color appears yellow. This is in contrast to most nickel salts, where the absorbance minimum occurs in the range of 525 to 550 nm and the compounds appear green.

The Dq parameter for the aragonite specimen is just about what one would expect of Ni^{2+} in normal nickel salts or in aqueous solution and, therefore, the specimen exhibits a "normal" nickel green. It is not reasonable, however, to expect such behavior from Ni^{2+} in the aragonite structure. The Ca^{2+} site in aragonite is larger than in calcite and the coordination is 9-fold instead of 6-fold. One would, therefore, expect a value of Dq of Ni^{2+} in aragonite smaller than its value in calcite, and such clearly is not the case.

Because the solubility of $NiCO_3$ in aragonite is expected to be small, our first thought was that $NiCO_3$, the newly discovered mineral gaspeite, had perhaps exsolved as a separate phase. To test this idea, electron microprobe analyses were made of the green aragonite chip. The results are shown in Fig. 11. Rather than occurring as $NiCO_3$, nickel is localized in small blebs of material which appear to be a magnesium silicate. Figure 11 shows that calcium is depleted in these areas, whereas magnesium and silicon are enriched. The average chemical analyses of Table 1 likewise show that the green aragonite contains more silica than the other speleothems. Exact identification of the mineral in the nickel-rich phase has not been possible. Quantitative electron microprobe analysis indicates (by weight) 52% MgO, 38% SiO_2, and up to 2% NiO. A hydrated magnesium silicate of the serpentine family is the most likely candidate. Because the Ni^{2+} is substituting for the smaller Mg^{2+} ion, the colors and optical absorption spectrum of the hydrated silicate are more similar to those of the reference oxides.

Acknowledgments

We are indebted to Monument Superintendent Don H. Castleberry for permission to carry out this study and to Naturalist Roger Siglin and Ranger Lloyd Jacklin for their assistance in the cave. The mapping was done with the assistance of Elizabeth L. White, Judith Van Gundy, and Jerry and June Underwood.

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Unusual Mineralogy of the Crystal Pit Spatter Cone, Craters of the Moon National Monument, Idaho.

Stewart B. Peck *

ABSTRACT

Crystal Pit Spatter Cone lies on the Idaho Rift System of the Snake River Basin Lava Plateau in south-central Idaho. The former magma chamber under the spatter cone, about 80 feet long and 35 feet wide, is still open for examination. It is reached by a descent of 90 feet through the narrow spatter cone throat. The magma chamber contains large quantities of the secondary sulfate minerals gypsum, mirabilite, and jarosite, all of which seem to be scarce or absent in caves in other volcanic regions. The gypsum and mirabilite probably were deposited from mineralized capillary groundwater seeping into Crystal Pit, rather than by condensation from mineralized volcanic gasses. However, the basaltic rocks overlying Crystal Pit seem to be poor in calcium-, sodium-, and sulfur-containing primary minerals that could be leached by the capillary groundwater. This extensive mineralization has occurred within the past 1000 to 2000 years.

Introduction

Few data have been reported on the secondary mineralization of volcanic caves in the United States. Secondary calcite mineralization, so common in limestone caves, is scarce or absent. The description of coralloidal opal from the ceilings of lava tubes in Lava Beds National Monument, California by Swartzlow and Keller (1937) is the only other report known to me concerning extensive secondary mineralization in a volcanic cave in the United States. The purpose of this paper is to describe Crystal Pit and to bring its minerals to the attention of interested speleologists and mineralogists.

History of Exploration

The earliest known exploration of Crystal Pit was that of Bob Limbert, in 1922. Limbert explored the lava plains around the Monument and published an article describing them in the March, 1924 *National Geographic Magazine.* Samuel A. Paisley, the first custodian of the Monument, probably entered the pit in 1925. Paisley displayed crystal-coated rocks, probably from Crystal Pit, for Monument visitors in 1926 (mentioned in the first of the *Superintendents Monthly Narrative Reports*). The geologist H. T. Stearns, who may have entered the pit in 1927 or '28, included a description of the pit in his unpublished manuscript on the geology of the Monument (Stearns, 1926-1928). A photograph by Paisley of the interior of the Crystal Pit was included by Stearns in his guide to the Monument (1959). R. C. Zinc (1956), in a short history of the Monument, reported that Boy Scouts from Pittsburgh, Pennsylvania, explored the pit to a depth of 163 feet in 1931 and found no floor. There are no other recorded explorations before mine of 1961 and '62.

Description of the Crystal Pit

Crystal Pit (Fig. 1) consists of the throat and empty lava chamber of a spatter cone (Fig. 2). Spatter cones are formed directly over lava-containing fissures by accumulated clots of expelled lava. The clots partially fuse and adhere to one another after falling. Spatter cones are not known to exceed 100 feet in height and rarely are more than 100 feet wide. They usually are formed during the closing stages of eruptions of degassed, viscous, basic lavas. Occasionally, the throats of spatter cones are open, but ones in which the magma subsided sufficiently before cooling to reveal the magma chamber are very rare.

Beginning at the southeast side of Big Crater Butte Cinder Cone, at Craters of the Moon, a chain of 11 spatter cones five to 50 feet in height trends southeastward. The chain is said by Stearns (1959) to be one of the most-nearly perfect in the world. None of the cones in the chain has an open throat except Crystal Pit. The first and second cones formerly could be entered to depths of 35 and of 50 to 60 feet, respectively, but both now are blocked by debris. Peck (1962) gave additional information about these cones.

Crystal Pit has an opening at the top about three feet in diameter. The throat maintains a circular cross-section for 60 feet, by which point it has become six to seven feet across. It resembles a well with walls of lava. It then abruptly bells out into a chamber 30 feet high (Fig. 3). The chamber is about 35 feet wide and 70 to 80 feet long. Its long axis represents the northwest-southeast trend of the rift from which the lava originally flowed. Extensive deposits of mirabilite and patches of gypsum occur in the chamber.

The floor of the chamber is formed almost entirely by loose cinders and lava talus. A five- to eight-foot wide ledge about 10 feet above the bottom was formed by a lava pool which cooled at this level. When the pool subsided, the ledge was left marking its former shoreline. There are indications of four other, less well-developed, lava-pool levels. In a southeasterly direction, the chamber soon ends in a small room where the ceiling descends to meet the floor. The ceiling here is covered by extensive deposits of mirabilite. To the northwest, the cave floor descends to a point 120 feet below the entrance and the chamber narrows to a passage 10 feet high and 60 feet long (Fig. 4). Sterns (1926-1928) suggested that from this lower passage, molten lava escaped to feed the margin of the flow surrounding Crystal Pit. Mirabilite and gypsum are abundant at the end of the passage and jarosite is common as a layer of powdery, loose, yellow material on

* Department of Biology, Carleton University, Ottawa, Ontario, Canada.

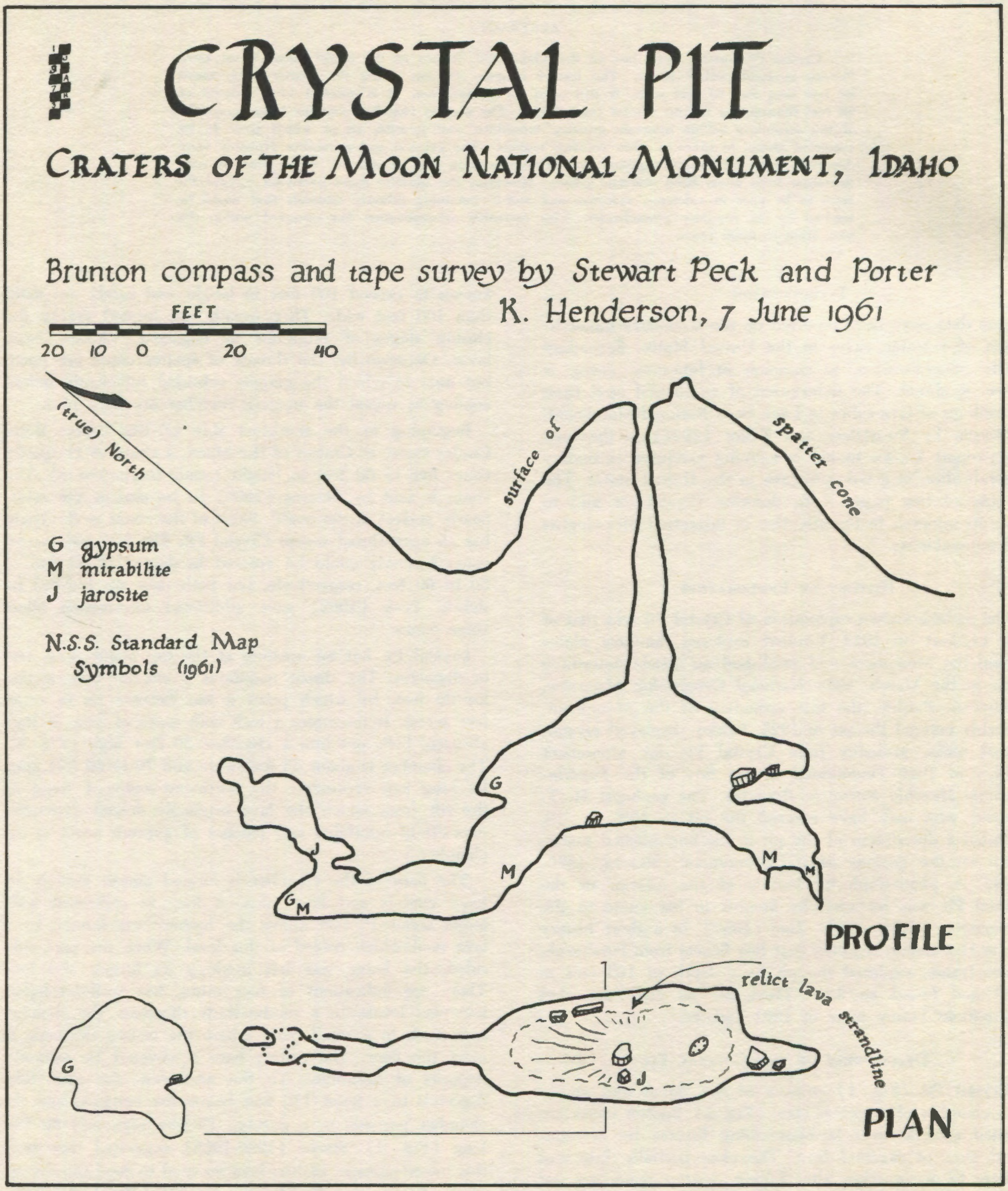

Fig. 1. Map of Crystal Pit.

Fig. 2. Portion of spatter cone chain, Craters of the Moon National Monument, Idaho. Crystal Pit Spatter Cone on the right.

the floor. Above the termination of the passage is a hole leading upward into a small chamber. Still another, smaller, chamber lies above that one. The walls of the chambers are three-inch thick crusts of lava, behind which are red, unfused cinders. These chambers probably underlie the adjacent spatter cone.

The walls, the ceiling, and the underside of the lava-pool ledge bristle with sharp lava stalatites, of which a few are nearly 12 inches long. Gypsum crusts cover many of the stalactites, at times so completely that their basaltic

Fig. 3. Interior of Crystal Pit, showing junction of spatter cone throat and magma chamber.

Fig. 4. Northwest end of magma chamber and entrance to northwest passage, showing mineral crusts—especially, those concentrated along fractures in the wall rock.

cores are obscured. Crusts also occur on the wall and on the floor. The crusts are composed of individual, simple, monoclinic crystals up to one inch in length (Figs. 5, 6) The crystals average about 1/8 inch long and have well developed crystal faces. Gypsum needles and gypsum hairs were not observed.

Description of the Minerals

Gypsum ($CaSO_4 \cdot 2H_20$)

This mineral frequently occurs in limestone caves, but seemingly is rare in lava tubes. Its mode of occurrence in Crystal Pit was described in the preceding paragraph. Gypsum occurs in other spatter cones in the series including Crystal Pit (Peck, 1962), also, and in Great Owl Caverns lava tube.

Mirabilite ($Na_2SO_4 \cdot 10\ H_20$)

This unstable, hydrated sodium sulfate is known from caves and, also, from prótected sites in recent lavas (Palache, Berman, and Frondel, 1951). It occurs as stalactitic masses in the Flint Ridge Cave System, Mammoth Cave National Park, Kentucky. Mirabilite is the stable phase in the system Na_2SO_4 - H_20 and forms below 32.38°F. Above this temperature, it becomes unstable and can change to thenardite. My sample was carried, sealed and under refrigeration, to Menlo Park, California for analysis. When not thus protected, and when occurring in warm or dry caves, the water is lost from mirabilite and thenardite results. The deep

Fig. 5. Mirabilite on lava slope. Jarosite lies as a loose powder on the mirabilite toward the bottom of the view. Carbide lamp provides scale.

Fig. 6. Mirabilite on walls, gypsum crystals on wall in lower right and lower left.

caves of the Monument are environmentally suitable for this sensitive mineral because their temperatures range from 28° to 35°F (the temperature of Crystal Pit was not measured).

The presence of thenardite in the Monument previously had been determined by analysis at the University of Utah. Also, both the hydrated and dehydrated phases are known to occur in the Monument in Arco Tunnel, Snow Cone, Lava River Cave, Indian Cave, and Great Owl Caverns.

Mirabilite deposits in Crystal Pit (see map, Fig. 1) occur as massive, efflorescent crusts. The crystals are so small that they are not visible to the eye. Neither crystals nor stalactitic deposits were noticed. The crusts are up to 6 inches thick and often cover areas greater than a square yard. Some of the crusts contain rounded depressions and other irregularities, as though they had been eroded by flowing or dripping water (Figs. 5, 7, 8).

Jarosite ($KFe_3(SO_4)_2(OH)_6$)

This mineral occurs as a widespread, ocherous to dark brown, secondary crust and coating on ferruginous ores and in cracks in the adjoining rocks. It has an appearance similar to that of limonite. It has not been noted from limestone caves or from basaltic lava flows (Palache, Berman, and Frondel, 1951). In Crystal Pit, the mineral was collected as a pale-yellow efflorescence mixed with gypsum, lying loosely on the floor and clinging to the walls (Fig. 5). Conspicuous deposits are found in the northwest corner of the large chamber and in the upper levels of the northwest passage.

Calcite ($CaCo_3$)

Thin, white crusts of what may be this mineral occur as patches on the walls of the large chamber. The material resembles the coralloidal opal of Swartzlow and Keller (1937), but effervesces with the application of hydrochloric acid.

Origin of the Minerals

Stearns (1926-1928) suggests that the gypsum in Crystal Pit may have been deposited from gaseous emanations accompanying the closing phase of the eruption, because gypsum is one of the common salts deposited at volcanic vents, or, alternatively, that sulfur compounds may have reacted at the orifice with calcium compounds in the deposits comprising the vent. Such mechanisms, however, are not likely to have brought about the large deposits of mirabilite because of thermal stability considerations, the limited amount of sodium compounds in the surrounding lavas, and the rarity of sodium compounds in vent emanations. Curiously, Stearns does not mention the presence either of mirabilite or of thenardite in Crystal Pit.

In my opinion, the minerals gypsum and mirabilite probably were deposited from capillary groundwater as secondary crystallization products, rather than from a gaseous carrier. Stearns (1926-1928, p. 75) reports a deposit of thenardite containing traces of K, Ca, Cl, and H_2O from

Indian Tunnel at a spot where water drips on sheltered ledges and evaporates, leaving a fluffy, white flour 1 to 2 inches thick. Likewise, the presence of mirabilite in limestone caves argues against deposition from heated vent gasses. If capillary groundwater supplied the ions for mineral growth, the process undoubtedly would have been very slow and regular, for the mineral deposits include no features suggesting growth rings or layers.

Several facts seemingly support an alternative hypothesis that the gypsum and mirabilite were precipitated from a sulfate-rich lake formerly existing in Crystal Pit. These are: the large size of the gypsum crystals, the absence of gypsum growths like those found in limestone caves, and the fact that both gypsum and mirabilite are precipitated from lakes in arid regions. This explanation is very unlikely, however, because the extremely porous and fractured rocks would not allow a lake to form.

If, as I suppose, the minerals were deposited from capillary groundwater slowly moving into an air-filled Crystal Pit, problems still would remain concerning the source of the minerals, their concentration, and their rates of deposition. The logical source of the minerals is the overlying lava. Recent lava from the Big Cinder Butte flow, some two miles from Crystal Pit, may be chemically similar to that of the Crystal Pit area. Stearns (1930) reports the percent composition of the entire sample of Big Butte lava to be: calcium oxide, 6.5%, sodium oxide, 3.59%; potassium oxide, 2.33%; ferric oxide, 2.15%; ferrous oxide, 12.97%. The sulfur source may have been the 0.15% iron disulfide, the only sulfur compound reported.[1] It seems to me that these quantities are low and that they would not provide a sufficient amount of elements to the groundwater. However, the unweathered lava may contain primary minerals which are sufficiently soluble in rainwater to provide the elements being considered. As the mineralogic composition of the lava is not reported, the solubility of its constituents can only be guessed at.

Another possibility is that these compounds may have been carried up into the cave from buried evaporite deposits by groundwater, but this seems unlikely under the circumstances. Malde (1965) presents a discussion of the stratigraphy of the region and other information pertinent to this problem.

The ages and rates of deposition of cave minerals often are of interest. Seemingly, the unusual minerals in Crystal Pit were deposited relatively rapidly and relatively recently. The appearance, degree of weathering, and vegetal cover of the Crystal Pit flow are very similar to those of the King's Bowl Rift flow, 35 miles to the south. The age of King's Bowl flow has been found to be 2130 ± 130 years before the present (Prinz, 1970), by means of C^{14}-dating of carbonized sagebrush at its base. Tree-ring data from the Monument also suggest that the youngest flows, including the Crystal Pit flow, are 1000 to 2000 years in age (Stearns, 1959 and unpublished).

[1] I have been given figures only for the elements under discussion here.

Fig. 7. Mirabilite on floor of pit. Carbide lamp provides scale.

Fig. 8. Gypsum crystals on wall and covering tips of lava stalactites.

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I wish to thank the following Park Service employees and associates for their aid in and coöperation with my exploration and observations in Craters of the Moon National Monument, and for the use of the Monument Library: William V. Appling, Franklyn N. Hambly, Porter K. Henderson, David C. Ochsner, and Merle E. Stitt. G. W. Moore, U. S. Geological Survey, Menlo Park, California, graciously identified the mirabilite and jarosite.

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ABSTRACTS OF RECENT LITERATURE (Earth Sciences)

Brown, M.C. (1973)—Mass Balance and Spectral Analysis applied to Karst Hydrologic Networks: *Water Resources Research* 9(3):749-752.

Analyses of data on flow volume and flow-through times of subterranean karst streams by cross-variance and by cross-spectral transfer function permits indentification of the systems as unitary or branching and as vadose or phreatic. (EWW)

Dodd, J. Robert; Siemens, Charles T. (1971)—Effect of Late Pleistocene Karst Topography on Holocene Sedimentation and Biota, lower Florida Keys: *Geol. Soc. America, Bull.* 82:211-218.

Filled or nearly-filled sinkholes, very common in the Big Pine Key—Bahia Honda Key area, are the sites of present-day sediment accumulations. Because of the varying thicknesses of the sediments in them, these sinkholes exert a strong influence on vegetation patterns. (EWW)

Baker, Victor R. (1973)—Geomorphology and Hydrology of Karst Drainage Basins and Cave Channel Networks in East-Central New York: *Water Resources Research* 9(3):695-706.

Cave streams form hydrologic links between surface streams and karst springs. They are an integral part of drainage organization in terms of Horton's hierarchial stream numbers, stream lengths, and stream drainage areas. Meandered reaches of free-surface cave streams have wavelengths dierctly related to upstream drainage area and to runoff. Subsurface streams are similar to surface streams both in morphology and hydrology. (EWW)

Bowen, R.; Williams, P.W. (1973)—Geohydrologic Study of the Gort Lowland and Adjacent Areas of Western Ireland using Environmental Isotopes: *Water Resources Research* 9(3):753-758.

Isotopic analysis of H^3 and O^{18} have shown that groundwaters in western Ireland fall into a single group. This indicates good mixing, as is expected to occur in karst regions. Modern levels of tritium indicate relatively rapid ground water circulation. Total replacement of the ground water occurs within a few months to a few years. (EWW)

Brucker, Roger W.; Hess, John W.; White, William B. (1972)—Rôle of Vertical Shafts in the Movement of Ground Water in Carbonate Aquifers: *Ground Water* 10(6):5-13.

Vertical shafts, commonly found in the Interior Lowlands and Appalachian Plateaus provinces, are very short-lived and occur only at the margins of clastic caprocks. They are formed by descending films or sheets of vadose water passing from the surface to master drains. Water takes up $CaCO_3$ and loses CO_2 in travelling from top to bottom of the shaft. (EWW)

Wigley, T.M.L.; Drake, J.J.; Quinlan, J.F.; Ford, D.C. (1973)—Geomorphology and Geochemistry of Gypsum Karst near Canal Flats, British Columbia: *Canadian Jour. Earth Sci.* 10:113-129.

Karst features include springs and related ponors, caves, tufa deposits, geologic organs, and solution breccias. Sinkholes may have a complex origin and are formed preferentially along contacts of the gypsum with other rocks. Statistically, water samples occur in five groups: gypsiferous springs, "downstream" surface waters, "upstream" surface waters, Lussier Valley sinkholes and Coyote Valley sinkholes. These groupings have genetic implications. Results suggest that modern groundwater flow is principally through carbonate rocks. (EWW)

Greeley, Ronald; Hyde, Jack H. (1972)—Lava Tubes of the Cave Basalt, Mount St. Helens, Washington: *Geol. Soc. America, Bull.* 83:2397-2418.

These lava tubes formed about 1900 years ago, chiefly between shear planes of laminar lava flow. Walls between tubes and country rock may be thinner than 25 cm, flows can erode pre-existing surfaces, collapse of roofs can occur after draining of the tubes and before the walls harden, tubes form above topographic lows in the previous land surface, and tubes can be extensively modified by later flows. Roof rupture may form tumuli, some of which collapse to form craters with raised rims. (EWW)

Hill, Carol A. (1973)—Huntite Flowstone in Carlsbad Caverns, New Mexico: *Science*, ns 181:158-159.

Huntite flowstone recently has been discovered in Carlsbad Caverns. This flowstone occurs as a thin, white layer of microcrystals (approximately one to 60 micrometers in diameter) which appears buckled and crinkled. It is believed that the huntite is being precipitated directly from magnesium-rich solutions rather than being formed by alteration of pre-existing minerals. (author)

ERRATA

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

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

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

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

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

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

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

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

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

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

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