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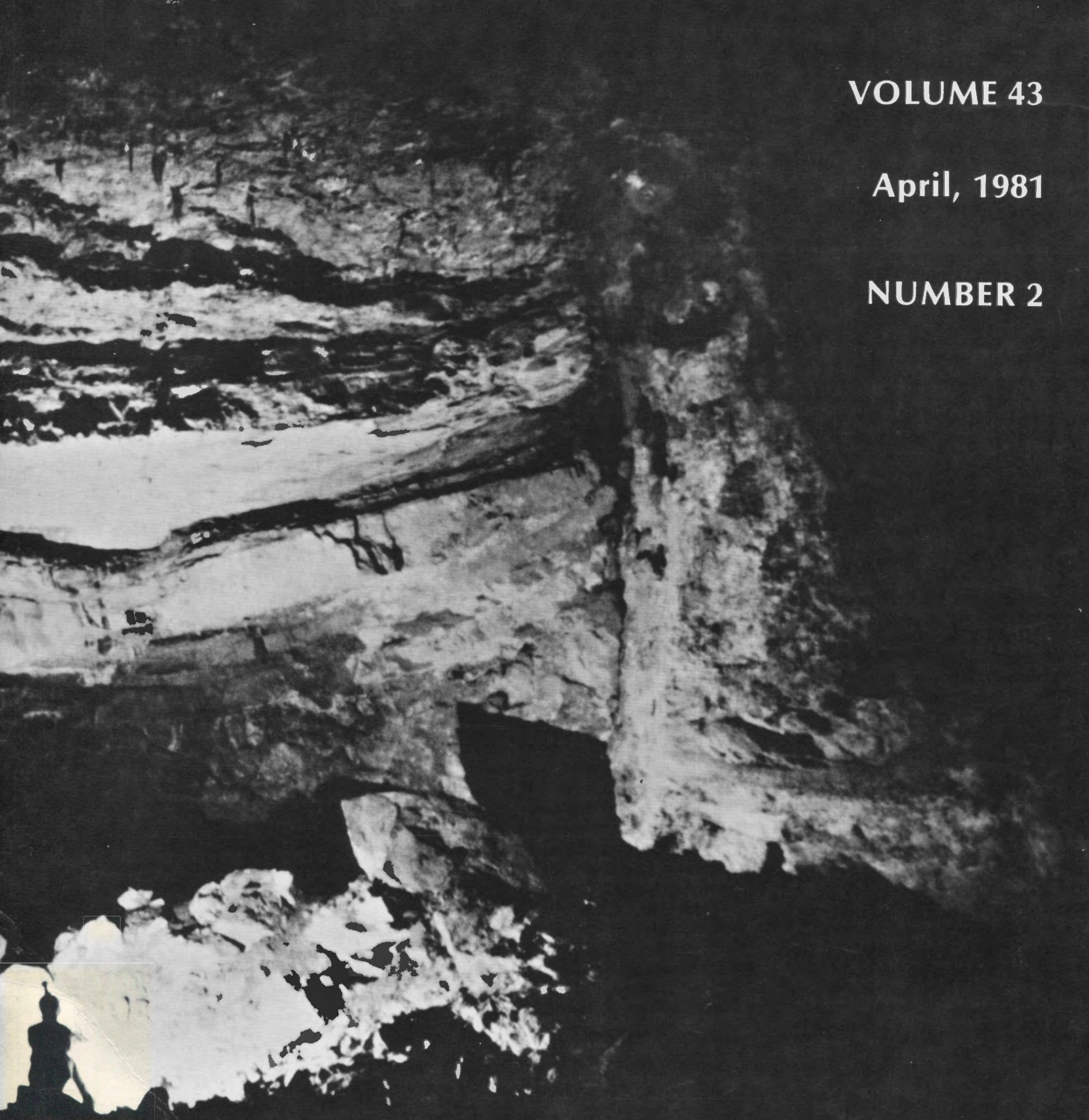

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COVER PHOTO: Upper Kane Cave, Wyoming. Photo by Stephen J. Egemeier.

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CAVERN DEVELOPMENT BY THERMAL WATERS

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SUMMARY

Several caves in Wyoming are forming by a process involving the replacement of limestone by gypsum and then solution of the gypsum. Artesian springs discharge thermal waters containing dissolved hydrogen sulfide into the caves. Atmospheric oxygen in the cave air dissolves into the spring waters and reacts with some of the hydrogen sulfide, producing sulfuric acid. The sulfuric acid reacts with limestone in the stream beds and dissolves it.

Much of the hydrogen sulfide in the spring waters escapes into the cave air. Some of it redissolves in water on the damp cave walls and ceiling where it is oxidized by dissolved oxygen in the water. Sulfur and sulfuric acid are produced. The acid attacks limestone and converts it to gypsum. Gradually a coating of gypsum up to half a meter thick forms on the cave walls and ceiling. Eventually the coating becomes so thick that it cannot support its own weight. As a result, some of the gypsum falls to the cave floor, where it is dissolved and removed by the cave stream. Passages are air-filled during cavern development. Caves that have formed by this process of "replacement-solution" are found in many states in the western United States.

The thermal springs in the caves are depositing metals in hydrogen sulfide-rich muds lining the springs. Small deposits of vanadium and uranium are being formed perhaps in a manner similar to the process that formed the economic deposits of these metals in nearby caves in the Big Horn and Pryor mountains. Other metals deposited include silver, copper, iron, lead, and zinc.

The history and development of some caves will have to be re-studied, as all previous work was done without knowledge of the replacement-solution process. Gypsum crusts, gypsum deposits, and "phreatic" cavern features can be formed by this process in air-filled "vadose" caves.

CAVES CONTAINING flowing thermal springs, although rare, are geologically important for a number of reasons. First, the solutional processes now forming thermal spring caves may have played a role in the development of many caves that no longer contain flowing springs. Second, the study of depositional processes in thermal spring caves may provide information on the formation of some ore bodies.

In the Big Horn Basin area of Wyoming (Fig. 1a), there are many caves. Most of the caves are dead; that is, the processes that formed them are no longer operating. Evidence of the processes that formed these caves is gradually being buried by ceiling collapse and calcite deposition. A few of these dead caves contain uranium ore that apparently was deposited at a time when streams flowed through the caves.

At least three caves in the Big Horn Basin are active; that is, they are being enlarged at the present time. Limestone in two active caves is being dissolved by a process involving sulfuric acid formed by oxidation of hydrogen sulfide. The hydrogen sulfide is discharged into the caves by thermal springs that rise from deep slots in the cave floors. The thermal waters have also deposited uranium- and vanadium-containing precipitates on the cave floors. Active caves such as these are apparently uncommon. However, there seem to be other caves in the Rocky Mountain Region that were formed by this process.

WATER FLOW IN THE BIG HORN BASIN

The Big Horn Basin is a large structural basin in northern Wyoming about 130 km east of Yellowstone National Park. The basin contains a sequence of gently-folded sedimentary rocks ranging from Cambrian to Quaternary in age. Cenozoic and Mesozoic rocks crop out in the basin. Paleozoic rocks crop out along the edge of the basin, and Precambrian rocks crop out in the cores of the mountain ranges surrounding the basin.

The locations of some of the numerous springs in the basin are shown in Figure 1a. Warm springs typically discharge from aquifers exposed in river canyons cut through the cores of anticlines. Nonthermal springs flow from the same aquifers, but at points near the mountain fronts at the edges of the basin.

Table 1 summarizes the water-bearing properties of the rocks in the Big Horn Basin. Inspection of this table reveals that there are two confined aquifers in the basin: the Tensleep sandstone and the Madison limestone. Both are basin-wide in extent and have relatively high permeabilities. As the hydraulic gradient in the Madison is much less than the gradient in the Tensleep, Stone (1967, p. 2096) concluded that the Madison limestone is the better aquifer. All the caves studied are in the Madison limestone.

The Big Horn Basin is very dry; average annual rainfall is less than 20 cm in most parts of the basin. Annual rainfall in the mountains to the west of the basin in Yellowstone National Park is 80 cm. Rainfall at an elevation of 1800 m in the Big Horn Mountains on the east side of the basin is 40 cm annually. Higher elevations probably receive even more rain.

The heavy rainfall in the mountains recharges the Paleozoic aquifers that crop out there. The water then moves under the basin to the springs where it is discharged. A more local source for the spring waters is unlikely, as local catchments above the springs are small, rainfall is intermittent and insufficient to account for the discharges observed, and there is little local surface exposure of the aquifers near the springs to permit recharge.

Figure 1b shows water quality in the Madison limestone using logarithmic Stiff diagrams (Stiff, 1951). Waters on the east side of the basin are generally more dilute than those on the west, indicating more flow on the east side. Therefore, the Big Horn mountains are probably the main source of recharge.

Many of the basin springs are thermal. White (1969, p. 269) defines a thermal spring as any spring 5° C (or more) warmer than the mean temperature of its locale. Waring (1965, p. 4) considers springs at least 15° F (8.3° C) warmer than their locale as thermal, but notes that any spring noticeably warmer than the mean annual air temperature of its locale could be classed as thermal. A comparison of the mean air temperature and spring temperatures shown in Fig. 1a reveals that there are many thermal springs in the Big Horn Basin. Most, as mentioned before, are in breached anticlines.

Although a magmatic source of heat cannot be ruled out, it appears unlikely. No igneous rocks, except Precambrian granitic basement, have been reported in any of the many deep drill holes in the basin. The geothermal gradient in the basin is not unusual except in areas of high water flow.

The warmth of the springs is probably due to deep artesian flow. A structure contour map of the Tensleep sandstone by Zapp (1956) indicates that the top of the Madison limestone is over 2000 m below the surface near the axis and 1000 m below the surface in many parts of the basin. Water in the Madison limestone must flow at these depths to reach the springs.

Rock temperatures at 1000 m would be 37° C and at 2000 m would be 67° C, assuming the lowest temperature gradient measured in the basin (Fig. 1b), 34 m/°C. As rock temperatures at depths where flow is likely to occur are higher than the spring temperatures, the warmth of the springs is most likely due to deep artesian flow.

KARST FEATURES AND THEIR RELATIONSHIP TO THE CAVES

Surface karst features are rare near the caves. Karst features are also rare in other limestone outcrop areas of the Big Horn and Pryor Mountains. The few sinkholes in the basin near the caves and in the mountains are clearly collapse features. No solution sinks are known in or near the basin, even though solution sinks are common in most karst areas. Other features common in karst areas such as karren, fissures, fluted limestone, and sinking streams are also very rare in the Big Horn Basin area. This is probably due to the low rainfall, which has limited the 'karstification' of the limestone. The caves show no relationship to surface karst development; in fact, they appear to be independent of it. This is logical, as the water enlarging the caves does not come from directly above the caves but by artesian flow from recharge areas miles away.

Shales overlie and underlie the Madison limestone (see Table 1), and substantial artesian pressures are found in the Madison in oil fields (Stone, 1967, p. 2093). Many of the springs that rise from the Madison are artesian.

The source of the springs is probably the mountains around the basin. Tilt of the oil/water interface in Big Horn Basin Tensleep oil pools is interpreted to be the consequence of hydrodynamic flow (Hubbert, 1966, p. 2516). Bond (1972, p. 10) pointed out, however, that in aquifers containing density-stratified brines, tilted interfaces can be maintained with zero flow. Water in the Madison limestone is much less saline than the water studied by Bond. In fact, potable water is obtained from many wells; hence, it is unlikely that the tilts are due to salinity stratification. This supposition is borne out by published chemical analyses which do not show a salinity (and hence, density) stratification of water in the Madison limestone. Thus, Hubbert's explanation is preferred. The tilt then should be in the same direction as the water flow. The amount of tilt depends upon aquifer permeability, oil gravity, and flow rate. From the tilts shown in Fig. 1a, it is apparent that flow in the Tensleep sandstone is from the perimeter, particularly the eastern perimeter, of the basin toward the springs in the east-central part of

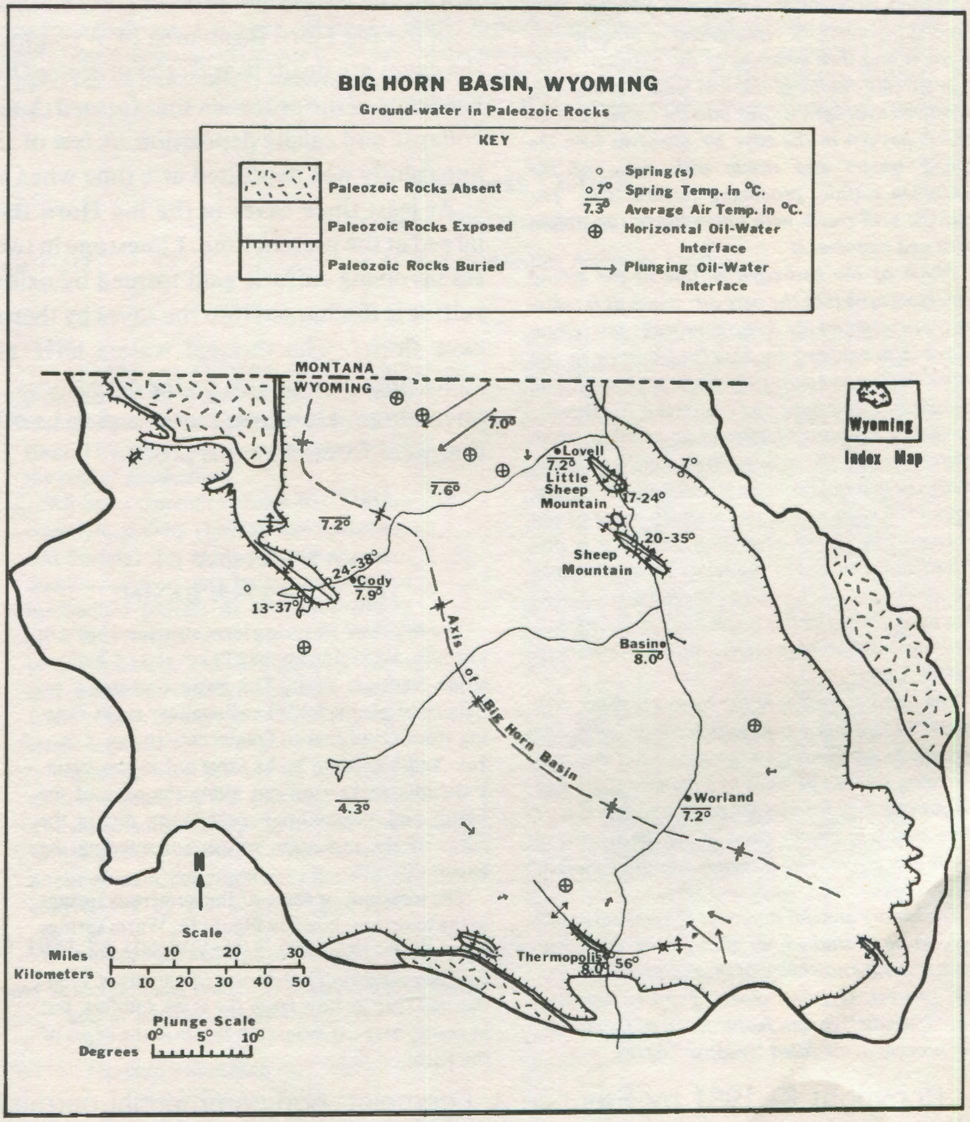

Figure 1a. Spring temperatures this study, except near Cody (Love and Good, 1970, pp. B19-21) and Thermopolis (Waring, 1965, p. 50). Mean air temperatures from Anon. (1965). Plunge on oil-water interface compiled from Zapp (1956) and Summerford (1952) pp. 99-102. Geology from Love, *et al.* (1955).

the basin. As the Madison limestone has the same general catchment and discharge areas as the Tensleep sandstone, its flow pattern is probably analogous. Water in the Madison is least saline near the eastern side of the basin (Fig. 1b), indicating fresh water recharge along this side of the basin. There apparently is less recharge to the Madison on the west side of the basin, perhaps because it is partially covered by volcanics. Figure 2 shows the general flow-path of water in the Madison limestone.

THE CAVES

Although a number of caves are mentioned, this report is chiefly a study of five caves. Four of these, Lower Kane, Upper Kane, Hellespont, and Salamander caves, are in the Little Sheep Mountain Anticline (see figs. 1a and 3). The fifth and largest cave, Spence Cave (figs. 1a and 8) is in the Sheep Mountain Anticline. With the exception of Lower Kane and Hellespont caves, they do not appear to be particularly noteworthy. Lower Kane and Hellespont caves, however, are very unusual caverns in several respects: 1) they smell of hydrogen sulfide; 2) they are uncomfortably hot and humid; 3) their walls are covered with a soft, wet gypsum crust (in places itself encrusted with sulfur crystals); and 4) large mounds of gypsum are present on the cave floors.

Lower Kane, Hellespont, and Salamander caves contain flowing water; Upper Kane and Spence caves do not. Salamander Cave contains the largest stream: an estimated 5000 l/min. of warm water emerges through breakdown that plugs the cave less than 20 m from its entrance. Because of this plug, little is known about Salamander Cave. The streams in Hellespont Cave and Lower Kane Cave are also fed by thermal springs.

The dead (streamless) caves apparently were once active caves. Collapse has altered, destroyed, and buried many of the features that were formed while the caves were active. Many features similar to those of the active caves are preserved in places in the dead caves, suggesting that the origins of all the caves are similar.

CAVE WATER AND AIR

The air in Lower Kane and Hellespont caves was analyzed *in situ*. The constituents and properties of the waters of these caves and in Salamander Cave were determined by *in situ* analyses and by laboratory analyses of samples.

Sampling and analytical methods

In general, standard methods were used; Table 2a lists the methods used and gives references for most of them. The standard methods for sulfate were found to be unsuitable; sulfate was determined on cation-exchanged samples by precipitation with barium chloride and titration of excess barium with CDTA.

Water analyses involved several *in situ* determinations, and at least two samples were taken at each location. At each sampling point, a one-liter raw sample and a 250 ml acidified sample were collected for future analysis. In some cases, duplicate samples were taken. (Chloride, nitrate, phosphate, silica, and sulfate were determined in the raw samples, and metal cations were determined in the acidified samples.) *In situ* determinations were made of water characteristics likely to change in a sample bottle after collection. At the time of water sampling, bicarbonate, pH, and temperature were determined; conductivity, hydrogen sulfide, and dissolved oxygen were determined *in situ* at another time. Several duplicate samples were taken that differed as much as a week in the time of sampling. For most of these duplicate samples, analytical results differed less than 5% for a pair. In many cases, this is within the analytical error which is given in Table 2b. Thus, sampling errors even for samples taken at different times are not significant.

An Edmund Scientific Corporation air pollution analyzer was used to determine the concentrations of a few gases in the cave atmosphere. The analyzer consists of a small hand pump which draws air through treated ampules and is similar to one described by Littlefield, *et al.* (1935) for determination of hydrogen sulfide concentration in air.

Gases in Cave Air

The air in the active caves (Lower Kane Cave and Hellespont Cave) contains hydrogen sulfide gas, concentrated enough to cause nausea and headache in a few hours.*

Results of analyses of cave air are listed in Table 3. The hydrogen sulfide concentrations found agree very well with estimates based on the strength of the odor. The gas concentration de-

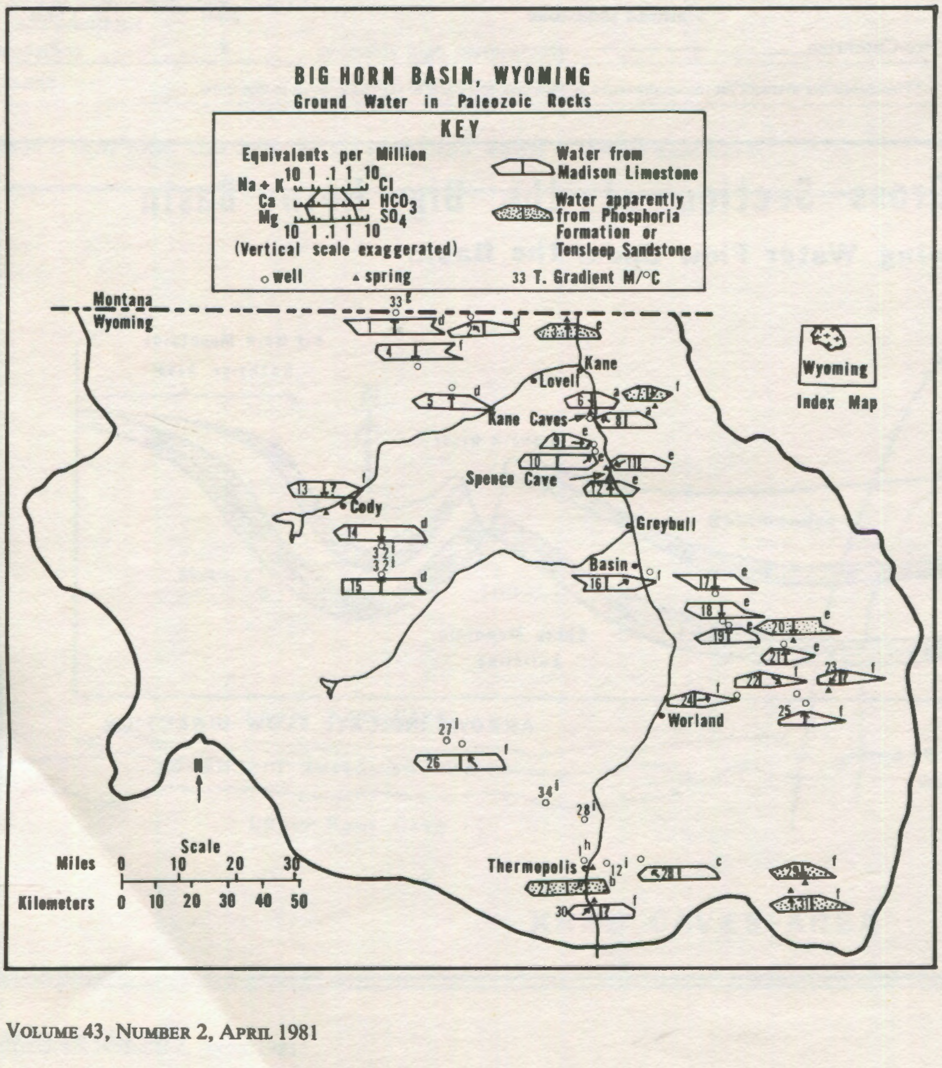

Figure 1b. Chemical analyses: **a.** this report. **b.** Burke (1952), p. 92. **c.** Crawford (1940), p. 1304. **d.** Crawford (1949) pp. 274-277. **e.** Shell Oil Company. **f.** Lowry and Lines (1972) pp. 12-14. Thermal gradients: **g.** Espach and Fry (1951) pp. 10-11. **h.** Burke (1952) p. 92. **i.** Van Ostrand (1934) p. 1021. Arrows within Stiff diagrams point to sample location. Question marks within Stiff diagrams indicate source of water is uncertain.

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*Hydrogen sulfide is poisonous. Exposure to the gas can produce nausea, headache, and death. Symptoms appear in a few hours or less in Hellespont Cave and after less than eight hours of exposure in Lower Kane Cave. Anyone entering these caves should consider these risks.

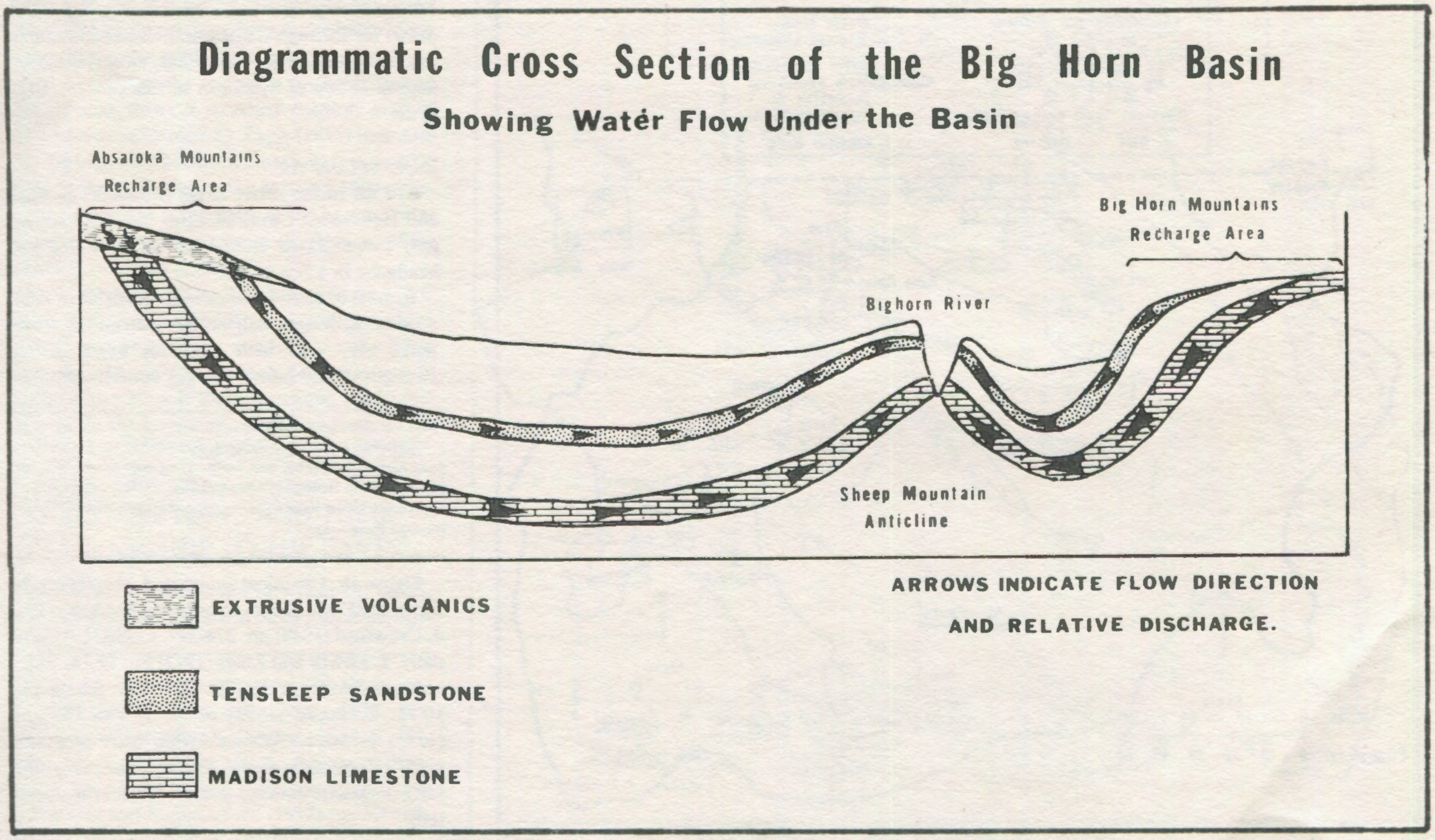
creases both upstream and downstream of the springs, which are the apparent source of the gas. The decrease downstream could be due to oxidation of the gas by the reaction $H_2S + 2O_2 \rightleftharpoons H_2O + SO_3$, diffusion of the gas out of the cave, or dilution of cave air with outside air. Upstream of the springs, dilution is not possible, so the decrease must be due to oxidation of the gas. The concentration of carbon dioxide is apparently uniform above the Upper Spring, but decreases downstream from the spring. This gas must be brought in by water entering the cave or be produced in the cave, as the proportion of carbon dioxide in the cave air significantly exceeds that in the atmosphere. Analysis of the spring water indicates that it is not losing carbon dioxide. Therefore, the gas must be being produced in the cave or be coming from seepage water entering the cave. The concentration of carbon dioxide is dependent upon rates of production in the cave, escape from seepage water, dilution of cave air by air from outside the cave, and diffusion of the gas out of the cave. If carbon dioxide is produced in the cave, then the highest carbon dioxide concentrations should be, and are, in the areas farthest from the entrance.

If the carbon dioxide came from seepage water, the greatest concentration would be near the one area of seepage. The concentration near this point is 0.05%, which is equal to the lowest measure-

Table 1. Water-bearing properties of Big Horn Basin rocks.

System	Name	Thickness*	
		M	Ft
Triassic	Chugwater formation	120-300	400-1000
Permian	Phosphoria formation	80-6	250-20
Pennsylvanian	Tensleep formation	110-30	350-100
	Amsden formation	80-45	260-150
Mississippian	Madison limestone	150-250	500-800
Devonian	Three Forks and Jefferson formations	0-90	0-300
Silurian	(absent)	0	0
Ordovician	Bighorn dolomite	0-140	0-450
Cambrian	Gallatin formation	120-150	400-500
	Gros Ventre formation	180-150	600-500
	Flathead sandstone	50-0	170-0
Pre-Cambrian	———	?	?

*Thicknesses are averages for the southwestern and northeastern parts of the basin, given in that order.



Diagrammatic Cross Section of the Big Horn Basin

Showing Watér Flow Under the Basin

ARROWS INDICATE FLOW DIRECTION AND RELATIVE DISCHARGE.

Table 1. (continued)
Stratigraphy after Stone (1967) pp. 2061-2069, and Thomas (1965) pp. 1867-1872.

Lithology	Permeability	Permeability reference
red siltstone, red shale and evaporites	tight	Stone (1967) p. 2069
red shale, phosphorite and limestone	16 millidarcys	McCaleb and Willingham (1967) p. 2124
	10 millidarcys	Stone (1967) p. 2066
sandstone and dolomite	few-800 millidarcys	Stone (1967) p. 2065
	0.1-600 millidarcys	Coe (1967) p. 2119
red shale, sandstone, and dolomite	lack of porosity, poor (oil) production	Stone (1967) p. 2065
limestone and dolomite	transmissibility uniformly high	Stone (1967) p. 2096
limestone and green shale	produces oil shale may reduce vertical permeability	Stone (1967) p. 2063
finely crystalline dolomite (cherty)	produces oil	Stone (1967) p. 2063
limestone and dark green shale	probably low permeability	———
glauconitic and sandy limestone, interbedded shale and sandstone	probably low permeability	———
sandstone	probably high permeability	———
granite	probably low permeability	———

ments in the cave. Therefore, seepage water is not a likely source of carbon dioxide.

The highest H_2S concentration in the caves is near the thermal springs, which suggests that the springs are the source of the gas. The water discharged by the springs contains dissolved hydrogen sulfide. Hydrogen sulfide is fairly common in some limestone ground water (Gardner, 1935) and has also been reported in water-filled Florida caves by Summers (1972, p. 21). In the Florida caves, the gas seems to be produced by the decay of organic debris that has fallen into stagnant water. As far as can be determined, these Wyoming caves are the only air-filled caves reported to contain an appreciable quantity of hydrogen sulfide.

Humidity

The humidity in the active caves is 95 to 100%, and in the dead caves it is 84% in Upper Kane Cave and even less in Spence Cave.

Temperature

Cave temperatures are typically close to the mean air temperature near the cave. A few exceptions are known, however, such as Warm River Cave, Virginia (Douglas, 1964, pp. 81-84) and Piercys Cave, West Virginia (Waring, 1965, p. 45). The average air temperature of the Big Horn Basin (Fig. 1a) is 7 to 8° C; all the caves studied are significantly warmer than this (Fig. 3).

The only known source of heat in the basin is

Figure 2. *(left)* Water flow under the Big Horn basin. **Figure 3.** *(below)* Kane Caves.

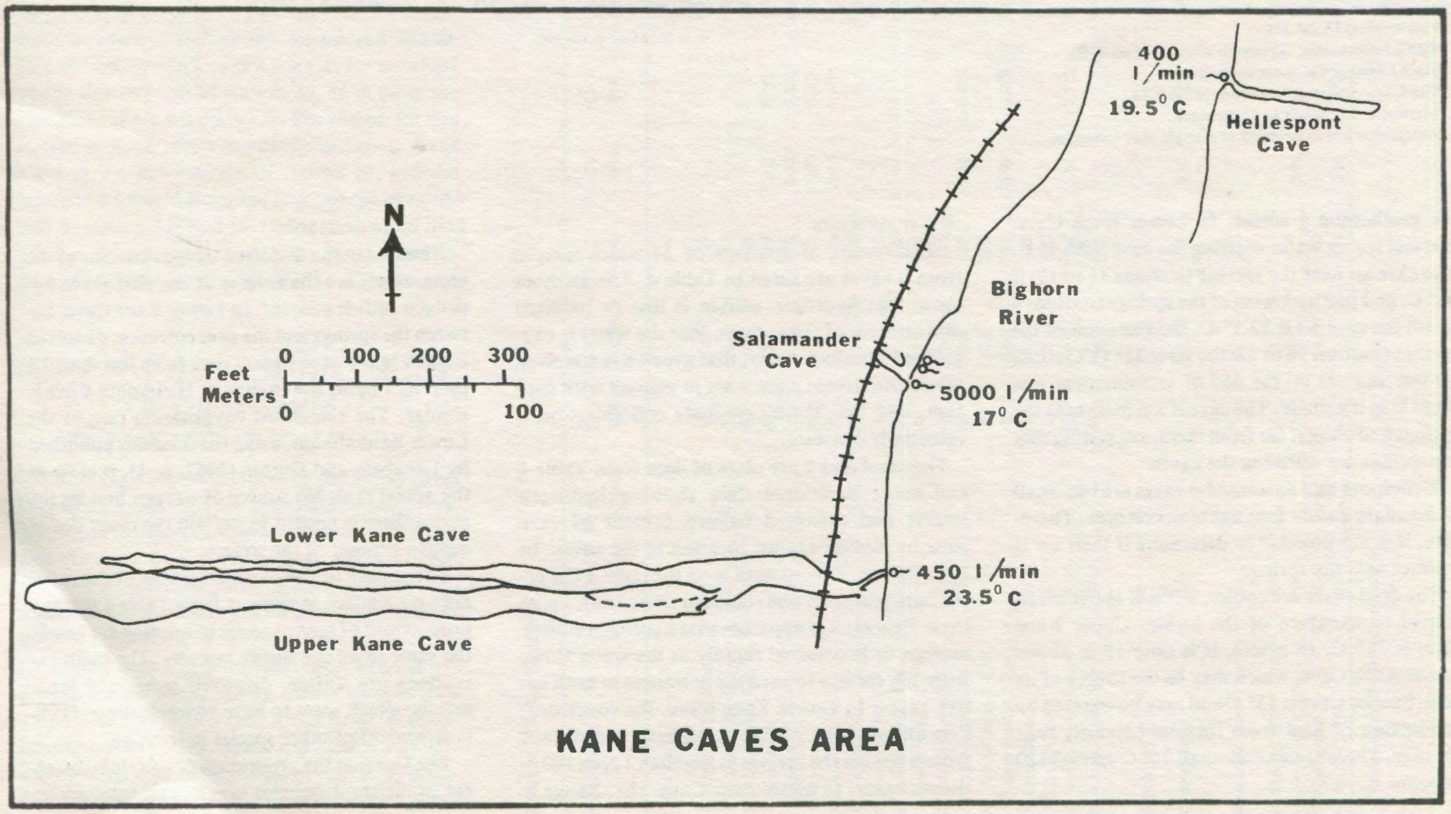


Table 2a. Analytical Methods.

Determination	Method	References
Ag	Atomic absorbtion[1] (extraction required)	Platte (1968)
Ca	Atomic absorbtion[1]	Brown, *et al.* (1970)
Cl	Mohr titration	Brown, *et al.* (1970)
Conductance	Conductance bridge[2, 6]	
Co	Atomic absorbtion[1] (extraction required)	Platte (1968)
Cu	Atomic absorbtion[1]	Platte (1968)
Fe	Atomic absorbtion[1] (extraction required)	Brown, *et al.* (1970)
HCO_3	Electrometric titration[6]	Barnes (1964)
H_2S	Lead acetate test paper[3, 6]	APHA (1971)
K	Flame photometry[4]	Brown, *et al.* (1970)
Li	Atomic absorbtion[1]	Brown, *et al.* (1970)
Mg	Atomic absorbtion[1]	Brown, *et al.* (1970)
Na	Flame photometry[4]	Brown, *et al.* (1970)
NO_3	Chromatropic acid[3]	APHA (1971)
O_2	Winkler (azide modification)[6]	Slack (1965)
pH	pH meter and pH 4 and 7 buffers[6]	Barnes (1964)
PO_4	Phosphomolybdate[3]	Rainwater and Thatcher (1960)
SiO_2	Molybdosilicate[3]	APHA (1971)
SO_4	see text	
Sr	Atomic absorbtion[1]	Brown, *et al.* 1970)
Temperature	Mercury thermometer[5, 6]	

[1] Perkin-Elmer Model 303.

[2] BETZ Laboratories "Conducto Bridge" model X-50.

[3] Hach Chemical Co. water analysis kit.

[4] Baird Atomic flame photometer model KY2.

[5] Matheson -10° to 100°C, 1° divisions.

[6] Analyzed *in situ* as promptly as possible after collection.

Table 2b. Analytical Error of Water Samples.

Determination	Inter-laboratory Results			This Study	
	σ	R.E.*	Source	σ	Drift
Ag	17.5%	10.6%	APHA (1971)		5%
Ca	3.5%		Brown, *et al.* (1970)	2.5%	3%
Cl	4.2%	2.2%	Hume (1967)		
	4.4%		Brown, *et al.* (1970)		
Conductance	1.6%		Brown, *et al.* (1970)		
	5 %		APHA (1971)		
Co					2%
Cu	11.2%	3.4%	APHA (1971)		5%
Fe	16.5%	0.6%	APHA (1971)	1.5%	6%
	6.6%		Brown, *et al.* (1970)		
HCO_3				1.4%	
H_2S				20-30%	
K	15.5%	2.3%	APHA (1971)	2.1%	4%
	15.2%	3 %	Hume (1967)		
Li	5.2%		Brown, *et al.* (1970)	0.5%	4%
Mg	10.5%	6.3%	APHA (1971)	1.7%	10%
	15 %		Brown, *et al.* (1970)		
Na	17.3%	4.0%	APHA (1971)	0.6%	1%
	9.2%	0	Hume (1967)		
NO_3	61.4%	12.5%	APHA (1971)		
O_2	2 %		Burke (1962)		
pH	0.13 pH		APHA (1971)		0.04 pH
	0.12 pH		Brown, *et al.* (1970)		
PO_4	20.8%	1.2%	APHA (1971)		
SiO_2				2.4%	
Sr	10 %		Brown, *et al.* (1970)	1.7%	2%
Temperature	3°C		APHA (1971)		

σ = Standard deviation. *R.E. = Relative Error

the geothermal gradient. In Lower Kane Cave, thermal spring water entering the cave is 23.5° C. The cave air near the springs (stations 11 to 19) is 23° C, and just upstream of the springs (stations 4 to 10) the cave air is 22.5° C. Downstream of the springs (stations 20 to 22) the air is 22° C. Outside air temperature on the day of measurement was 30.5° C in the shade. The cave is warmest near the springs and coolest far from them, suggesting that the springs are warming the caves.

Hellespont and Salamander caves are too small to maintain steady internal temperatures. Therefore, it is not possible to determine if their air is warmer near the springs.

The dead caves are cooler, but still above mean annual temperature of the locale. Upper Kane Cave is 21° C. In places, it is only 15 m above Lower Kane Cave, which may be the source of its heat. Spence Cave is 13° C and may be warmed by conduction of heat from thermal conduits near the cave. There are several small 34° C springs 120 m below it.

Water Analyses

The results of analyses of 24 water samples from 3 caves are listed in Table 4. The analyses show that hydrogen sulfide is lost or oxidized downstream of the springs, that the water is oxygenated simultaneously, that gypsum is dissolved where the stream waters are in contact with gypsum, and that the bicarbonate concentration is essentially constant.

Figures 4 and 5 are plots of data from Table 4 and some additional data showing hydrogen sulfide and dissolved oxygen content of cave streams plotted against location in the caves. In both figures, downstream is on the right. (The error bars represent one standard deviation). From these figures, it is apparent that hydrogen sulfide escapes or is oxidized rapidly as the water flows from the springs toward the entrances in both active caves. In Lower Kane Cave, the concentration of hydrogen sulfide in the cave stream falls from 6 ppm at the springs to less than 1 ppm 100 m downstream. In Hellespont Cave, the change is even more dramatic.

Changes in the dissolved oxygen content of the cave waters are the reverse of the changes in hydrogen sulfide content. In Lower Kane Cave, between the springs and the cave entrance, dissolved oxygen in the cave stream rises from less than 0.1 ppm to 5 ppm; the change in Hellespont Cave is similar. The calculated oxygenation rate of the Lower Kane stream, using the relations published by Langbein and Durum (1967, p. 1), is close to the actual rate. No source of oxygen besides the atmosphere is needed to explain the rapid rise of oxygen content of the stream.

The waters of the various springs in the caves are very similar. It appears from Table 4 that the proportions of many species in solution are nearly the same in all the spring samples. The main exceptions are sulfate, dissolved gases, and trace metals, which seem to have concentrations fairly independent of other species in solution.

The fact that the proportion of sulfate is different in different samples suggests that the waters

Table 3. Analyses of cave and outside air for some gases. Station locations are shown in figures 6 and 7.

Sampling point	CO_2 %	H_2S* ppm	SO_2† ppm
Lower Kane Cave			
Station 24	0.03	0.x	0
Station 22	0.05	0.x	0
Station 20		0.x	
Lower Spring	0.05	3	0
Upper Spring	0.14	2	0
Station 2	0.14	0.x	
Hellespont Cave			
Station H4	0.05	7	0
Station H5	0.09	11	

*Detection limit is 1 ppm. A hydrogen sulfide odor is present at the outside sampling point and everywhere within both caves. Therefore, the H_2S concentration must be >0.1 ppm, according to Littlefield, *et al.* (1935).

†Detection limit is 0.5 ppm.

are not saturated with respect to gypsum. Indeed, this is the case; a few samples are close to 10% of saturation, but the remainder are less saturated. Large blocks of gypsum (about 20 kg) placed in the cave streams dissolve completely in a week.

The sulfate and bicarbonate content of cave waters versus location is plotted in figures 4 and 5. The bicarbonate content is essentially constant in both Lower Kane and Hellespont caves, except possibly near the entrances to both caves. Apparently, carbonic acid does not play a significant role as far as enlargement of these caves is concerned. Sulfate content of stream water is essentially constant in Hellespont Cave, but rises rapidly downstream in Lower Kane Cave everywhere except downstream of station 20, 45 m from the entrance, where it is constant. The reason for this is readily apparent; everywhere in Lower Kane Cave except downstream of station 20, gypsum mounds are found along the edges of the streams. Below station 20 and in Hellespont Cave, the streams flow on mud, are not in contact with gypsum, and therefore are not able to dissolve gypsum.

AN HYPOTHESIS FOR THE ORIGIN OF THESE CAVES: REPLACEMENT-SOLUTION

The many unusual features of the active caves suggest an hypothesis for the origin of the caves. Hydrogen-sulfide-bearing waters are discharged into the caves by thermal springs. Oxygen from the cave air dissolves into the cave streams, oxidizes some of the hydrogen sulfide, and produces sulfuric acid. The acid reacts with the limestone in the stream beds and dissolves it. Much of the hydrogen sulfide in the spring waters escapes into the cave air. Some of it redissolves in water droplets on the cave walls, where it is oxidized by dissolved oxygen in these droplets. Sulfur and sulfuric acid are produced. The acid attacks lime-

Table 4. Water Samples.

No. Location	Date 1970	Ag ppm	Ca ppm	Cl ppm	Cond. Mhos	Co ppm	Cu ppm	Fe ppm	HCO_3 ppm	H_2S ppm	K ppm	Li ppm	Mg ppm	Na ppm	NO_3 ppm	O_2 ppm	pH	PO_4 ppm	SiO_2 ppm	SO_4 ppm	Sr ppm	Temp. °C	epm Balance†
Lower Kane Cave																							
1. Upper Spring 0 m*	7/3		90	12.3	550			0.0	213	6	3	0.4	29	11		0.0	7.00			182	1.0	24	+0
2. Upper Spring 0.3 m	7/3	0.00	98	12.5	550			0.0	215	6	3	0.4	29	11		0.0	6.90			184	1.0	23.5	+2
3. Upper Spring 1 m	7/3		97	12.5	550			0.0	218	6	3	0.4	31	11		0.0	6.90			189	1.0	23.5	+2
4. Station 19	7/6		107	12.5	525	0.00	0.003	0.0	213	1	3	0.4	33	11		6.7	7.10	0.2		245	1.0	23.5	−2
5. Station 15	7/6		110	12.5	525	0.00	0.003	0.0	213	4	3	0.4	30	11		1.3	6.90			229	1.0	24	+2
6. Station 17	7/6		108	12.5	525	0.00	0.001	0.0	215	1	3	0.4	33	11		4.1	7.10		15	242	1.0	23.5	−1
7. Station 23 ent.	7/8		99	12.5	520	0.00	0.002	0.0	213	0.4	3	0.4	32	11		6.5	7.30	0.3	15	223	1.0	23.5	−1
8. Station 20	7/8	0.00	114	12.5	525			0.1	211	0.5	3	0.4	31	11		5.2	7.10			223	1.0	23.5	+4
9. 5 m N of Sta. 19	7/8		99	12.8	525			0.0	215	1.5	3	0.4	31	11	0.0x	3.0	7.30	0.2	15	201	1.0	23.5	+0
10. Lower Spring 0 m	7/16		96	12.5	525			0.0	218	5	3	0.4	32	11		0.0	7.25			195	1.0	23.5	+1
11. Lower Spring 0.3 m	7/16	0.00	104	13.5	525			0.0	215		4	0.4	28	11		0.0	7.25			195	1.0	23.5	+2
12. Lower Spring 1.2 m	7/24		104	12.0	525			0.0	218	5	3	0.4	31	11		0.0	7.10			218	1.0	23.5	−1
13. Upper Spring 0 m	7/24		101	13.0	525	0.00	0.002	0.0	236	6	3	0.4	33	11		0.0	7.10	0.3		189	1.0	23.5	+3
14. Upper Spring 0.6 m	7/24		98	12.5	525			0.0	220	6	3	0.4	32	11		0.0	7.10		15	195	1.1	24	+2
15. Upper Spring 1.2 m	7/24		89	12.0	525	0.00	0.002	0.0	213	6	3	0.4	31	11		0.0	6.80	0.2	15	195	1.0	23.5	−0
16. Dead Spring	7/28		140	14.5	695			0.0	195	5	4	0.5	34	10			7.25		17	319	1.2	23.5	+2
17. Outlet Spring	7/28		104	12.5	520			0.0	229	0.0x	4	0.4	36	11		5.0	7.20		15	223	1.0	23.5	+1
18. Lower Spring 0.9 m	7/29		101	12.5	525	0.00	0.003	0.0	213	5	3	0.4	29	10		0.0	6.80		15	203	1.0	24	−0
19. Lower Spring 0 m	7/29		102	12.5	525			0.0	218	5	3	0.4	30	11		0.0	6.80		15	192	1.0	23.5	+3
Salamander Cave																							
20. Entrance	7/28		60	5.5	330	0.00	0.002	0.0	207	0.0x	1	0.2	26	6		3.3	7.20		11	105	0.6	17	−3
Hellespont Cave																							
21. Station H2	8/4		116	19.5				0.0	218	2	4	0.6	35	15		3.8	7.30		16	275	1.3	19	−4
22. 2 m W of H3 ent.	8/4		122	18.5		0.00	0.03	0.0	236	8	4	0.6	32	15		0.0	7.25		16	275	1.2	19.5	−2
23. Spring 0 m	8/4	0.00	119	18.5	600			0.0	236	13	4	0.6	39	15		0.0	7.15		16	269	1.2	19.5	+0
24. Spring 0.75 m	8/4		122	19.5	600	0.00	0.02	0.0	229	15	4	0.5	36	16	0.0x	0.0	7.20	0.2	17	275	1.3	19.5	−0

*Sampling depth (Below water surface).

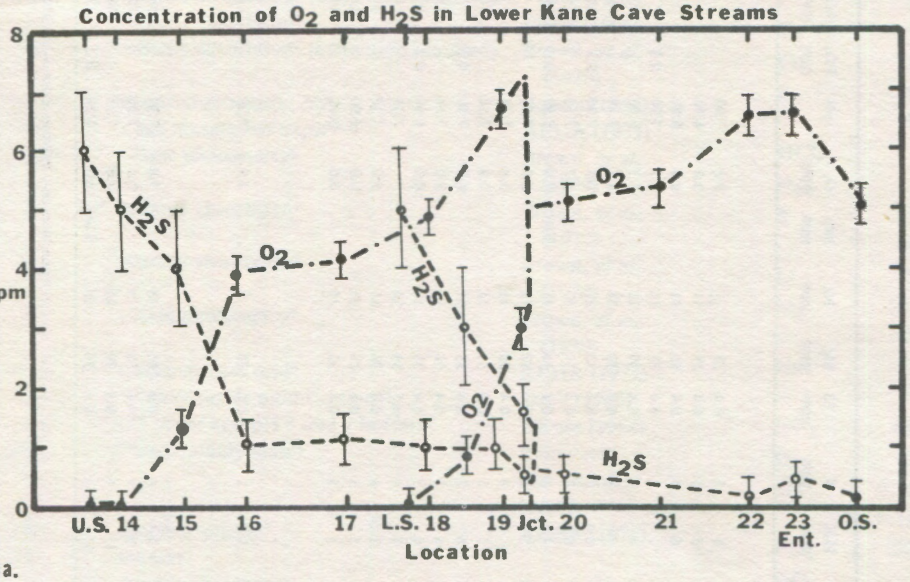
†epm balance = 100 (epm cations-epm anions)/([epm cations + epm anions]/2).

stone and converts it to gypsum, forming a gypsum crust on the cave walls and ceiling. Eventually the crust becomes too thick to support its own weight, and some of it falls to the cave floor, where it forms gypsum mounds. These mounds are dissolved and removed from the cave where they are in contact with cave streams. The passages lengthen headward and enlarge at the same time.

The oxidation of hydrogen sulfide by oxygen to sulfur occurs readily. The reaction

$$(1)\quad 2H_2S_{(aq)} + O_{2\,(aq)} \rightleftharpoons 2S_{(s)} + 2H_2O_{(l)}$$

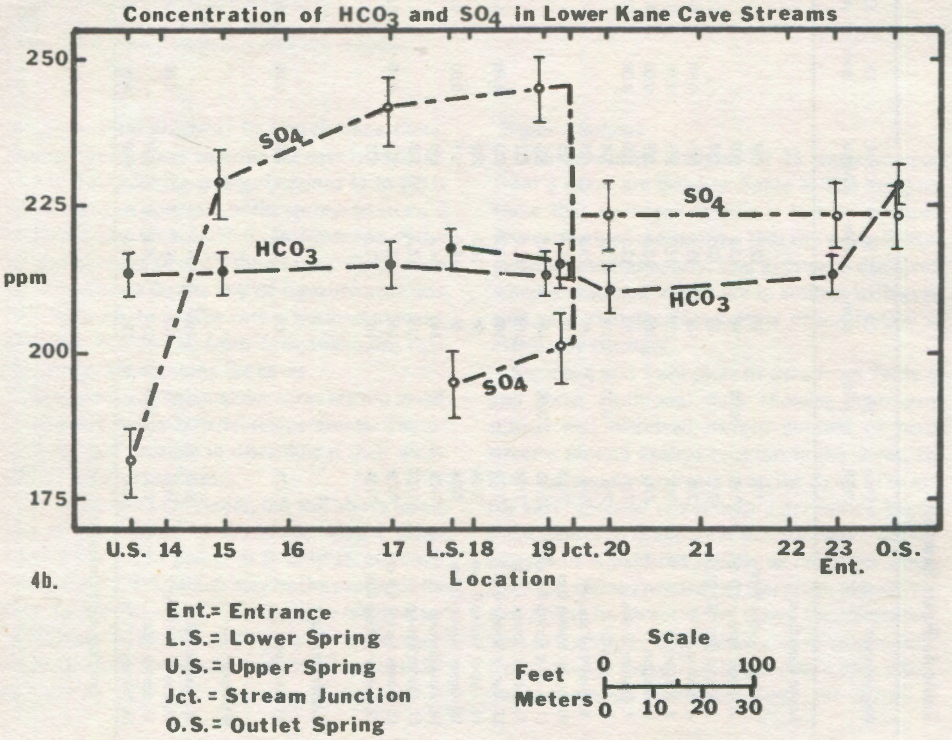
has a standard free energy change of -100 kcal/mole, which indicates that given enough time the reaction will proceed to the right. This prediction has been confirmed by Davis, *et al.* (1970, p. 2445), who found that the reaction produced visi-



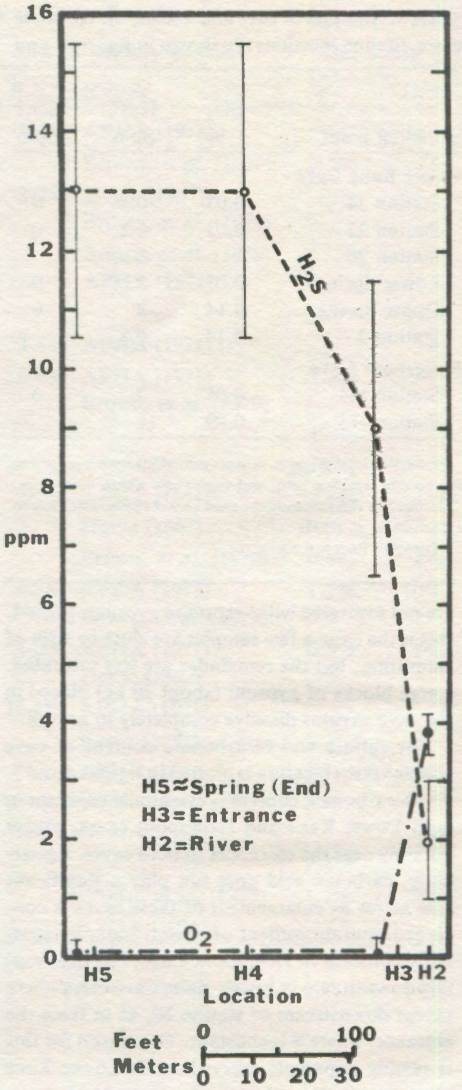
4a.

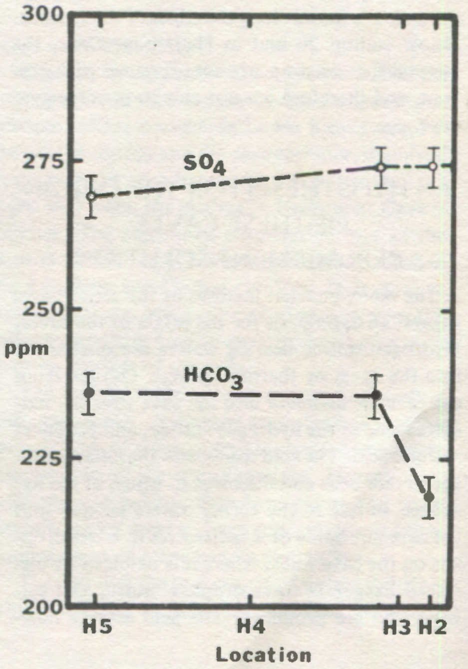
Figure 4a. *(above)* Concentration of O_2 and H_2S in Lower Kane Cave Streams. **Figure 4b.** *(below)* Concentration of HCO_3 and SO_4 in Lower Kane Cave Streams.

Figure 5a. *(right, top)* Concentration of H_2S and O_2 in Hellespont Cave Stream. **Figure 5b.** *(right, bottom)* Concentration of HCO_3 and SO_4 in Hellespont Cave Stream.



4b.





ble sulfur crystals in an hour at 70° C. It is quite reasonable to expect this reaction to occur in the cave.

The oxidation of sulfur to sulfate also occurs readily. The reaction

$$(2)\ 2H_2O_{(l)} + 2S_{(s)} + 3O_{2(aq)} \rightleftharpoons 2HSO^-_{4(aq)} + 2H^+_{(aq)}$$

has a standard free energy change of -246 kcal/mole, which indicates that the reaction will proceed to the right given enough time. According to Lundgren (1971, p. 41), iron-oxidizing bacteria can act as a catalyst in this reaction.

The replacement of limestone by gypsum proceeds according to the reaction

$$(3)\quad H^+_{(aq)} + HSO^-_{4(aq)} + CaCO_{3(s)} + H_2O_{(l)} \rightleftharpoons CaSO_4 \cdot 2H_2O_{(s)} + CO_{2(g)}$$

which has a standard free energy change of -17.5 kcal/mole. Dolomite is converted by a similar reaction to gypsum and epsomite. Reaction 3 goes very rapidly to the right at room temperature.

The reactions described are the basic reactions causing enlargement of the active caves. The oxidation of hydrogen sulfide may proceed in more than the two steps suggested; however, intermediate products other than sulfur, such as sulfur dioxide and thiosulfate, have been tested for and have not been detected. Other ions such as HS^- and $SO_4^=$ may take part in reactions as well as or instead of H_2S and HSO_4^-; all the reactions would still proceed as written and the products would be essentially the same.

Durov (1956) proposed that a significant part of the solutional development of some caves in the U.S.S.R. was due to oxidation of pyrite, which produces sulfuric acid. Eight samples of Madison limestone from the caves had mean iron content of 0.09%. The highest value was 0.18%. Thus, the iron sulfide content of the limestone is fairly small.

Cavern development has generally been attributed to the solution of limestone or dolomite by carbonic-acid-bearing cold vadose and/or phreatic underground waters. The caves studied seem to be formed by another process involving the replacement of limestone and dolomite by gypsum and epsomite and then solution by cave waters. Because the first process only involves direct solution of soluble bedrock, I have chosen to call it "simple-solution." As the second process, proposed here, involves first replacement, then solution, I have named this process "replacement-solution." (Park and MacDiarmid [1964, p. 109] state: "Replacement ordinarily implies little or no change in the volume of the replaced rock, although in some rocks considerable shrinkage or expansion takes place." According to this definition, the process described above is replacement even though there is a volume increase.) Caves formed by either simple- or replacement-solution processes are solution caves.

The hypothesis of replacement-solution explains the chemistry of the cave air. The decrease in concentration of hydrogen sulfide in cave air away from the springs is due to reactions 1 and 2. The carbon dioxide present in the air is produced by reaction 3.

This hypothesis also explains the observed changes in the cave water. Hydrogen sulfide in the cave streams decreases downstream of the springs (figs. 4 and 5) due to loss of the gas into the air and reactions 1 and 2. Constant bicarbonate content and a rise in sulfate content (Fig. 4) indicate that the gypsum is dissolving, thus enlarging the cave.

The physical aspect of cavern development is not as clear as the chemical process involved. The process does seem to be a simple one. There is no evidence of any pre-existing cave. No abandoned passages, channels, tubes or the like are intersected by either of the active caves. They appear to be "one cycle" caves.

The Madison does contain breccias that may represent a previous karst cycle of erosion. They are, however, cemented into solid rock today. Where they are cut by the caves; there is no evidence of influence on the cavern passages.

The spring slots are enlarging and appear to be contemporaneous with the caves. If, however, they did exist before the cave and if they channeled flow through the limestone then they would have influenced the development of the caves.

Hellespont Cave probably represents a cave early in its development. The cave is enlarging headward at the spring, due to vigorous replacement-solution where H_2S first enters. Simultaneously, the passage diameter is also increasing. The cave probably started as a seep by the river and grew headward to its present form in the same way it's enlarging today. Headward cavern development occurs in other areas (Egemeier, 1969, p. 109).

If the caves enlarge headward by solution in a manner similar to the way a gully enlarges by suspension, then development should follow a profile of equilibrium. The gully profile is controlled by suspended load, while in this case the profile is controlled by solution and therefore is much flatter. As the profile flattens oxygenation of the stream slows reducing its corrosive power. In steeper areas corrosion is strong and the profile flattens appropriately. In short the cave develops much as steephead valleys are developed by springs in limestone areas, such as Florida. In both cases headward development follows the path of maximum ground water flow toward the cave or spring.

These caves are in the desert in breached anticlines that allow flow of confined water out of the Madison limestone. The only recharge areas are tens of kilometers away. Thus, there is almost no cold water moving down from the surface and entering the caves. The surface water seepage into lower Kane Cave is too small to measure, but is less than 100 ml/min. The thermal spring flow is 450 l/min, around 5000 times greater. Even so, this small seep has washed away gypsum and caused the development of a small collapse dome. Thus, in other areas where seepage is greater it could have a substantial influence on cavern development and the caves formed might be quite different.

Features and Characteristics of the Caves

Replacement-solution caves have many distinctive characteristics and features that serve to distinguish them from simple-solution caves. The active caves show these features better than the dead caves, because ceiling collapse and the growth of speleothems have modified, removed, or hidden some of the features of the dead caves. (While reading the detailed descriptions of cave features, the reader should refer to the cave maps [figs. 6 through 8], as the descriptions are keyed to these maps. For larger-scale maps, see Egemeier, 1973.)

Gypsum crusts. Despite the fact that the active caves are limestone caves, almost no limestone is exposed to air in them. Fresh limestone brought into the caves forms a paper-thin gypsum crust within one year (reaction 3). Dolomite forms a gypsum-epsomite crust. All the limestone is covered with a gypsum crust, except: 1) within a few meters of the cave entrance; 2) where water is flowing over the limestone; and 3) where the limestone is coated with or buried by mud. The crust consists of white, finely-grained gypsum, in places coated with 2 mm sulfur crystals formed according to reaction 1. Where wet, the crust has a consistency similar to that of mud. Visible gypsum crystals are rare, but where present have a maximum length of 10 mm.

The gypsum crust on the limestone is irregular. Its surface is very rough (Fig. 9), and its thickness differs markedly when measured at points only a few centimeters apart horizontally, with the thickest crust being in pockets in the limestone.

The average thickness of the crust differs from one area of a cave to another (see Fig. 7). The crust is thickest on the cave walls nearest the thermal springs. Near the Upper and Lower springs, the crust is 2 to 50 cm thick. Near the entrance of Lower Kane Cave, there is little gypsum on the walls and ceiling of the cave; the limestone here is covered by a thin coating of mud. At the entrance, uncoated dry limestone is exposed. The dry bedrock shows no tendency to react with the cave air, as the reactions proposed only proceed rapidly in water. Damp limestone broken clean in 1971 near the lower spring in Lower Kane Cave had a spotty crust about ½ mm thick in 1975. The south wall of Lower Kane Cave, 45 m from the entrance, is wet with flowing seepage water. Very little gypsum crust is evident on this wall. Also, no gypsum crusts are present on the limestone in cave streams. In both these cases, the flowing waters are undersaturated with respect to gypsum; hence, none crystallizes on the limestone. In Hellespont Cave (see Fig. 6), the coating has a maximum thickness of several centimeters near the spring; the coating becomes thinner away from the spring, reaching zero on dry limestone at the entrance. Sulfur crystals are very common on the gypsum crust near the springs in both caves, prob-

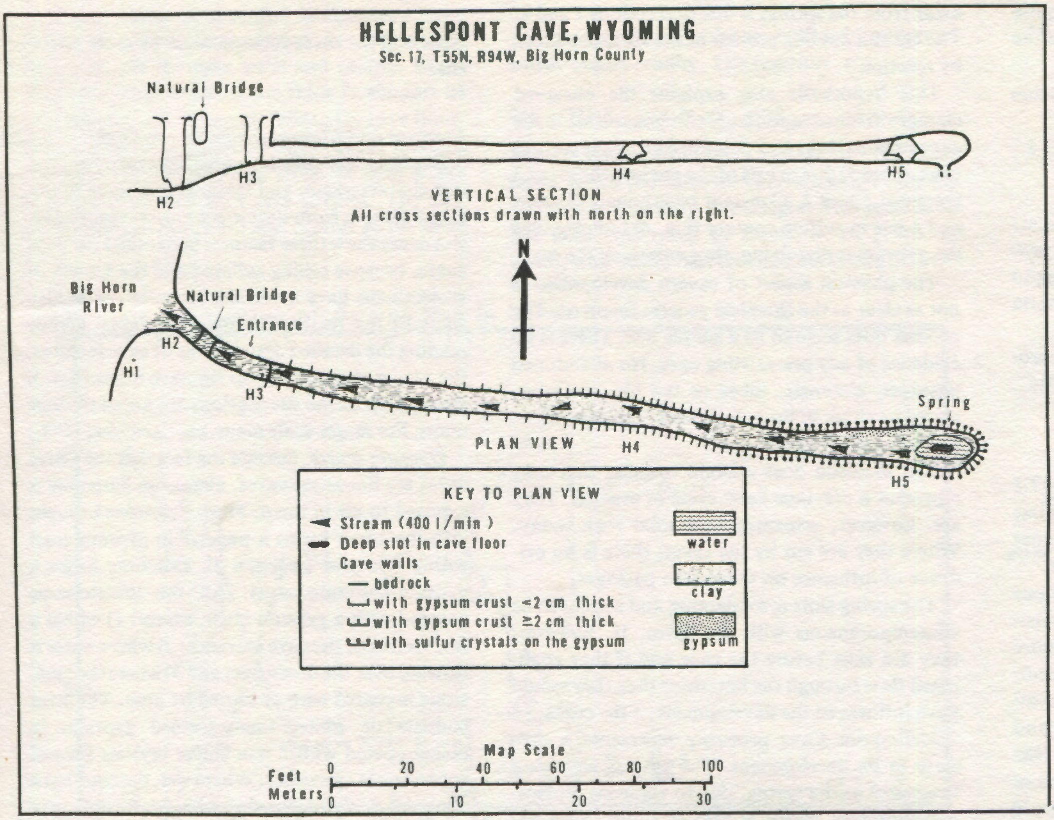


Figure 6. Map of Hellespont Cave.

Figure 8. Map of the Spence Caves.

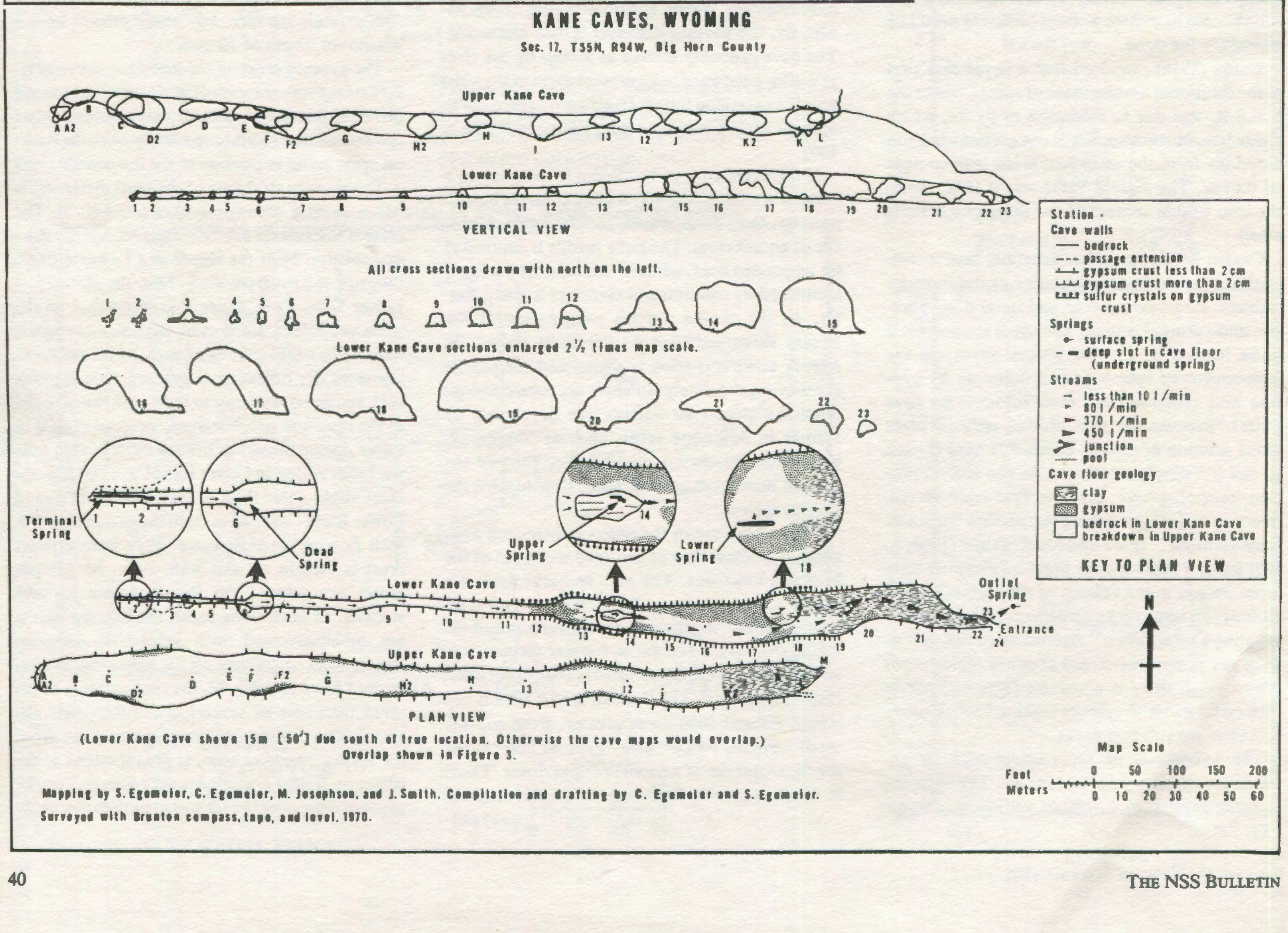


Figure 7. Map of the Kane Caves.

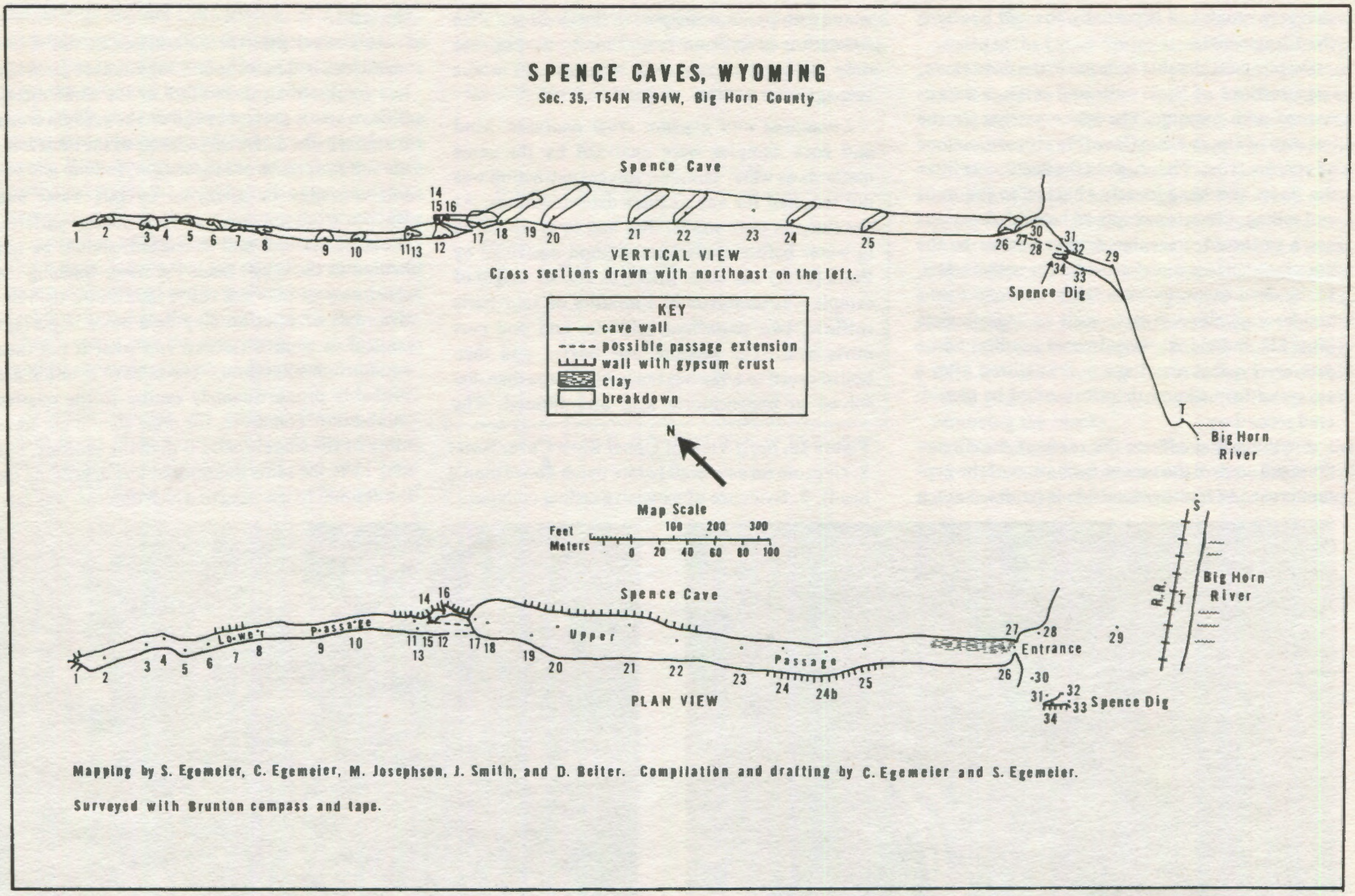


Figure 9. Stereo view of Lower Kane Cave taken from just upstream of the upper spring (looking downstream). Pool in foreground is the upper spring. **Note: 1.** Gypsum mounds on the stream banks. **2.** The joint exposed in the ceiling. **3.** The etched horizontal lines on the gypsum blocks just to the right of the stream. These were etched in a few weeks by the water when the stream was dammed. **4.** The roughness of the cave walls. **5.** The flat floor and bell-shaped passage cross section partially obscured by gypsum mounds.

ably as the result of partial oxidation of H_2S.

In the active caves, the crust, even where it is relatively thick, preserves some features of the underlying limestone. Faint traces of relict bedding are evident in the crust exposed on the walls. Chert bands in the limestone continue, although somewhat broken, into the gypsum crust. Where the crust is developed on brecciated limestone, faint suggestions of the breccia pattern are visible in the gypsum. This is further evidence that the gypsum is a replacement of the limestone.

The limestone surface beneath the gypsum crust is much like the surface of a corroded iron meteorite, having similar roughness, pockets, projections, edges and concavities. The scale of the roughness differs in different parts of the cave; depth of the pockets may be as little as 1 mm or as great as 1 m. The contact between limestone and gypsum is fairly sharp; the transition zone is typically 1 mm or less in thickness. The limestone bedrock in stream beds in Lower Kane Cave is also very rough and is partially covered by sharp chert fragments.

Despite considerable collapse in the dead caves, large sections of their walls and ceilings are encrusted with gypsum. The white patches on the left wall of Upper Kane Cave (Fig. 10) are sections of gypsum crust. The crusts in the dead caves litter the floors and hang loosely attached to the walls and ceiling. Limestone exposed beneath the crusts has a pocketed, irregular surface similar to the limestone surface under crusts in the active caves. In the dead caves, gypsum takes on many forms besides crystalline crusts, such as oulopholites (Fig. 11), individual crystals, and needles. Some primary gypsum may have re-crystallized after a cave was formed and therefore could be considered secondary.

Anything that affects the rates of these reactions will control the rate of formation of the gypsum crust. As hydrogen sulfide is oxidized much more rapidly in aqueous solution than in air, no crusts form on dry limestone in active caves. Coatings, such as mud, on the cave walls can and do prevent the formation of a crust by preventing sulfuric acid from coming in contact with limestone. The amount of hydrogen sulfide available to react with the limestone also controls the thickness of the crust. Near the springs, where the concentration of hydrogen sulfide in the cave air is the greatest, the gypsum crust is thickest, as would be expected (figs. 6 and 7). No gypsum crust forms in cave streams, as this water is undersaturated with respect to gypsum. No crust is formed in areas of seepage, probably for the same reason. The distribution and thickness variations of the crust in the active caves are logically explained by replacement-solution.

Figure 10. *(left)* View of Upper Kane Cave. **Note: 1.** Gypsum on cave wall (white patch above man's head). **2.** Evidence of extensive ceiling collapse.

Figure 11. *(right)* Oulopholite in Spence Cave. The carbide lamp has a 10 cm reflector.

Gypsum crusts similar to those in the active caves have been found in many caves, including Carlsbad Caverns, N.M. and Mammoth Cave, Ky. Relict algal structures are preserved in gypsum in the Big Room in Carlsbad. Crusts in Mammoth Cave have been studied by Pohl and Born (1935) and Pohl and White (1965). These authors also suggest that the Mammoth Cave gypsum crusts were formed by the reaction of oxygenated water with pyrite, transport to the cave, and then formation of gypsum from limestone. This has only occurred in relatively dry passages where seepage rates are low.

Limestone and gypsum crust analyses. Mud and rock samples were analyzed by the same methods as water samples; solvent extraction was not required for any of these determinations. As the samples were solids, they had to be dissolved in water before analysis; a method described by Pinta (1971) was used. Briefly, a known weight of sample was dissolved in a mixture of four parts sulfuric, two parts hydrochloric, and one part nitric acid. The mixture was heated and then boiled down to a viscous mass which was then dissolved in hydrochloric acid and filtered. The residue was washed with hot hydrochloric acid and the washings were added to the sample. The residue was weighed and the samples were ready for analysis.

If the gypsum crust is a replacement of the carbonate bedrock, the crust perhaps would show chemical similarities to the carbonate bedrock to which it is attached; crusts should show variations in composition where developed on bedrock of different compositions. With this in mind, fourteen paired specimens were taken from four caves. Each specimen was a piece of carbonate bedrock with gypsum crust attached. Before analysis the crust was removed to within 1 mm of the carbonate rock. This last millimeter of crust was then removed and dissolved for analysis as described previously. A millimeter of carbonate rock from just below the crust was similarly removed and prepared for analysis. The purpose of removing the analytical samples from as closely adjacent parts of the specimen as possible was to minimize the sampling error due to small-scale variations in the composition of the bedrock and the crust.

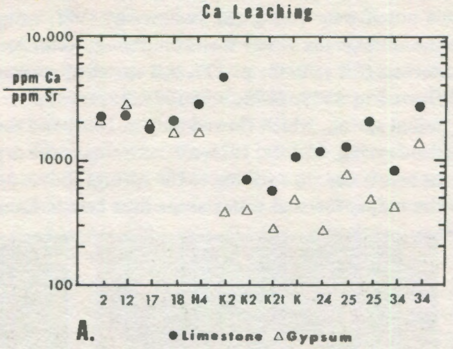
At neutral pH, the carbonates of the major metal ions in limestone and dolomite are generally less soluble than the sulfate of the same metal. Since water is present in and on the gypsum crusts forming in the active caves, some of the most soluble sulfates, those of magnesium, sodium and potassium, may be dissolved by this water and leached from the crust. The less soluble sulfates, those of calcium and strontium, should be left behind in the crust. Since the water leaching the crust may or may not reach saturation, the relative rates of solution may be a more important control of what is leached and what is not then equilibrium solubilities. The rates of solution are probably proportionately similar to the relative solubilities. Therefore, the crust should not have more of the soluble cations than the bedrock, but may have the same quantity or less depending on the amount of leaching to which the crust was subjected. Likewise the crust should be enriched in the least soluble cations relative to the bedrock. This systematic relationship should apply to all samples.

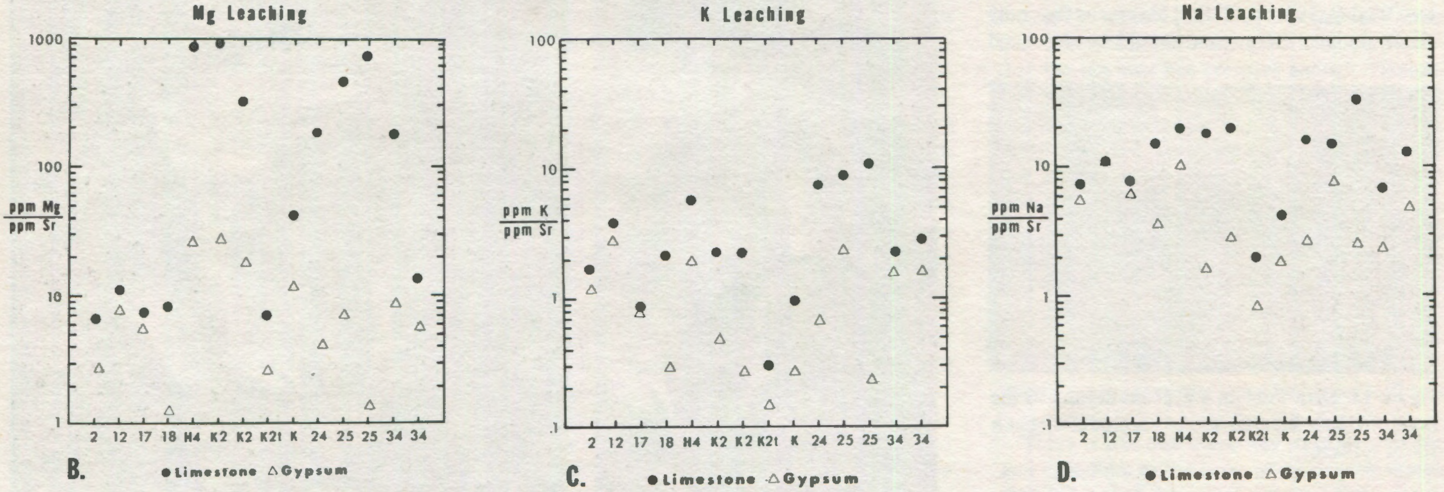
Figure 12 shows the relative amounts of several metals in the gypsum crust and limestone bedrock samples. Strontium is used as a basis of comparison, as it forms the least soluble sulfate and thus is probably leached the least. The graphs show that, within analytical and sampling error, the cation-to-strontium ratio is less in the crust than in the associated bedrock, even though there is more than an order of magnitude difference in the cation-to-strontium ratio among the samples. Such a close relationship of crust to bedrock composition indicates that the crust and bedrock are related, as would be the case if the crust were a replacement of the bedrock. A crust deposited by precipitation of gypsum from ponded or seepage water in the cave would be fairly uniform in composition and need not relate systematically to the composition of the bedrock on which it was precipitated.

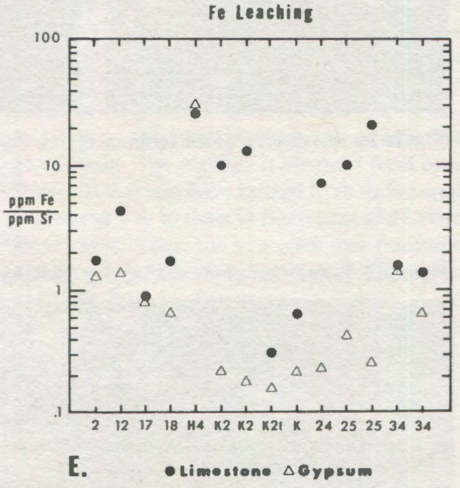
Figure 12e shows the iron content of the same bedrock and gypsum crust samples. Most of the iron in the bedrock seems to be limonite. The iron-bearing minerals in the crust are not known, but apparently they are more soluble, as the iron content relative to strontium is consistently less in the crust than in the bedrock. This is further evidence that the crust and bedrock are related.

It seems clear that the gypsum crusts in all the caves are replacements of the bedrock beneath the crust. With few exceptions, all probably within analytical error, the composition of gypsum crust samples was systematically related to the composition of the limestone beneath. Certainly this is as indicative that the crust is a replacement as are evidences of bedrock chert bands and relict bedding found in the crusts in the active caves. Thus it seems that the gypsum crusts in the caves studied are replacements of limestone.

Figure 12. Chemical comparison of limestone and its associated gypsum crust. Note that metal sulfates more soluble than strontium sulfate are selectively leached from the gypsum crust.







Gypsum mounds. An unusual feature of the caves is the large gypsum mounds that cover a substantial part of the cave floors. In Lower Kane Cave and Hellespont Cave, the mounds consist of unbedded, finely-crystalline, soft gypsum similar to that which makes up the crust above the mounds. In the dead caves, the mounds are dry and crystalline, occasionally cut by 2 m-deep drip tubes dissolved by dripping water (Fig. 13). Where the mounds are under crusts containing chert, they also contain chert; otherwise, they do not. With one exception, 2 m-deep holes drilled into the mounds revealed no limestone. The exception is a mound built against the south wall of Lower Kane Cave, 45 m from the entrance, in the place where seepage is entering the cave.

In Lower Kane Cave, active enlargement of floor gypsum mounds has been observed as a result of the spalling of parts of the gypsum crust from the cave ceiling. During the many days spent in Lower Kane Cave, several falls of gypsum ceiling crust were seen and/or heard; most occurred in the area between the Upper and Lower springs (stations 13 to 20). The largest mound (5 m thick, 10 m wide and 30 m long) and thickest crusts have formed in this section of the cave, which is between the two thermal springs. The upstream end of the largest mound is shown in Figure 9. A July, 1975 visit to the cave revealed that a gypsum fall of several cubic meters has produced a new mound near station 18.

The cave streams are dissolving gypsum in Lower Kane Cave. A 20 kg block of gypsum placed in the Upper Spring stream near station 15, between the Upper and Lower springs, dissolved completely in a week. Gypsum blocks used to build dams for weir stilling ponds dissolved and caused failure of the dams in a few days unless the gypsum blocks were isolated from the water by a coating of mud. The horizontal lines etched in the gypsum blocks on the floor in Figure 9 resulted from damming of the stream for less than a week. The rapid breakdown of the gypsum blocks could be caused by solution of epsomite in the blocks, exposing fine-grained gypsum to the water. The blocks break down equally rapidly in fairly stagnant pools, indicating that solution is probably more significant than erosion as the process removing the blocks.

The mounds, too, are being dissolved by the cave streams. The mounds in Lower Kane Cave have oversteepened banks along the cave streams, probably due to solutional undercutting of the gypsum mounds by the cave streams. The continued presence of mounds indicates continuing addition of gypsum, or the mounds would have been removed by solution. Hence, continued falls of crust must be maintaining the mounds. Falling crust has been observed in Lower Kane Cave on several occasions. Since this is so, then the crust must be replaced or limestone would be exposed in the cave, and almost none is exposed. Replacement-solution is an on-going process in this cave.

All the caves studied contain gypsum mounds. Hellespont Cave has small mounds almost identical to the ones in Lower Kane Cave. The dead caves also have some very large mounds, most of which are partially covered by breakdown. The mounds in Upper Kane and Spence caves consist of dry, finely-crystalline gypsum, except for mounds in the driest parts of Spence Cave, which are loose piles of gypsum sand and dust. In places, dripping water has dissolved long tubes through the mounds, such as the tube shown in Figure 13. Some of the tubes in Spence and Upper Kane caves penetrate almost 2 m of gypsum. None of the tubes intersect limestone blocks within the gypsum.

Gypsum deposits in the Big Room in Carlsbad Caverns in New Mexico (Bretz, 1949, p. 454; Good, 1957, p. 12) are very similar to the deposits in the Wyoming caves. The Carlsbad deposits, with one possible exception, do not contain limestone blocks (Good, 1957, p. 12); they are not bedded and they contain drip tubes. The walls of the Big Room have remnants of gypsum crusts, especially in solution pockets along the sides of the room. Good (1957, p. 19) proposed that the floor deposits were precipitated from gypsum-saturated surface water that cooled upon entering the cave. It does not appear that the gypsum in the Wyoming caves could have been formed in this way, since the cave streams are dissolving and removing the deposits, not precipitating them.

There are problems with this theory of origin for the gypsum deposits in Carlsbad as well. Gypsum solubility is not very sensitive to temperature (see Wigley, 1971, p. 472). This means that only small amounts of gypsum would be deposited from large quantities of water. Over the period of deposition, it is likely that sand and silt might be washed into the cave and deposited with the gypsum. Therefore, it seems that the Carlsbad deposit should contain sand and silt partings, but it does not (Good, 1957, p. 12).

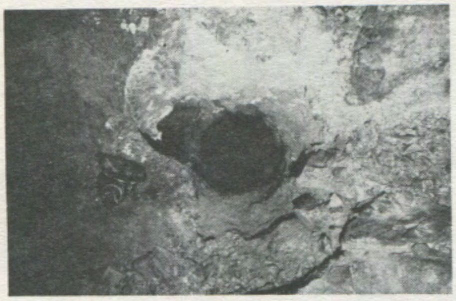


Figure 13. Drip Tube in a gypsum mound in the floor of Upper Kane cave. The carbide lamp has a 10 cm reflector.

Today, the surfaces of the gypsum deposits are beneath fallen limestone. One would expect that limestone was falling during deposition also, but only one block of incorporated limestone has been found in the mounds (Good, 1957, p. 12). The gypsum mounds in the Wyoming caves are also free of incorporated limestone blocks. In a few places, the gypsum in Carlsbad has patterns that appear to be algal stromatolites inherited from the limestone the gypsum replaced. It is possible that the gypsum deposits in Carlsbad Caverns were formed by replacement-solution.

Thermal springs. There are several thermal springs in and near the Wyoming caves. At river level below Spence Cave are several seeps whose waters are 34° C, and two larger springs with temperatures of 20° and 21° C (Fig. 1b). In the Kane Caves area, there are also several springs (Fig. 3); the coolest and largest is the one that feeds the Salamander Cave stream. Unfortunately, this spring is inaccessible due to collapse of the cave under railroad tracks. The discharges indicated on Fig. 3 were determined with a 90° V-notch weir during the summer of 1971, using the tables in the Water Well Handbook, Keith Anderson (Ed.), 1948, p. 127. All the cave springs flowed in 1969, 1970, and 1971, except the terminal spring, which flowed only in 1969, and the dead spring. In July, 1975, all the springs were dry or nearly dry. In addition to the springs shown on the maps, thermal water seeps into Lower Kane Cave at several points.

The thermal springs in the caves rise through slots in the bedrock floors of the caves. The slots are rectangular in plan and many times longer than they are wide. This is because they are situated along joints. The slots studied were 15 to 30 cm wide and 1 to 10 m long; typical slots are shown in figures 14 and 15. The author has not observed flow from the floor slot shown in Figure 15 (Dead Spring). However, the similarity of this slot to the other spring slots suggests that it is a dead spring.

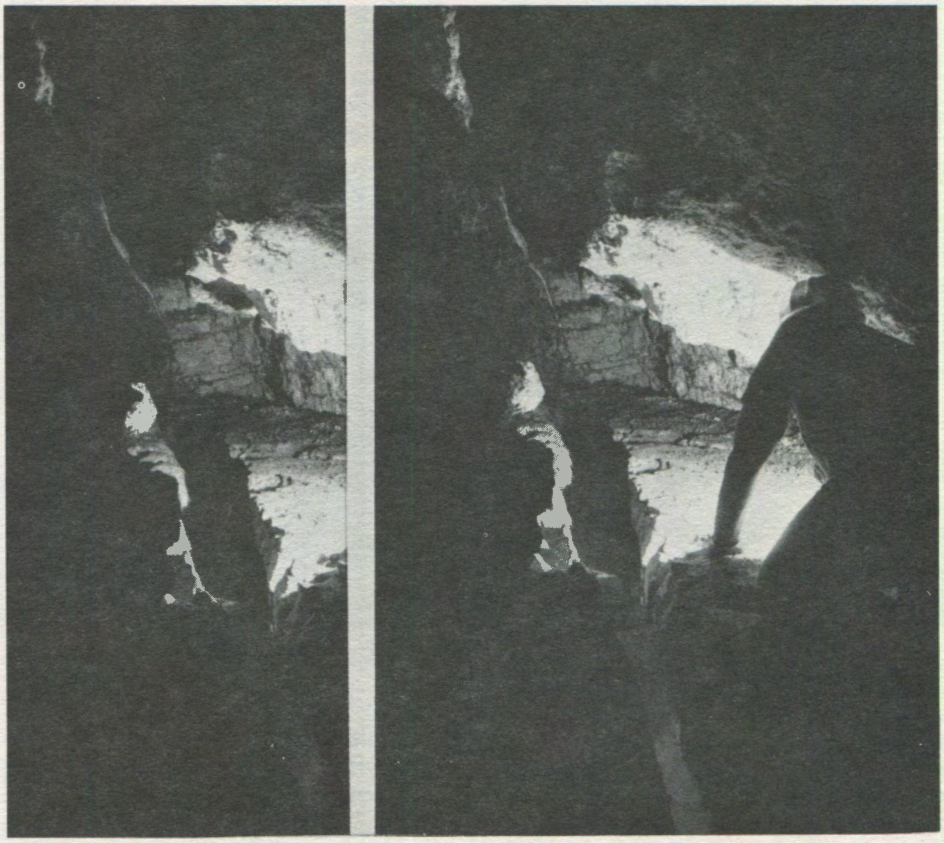


Figure 14. Stereo view of the terminal spring showing feeder slot in the floor (looking upstream).

Figure 15. Stereo view of the dead spring (looking upstream). Note the irregular contour of the cave wall.

Efforts to plumb the depth of the springs met with only partial success. Lower Spring in Lower

Kane Cave is almost 2 m deep; at this depth the spring slot is offset to the south or is fed by a channel from the south. The depth of Upper Spring is undetermined, but exceeds 3 m, which is the length of a pole that was maneuvered into it. The same pole also failed to reach the bottom of Terminal Spring. There is no way to know how much the depth of the springs exceeds the length of the pole. The water has come tens of kilometers from the rechange areas and flowed at depths of a kilometer or two. How much of this route is seepage and how much is pipe flow is unknown.

The edges formed by the walls of the slots and the cave floor are embayed, as shown in Figure 14. Apparently, water at the surface of the spring vigorously dissolved away the edges of the slot, producing anastomosing channels and upside-down pendants in the limestone.

All the flowing cave springs contain deposits of sharp chert fragments and black, hydrogen sulfide-rich, radioactive muds. Mud removed from Upper Spring and left out of water in the cave for a year oxidized to a red clay. These muds consist mainly of clay and sulfide precipitates from the spring waters.

For a few days in 1970, Upper Spring contained numerous 2 to 4 mm diameter balls of oil in the spring orifice. There is apparently enough organic material in the spring water to feed bacteria, which in turn are food for hundreds of colorless snails that live just downstream of each spring.

The springs rise under moderate pressure, which may change slightly over time. The water level in Terminal Spring has been as low as 1 m below the lip of the spring at times, and has flowed over the lip at least once since 1969. A dam especially built to raise the level of Upper Spring pool ½ m did not measurably slow its flow. The measured flows of the springs were steady over a period of a month. Thus, it seems that the pressure is very nearly constant, and, as a result, the flows are also nearly constant.

No spring slots are present in the dry caves, nor have similar slots been reported in other caves. A hole in the floor of Upper Kane Cave 20 m from the entrance, near station K, may be a part of a spring slot, but it is too small to be explored. It appears to be round rather than slot-like, but the silt deposits around it may be masking its true shape. In this cave and in Spence Cave, the bedrock floor is covered by breakdown; thus, if slots are present they are very difficult to find because they are buried.

Precipitate and mud samples. The thermal springs contain aqueous deposits of muds and precipitates. While in the springs the deposits are black, but if removed from the water or if the water level falls they turn hematite-red after several months, perhaps due to oxidation in the muds. The muds smell very strongly of hydrogen sulfide. If the muds are disturbed under water, bubbles of flammable gas with a strong hydrogen sulfide odor rise to the surface, possibly indicating that some hydrogen sulfide and perhaps methane is produced in the muds. If hydrochloric acid is added to the muds, hydrogen sulfide gas is given off, probably due to reaction of the acid with metal sulfides in the mud. Enough organic material is present in some samples of the mud that the sample will smolder in an open flame. The organic material seems to be oil brought into the caves by the springs. All the samples checked were radioactive, most likely indicating the presence of uranium or thorium.

The discharge of water into the caves is very rapid. The slot of Upper Spring has a cross-sectional area at the spring of 0.1 to 0.5m^2. Since the discharge is 370 l/min, the flow velocity must be about 1 m/min at the spring and more at depth. Such a strong flow could carry solids precipitated at depth to the spring pools, where they might settle out. The waters may travel 1500 m vertically in one day, which could result in a maximum change in temperature of 10° to 20° C.

The source of the hydrogen sulfide in the muds is probably microbial activity. Dr. James A. Lautenberger of the Department of Biochemistry, University of California, Berkeley,* kindly performed several tests on samples of waters and muds from Upper Spring in Lower Kane Cave. Using the procedure of Grossman and Postgate (1953), it was determined that the muds contained sulfate-reducing micro-organisms. Tests on the water were negative, indicating that sulfate reducers were absent or in much smaller quantity in the water. Microscopic analysis revealed the presence of many slightly-curved, long, rod-shaped organisms in a sample cultured from the spring mud.

The muds in the cave springs have a moderate heavy-metal content. Analyses of muds are listed in tables 5 and 6; the analyses in Table 5 are more precise, but less complete than the ones in Table 6. Only one metal, vanadium, is in a concentration of any economic interest; however, the deposits in the active caves total only a few cubic meters. The radioactivity is probably due to uranium in the cave muds. Economic tyuyamunite deposits are found in caves in the Big Horn and Pryor mountains, along the north and east edges of the Big Horn Basin. The uranium is obtained from cave floor silts that may have formed from spring mud deposits similar to those in Hellespont and Lower Kane caves. Thus, the processes that transport and deposit vanadium and uranium in these caves are of interest.

The gypsum crusts from the caves were also checked for heavy metal and sulfide content. Dr. Martin S. Rutstein of SUNY, New Paltz, N.Y. analyzed the crusts by X-ray fluorescence. No sulfide minerals were detected. Optically, less than 1% of the crusts were opaque minerals. Atomic absorption analyses of eight gypsum crust samples averaged 0.06% iron, the highest value being 0.3%. Silver, copper, and cobalt were not detected at the 50 ppb level. The heavy-metal content of the crust is 2 to 4 orders of magnitude below that in the muds.

Uranium is presently being transported in slightly oxidizing waters very similar to the cave waters (Butler, 1969, p. 83). However, since uranium and vanadium are essentially insoluble in a reduced state, hydrogen sulfide water such as found in the cave springs cannot transport either metal. If all the hydrogen sulfide in the water is produced by sulfate-reducing bacteria in the springs, the uranium and vanadium could be transported to the cave, where they would be reduced and precipitated in the spring.

However, there is some evidence that suggests that not all the hydrogen sulfide present in the spring waters is produced there. First, sulfate and organic matter are present in the cave waters; thus, it is possible for sulfate-reducing bacteria to produce hydrogen sulfide in the water as well as in the spring muds. The fact that they were detected only in the muds and not in the water may mean that none are in the water, or it may only mean that the test was not sensitive enough. Second, samples taken from the spring orifices had as much hydrogen sulfide as did samples from the surface of the same springs. Certainly these samples should have less hydrogen sulfide if most of the hydrogen sulfide were produced in the springs. Of course, it is possible that the hydrogen sulfide is produced at a shallow depth below the caves, and that it and the precipitates are carried up to the springs, where they settle out. Third, oil wells in the Madison limestone often produce hydrogen-sulfide rich water (Crawford, 1940, p. 1304); this suggests that some hydrogen sulfide is in the water before it reaches the cave.

There is still one more way to carry uranium and vanadium to the cave: in the organic matter that is carried into the springs. The organic material contains six orders of magnitude more vanadium than the spring water (see Table 6). The vanadium content of Big Horn Basin oil is around 30 to 50 ppm (Stone, 1967, p. 2082). Organic matter in the active cave springs is radioactive, indicating that it contains uranium. Vanadium and uranium in oil or organic matter could be concentrated by bacterial destruction of the oil that dilutes it. Thus, the hydrothermal fluid that formed this deposit may be oil, and not water as would be expected.

The source of the vanadium in the oil is uncertain; it could be from the Phosphoria formation. According to Rubey (1943), some beds in the Phosphoria formation contain up to 1% vanadium.

Possible sources of the uranium are not clear. Elliot (1963, p. 5) suggests that Pleistocene ash beds (since eroded) were the source for uranium in the Pryor and Big Horn Mountain caves. Unless the water circulation to the active caves in the basin is incredibly slow, these eroded beds cannot be the source of uranium being deposited now. Perhaps the uranium was temporarily stored in the bedrock or in the oil, or perhaps it came from

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Table 5. Precipitate and Mud Samples.

No. Location	Date	Ag ppm	Ca %	Co ppm	Cu ppm	Fe %	K %	Li ppm	Mg %	Na ppm	Sr ppm	Radioactivity MR/hr†	Undissolved Residue %
Lower Cane Cave													
1. Lower Spring	7/16/70	3	0.8	14	32	1.0	0.9	37	0.1	490	35	0.025	74
2. Upper Spring	8/8/70	3	3.7	7	230	0.9	0.6	22	0.4	410	96	0.015	67
3. Lower Spring	9/18/71	9	2.0	14	27	1.4	0.7	170	0.4	510	64	0.025	57
4. 3 m downstream of #3	9/18/71	3	1.2	13	26	1.7	0.7	58	1.2	500	58	0.025	68
5. 15 m downstream of #3	9/18/71	3	0.9	10	21	2.0	0.9	110	0.5	420	47	0.025	83
6. Upper Spring 1 m*	7/18/71	6	1.1	15	35	1.5	0.6	110	1.5	270	50	0.020	73
7. Upper Spring 0.3 m	7/18/71	5	1.4	15	21	1.7	0.4	170	0.1	430	74	0.035	51
Hellespont Cave													
8. Spring	9/19/71	3	0.2	9	21	0.9	1	64	0.3	330	28	0.020	73

*Sampling depth (below water surface). †Background count 0.005 MR/hr.

Table 6. Results of semi-quantitative flame emission spectrochemical analyses.*

Element	Limestone sample #18 Station 18	Mud sample #7 Upper Spring %	Water Upper Spring ppm
Ag	0.0000	0.0001–0.0002	0.0008–0.001
Al	0.0–0.2	> 8	0.004–0.008
Ba	0.0000	0.3–0.5	1–2
Be	0.0000	0.00005–0.00008	0.000
B	0.00	0.02–0.03	0.1–0.2
Ca	major	> 6	> 40
Co	0.000	0.0005	0.001–0.002
Cr	0.0002	0.002–0.004	0.000
Cu	0.0001–0.0002	0.0005–0.0007	0.0004–0.0008
Fe	0.03–0.05	3–5	0.08–0.2
K	0.	2–4	20
Li	0.0	0.0	0.08–0.2
Mg	0.02–0.04	> 5	>40
Mn	0.0001–0.0002	0.002	0.000
Mo	0.000	0.01–0.02	0.003–0.004
Na	0.0	0.2–0.3	10-20
Ni	0.0001–0.0002	0.001	0.000
Pb	0.000	0.008–0.01	0.00
Si	0.3–0.5	>10	2–3
Sn	0.000	0.000	0.008–0.01
Sr	0.02–0.03	0.01–0.03	0.8–2.
Ti	0.005–0.007	0.1–0.2	0.00
U	0.0	0.0	0.
V	0.000	0.1–0.2	0.004–0.008
Zn	0.00	0.006–0.008	0.0
Zr	0.000	0.004–0.006	0.00

*Analyses by Charles Taylor, Stanford University.

some other source.

Besides uranium and vanadium, other metals such as chromium, copper, iron, manganese, molybdenum, nickel, silver, lead, and zinc are deposited in the cave springs. The deposition of many of these metals may be due to gradual cooling of the water as it rises into the cave, or to precipitation by hydrogen sulfide in the spring muds. The solubility of many of these metals is low in waters containing hydrogen sulfide, except when the concentration of hydrogen sulfide is very high (Barnes and Czmanski, 1967, p. 371) or when the chloride concentration is very high (Lebedev, *et al.*, 1971, p. 511). Most of the metals found in the springs are at concentrations of 1 to 100 ppb in the spring waters, which may mean that the water cannot dissolve significant quantities of these metals, or that significant quantities of these metals are not available to be dissolved, or that most of the metals are precipitated before the water reaches the caves. The Madison limestone is a poor source rock for these metals (Table 6). The fact that many caves do contain economic deposits of many of these metals suggests that, at least in some cases, when source rocks are present, these metals can be transported to and deposited in significant quantity in caves.

Sulfides of some of these metals have been mined from caves. Sphalerite encrusts caves in the Mascot-Jefferson City area in Tennessee (Harris, 1971, pp. 734-735). Caves in the Metaline Zinc-Lead district in Washington (Dings, *et al.*, 1965) and in the Illinois, Wisconsin, Iowa Tri-state district (Bretz, 1938) contain sphalerite and galena. Cave fills in some caves in southeast Missouri contain galena crystals (Buckley, 1908, p. 93). All these minerals, plus some silver and manganese, are found in caves in the Castle Mountain mining district in Montana (Winters, 1968, pp. 51, 55). Silver-containing zinc and copper sulfides are mined from caves in the Tintic district in Utah. These deposits may have been formed by hotter waters with more dissolved chloride or hydrogen sulfide than the waters in these cave studies.

Caves near ore bodies may be partly or wholly secondary. That is, they may have been formed or enlarged by oxidation of the ore (Morehouse, 1968, p. 7), and thus may be post-ore or partly post-ore features. Most of the caves with deposits listed above are thought to be pre-ore or contemporaneous with the ore, but some of the caves may prove to be post-ore features.

In summary, the cave springs are traps for some metals. Two metals, uranium and vanadium, are of particular interest because they are also found in economic quantity in similar caves in the Big Horn and Pryor mountains. The most likely cause of precipitation of these metals is hydrogen sulfide produced in the springs. Also, small quantities of heavy metals are precipitated in the springs probably as sulfides. No nearby caves contain these metals in economic quantity.

An estimate of the age of the caves.

In Tertiary time, the Big Horn basin was filled with alluvium. The Big Horn River flowed across the fill along a course that has been maintained as it cut downward. Thus, the river cuts across several mountains instead of flowing around them. As the mountains were exhumed from the fill by erosion, several were breached by the river. When the Madison limestone was breached, confined water and trapped petroleum were able to escape and the caves were formed. In unbreached domes both the water and oil remains trapped.

Since erosion proceeded unevenly, terraces were formed during periods of stability. Those in the basin are cut in the alluvial fill and those in the canyons are cut into bedrock.

Cavern development also requires stability and, hence, cave elevations and terrace elevations correlate very well. Spence cave is entered from a sizable terrace remnant. Upper Kane is entered from a tiny remnant of a terrace that is better preserved on the opposite side of the canyon. Lower Kane and Hellespont are at the present river elevation.

Brown clay and silt sediments are present near the cave entrances in all the caves. In the active caves, Hellespont and Lower Kane caves, the highest clay sediments are less than 4 m above the July 2, 1970, river level. The largest flood recorded at a USGS gaging station 1.5 km downstream from the cave, since it was installed in 1953, was 3.0 m higher than the July 2nd level. Thus, the sediments in the cave entrances are within flood range and probably are Bighorn River flood deposits. This would explain the mud coating on the walls and ceiling in the entrance area in Lower Kane Cave, as well as the clay floor deposits. Flooding may also have removed the gypsum crusts near the cave entrance. Clay deposits in Spence and Upper Kane caves are also found only near the cave entrances. Perhaps these clay deposits are also Bighorn River floodplain deposits, deposited at a time when the Bighorn River was much higher than it is now.

If this is so, these deposits should be at the elevation of a former floodplain. Vestiges of these floodplains—river terraces—exist at the elevations of the cave entrances. Mackin (1937, p. 864) recognized two terraces both at Kane and at Greybull, Wyoming (Fig. 1b). The lower of these, the Powell-Emblem bench, is 24 m above the Bighorn River at Kane and 33 m above the river at Greybull. Upper Kane Cave, which is located between these two points, contains silt deposits that are 27 m above the river, which is the probable elevation of the terrace near the cave. The other terrace recognized by Mackin is the Y.U.-Kane terrace, which is 130 m above the river at Kane and 135 m above river at Greybull. Spence Cave (see Fig. 8), which is located between these points, contains silt deposits 135 m above river level, or at the expected terrace level in this area. The cave and terrace elevations are correlated.

Table 7. Dimensions of Cave Passages (meters).

Passage	Length	Avg. Diameter	Volume	Max. Width	Max. Height
Hellespont	56	2	91	3	2
Lower Kane	320	6	10,000	17	9
Upper Kane	290	11	27,000	21	12
Salamander	14	2	20	4	2
Lower Spence	290	7	12,000	12	8
Upper Spence	420	14	62,000	33	30

Since the caves contain river silt deposited at the time the terraces were formed, they must be as old as, or older than, the terraces. A terrace remnant called the Tatman bench is preserved 370 m above the Greybull River. This terrace, which is higher than the terraces at the mouths of the caves, has been tentatively dated late Pliocene-early Pleistocene by Rohrer and Leopold (1963). Thus, the cave-mouth terraces and presumably all the caves are Pleistocene. The Cody terrace, which is below the Powell-Emblem bench, in the Shoshone Valley, was correlated by Moss and Bonini (1961, p. 551) with the Pinedale glaciation. This terrace merges into the Bighorn River floodplain (Moss, 1974, p. 309) indicating that the Bighorn River has not lowered its valley significantly since the terrace formed 10,000 years ago. All the active caves at present river level have, therefore, had at least 10,000 years available for development. If, as suggested by Moss (1951, p. 67), the higher terraces and benches also correspond to glacial stages, the Powell-Emblem bench may correlate with the Bull Lake glaciation. If so, its age is 40 to 130 thousand years BP (Birkeland, *et al.*, 1971, p. 219). This, then, may be the age of Upper Kane Cave. Spence Cave is below the Tatman bench and above the Powell-Emblem bench, and thus may be Pleistocene, but older than the Bull Lake glaciation.

Stream gradients

The very gentle floor gradients in the cave are probably controlled by oxygenation rather than the water table. A comparison of figures 4 and 16 reveals that oxygenation is most rapid where the gradient is steepest and slowest where the gradient is flattest. More rapid oxygenation due to greater turbulence in steeper stream segments apparently results in more rapid sulfuric acid production and more rapid solutional downcutting of the stream bed. As downcutting proceeds, turbulence, oxygenation, and passage slope will decrease until an equilibrium slope is attained. The result is a stream graded by solution.

Main passages

All the caves are characterized by a linear main passage or passages. The passages of Upper Kane, Lower Kane and Hellespont caves are developed along single joints. The dimensions of the main cave passages are given in Table 7. In the active caves, typical passage cross-sections are semicircular, but often the passage is wider or flared at floor level, so that many cross-sections are bell-shaped. The bedrock cave floors are flat, apparently even where hidden by mud and/or gypsum mounds. Hand drilling of these deposits where they are less than 2 m thick indicated an approximately flat bedrock floor in Lower Kane Cave. The flat floor is probably produced by meandering or braiding of the cave streams. The passage of Upper Kane Cave and the lower passage of Spence Cave are also semicircular in cross section; a flare or widening of the passage near floor level, if present, is buried by breakdown. The upper passage in Spence Cave has a cross section that resembles a rectangle dipping 30° NE (Fig. 8). One explanation for this cross-sectional shape might be homoclinal shifting of the cave-forming stream down the dip slope of the Madison limestone. The striking difference in shape of this passage, which is on the flank of an anticline, as compared with the others, which are in nearly flat-lying limestone near the axes of anticlines, suggests that the variety of passage shapes and patterns observed in the caves studied are restricted by the fact that all the caves are in anticlines. Caves formed by the same processes under different structural constraints might have different cross-sectional shapes and floor plans. This is known to be the case for caves in New York State (Egemeier, 1969).

Typically, cave passages narrow gradually upstream; the passages in the caves studied do not. In Lower Kane Cave above station 20, 50 m from the entrance, the main passage consists of successively smaller diameter segments, each starting above a spring (Fig. 7). Below station 20, the passage narrows toward the entrance. In Hellespont Cave (Fig. 6), the passage is largest near the spring and also narrows slightly toward the entrance. In both caves the passage is largest near the flowing springs. The processes enlarging the caves seem to be ineffective upstream of the springs, most efficient near the springs, with the efficiency dropping off slowly downstream of the springs. This observation fits well with what would be expected for the replacement-solution process.

Collapse has altered the passage shape in the dead caves; however, several similarities to the active caves are still evident. First, the passages do not taper up-gradient. In fact, both the upper passage in Spence Cave and the passage in Upper Kane Cave are largest near their far ends, like the passage in Hellespont Cave. Second, all the passages have blind terminations similar to the terminations at the springs in the active caves; there is no indication that the passages once continued and have collapsed or been filled. This is unusual, as in most caves, especially those in the eastern United States, passages narrow upstream until they become impassable; blind terminations are unusual except where due to collapse or fill.

It is interesting to note the strong similarity in the passage patterns of Spence Cave and the Kane caves. The lower passage in Spence Cave is blocked southeast of station 12 by ceiling collapse. Another collapsed passage wtih gypsum wall crusts was found in the Spence Dig. If Spence Dig is the outlet end of the lower passage, then Spence Cave consists of two passages, Lower Spence Cave and Upper Spence Cave, analogous to Lower Kane Cave and Upper Kane Cave. If Lower Kane Cave continues to enlarge upward it may cause a collapse, resulting in a connection be-

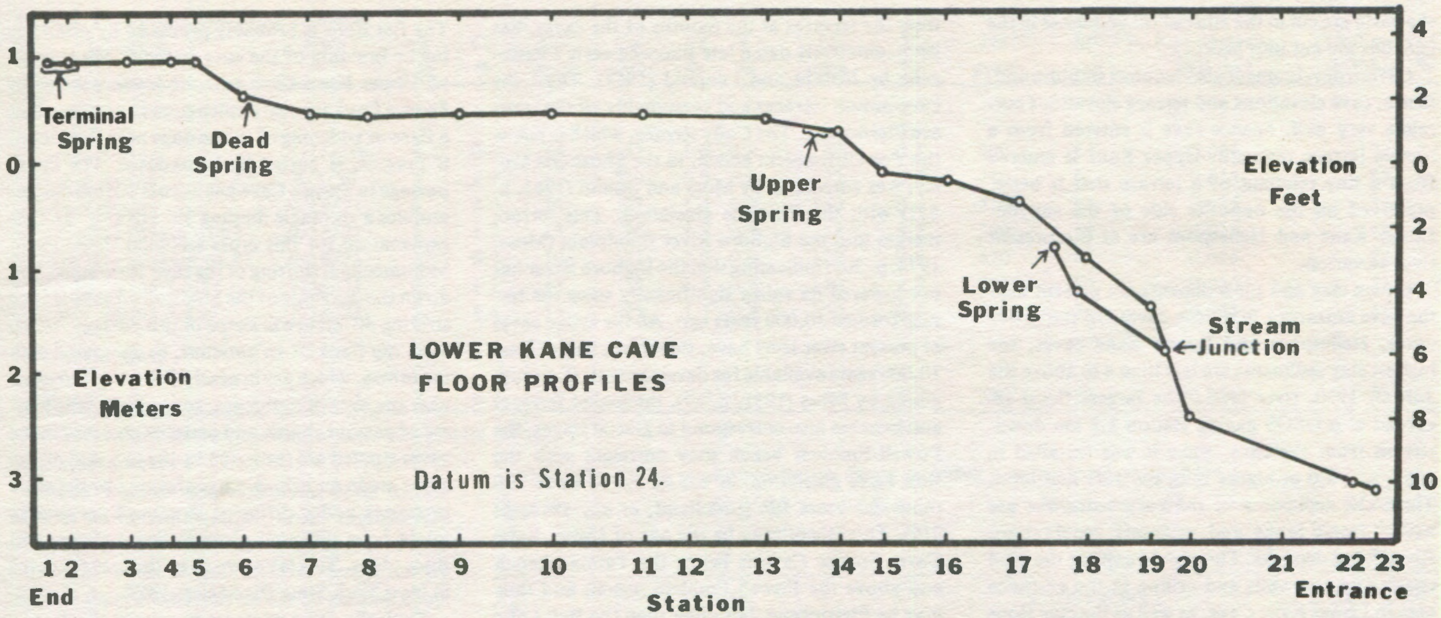

Figure 16. Longitudinal profiles of the streams in Lower Kane Cave. Vertical exaggeration is 25:1. The average slope is 1.3%.

tween the Kane Caves. This is apparently the way the collapse connection between the two passages of Spence Cave formed. Such a connection between the Kane caves would make Upper Kane an active cave again.

Side passages

The active caves have no side passages; of the dead caves, only Spence Cave has any side passages. Generally, side passages are rare in all the caves studied.

Flow features

Bretz (1942, pp. 720-733) lists pendants, ceiling channels, flutes, floor potholes, horizontal grooves in cave walls, incised meanders in cave walls, and domepits as features produced by flowing water. None of these features are present in any of the caves studied despite the presence of flowing water in the active caves.

The walls of the caves have enlarged upward and outward by continuous formation and spalling of gypsum crust. Streams were not in contact with the walls or ceiling during this process, so the cave walls and ceilings lack flow features. In Lower Kane Cave, the cave streams have several reaches over bedrock, but no flow features have formed in the stream bed; the reason for this is not known.

Phreatic features

According to Bretz (1942, pp. 698-720), spongework, wall and ceiling pockets, joint-determined wall and ceiling pockets, bedding and joint anastomoses, continuous rock spans, network patterns, boxwork, halftubes, and floor pockets are phreatic features, supposedly formed in water-filled caves when water circulation is very sluggish.

Phreatic features are fairly common in all the caves studied, despite the fact that there is no evidence that any of the caves were ever water-filled. However, the cave air reacts with the limestone in the active caves. It appears that the presence of a reactive fluid, either water or air, is all that is required for phreatic features to form.

Secondary mineral deposits

All the features described so far are primary deposits of the caves; that is, they were formed by the processes that formed the caves. Secondary deposits are those formed after the cave is no longer actively growing and hence provide no clues to the origin of the cave. Most secondary deposits are precipitates from seepage water.

Calcite. Stalactites and stalagmites are common in the dead caves. One wall of Upper Kane Cave near the entrance is covered with calcite flowstone containing a few thin gypsum bands. These bands indicate that the gypsum-forming process had several spurts of activity during deposition of the flowstone. Vugs of calcite crystals are present in Upper Kane Cave, but they appear to be a feature of the limestone rather than of the cave.

Silica. Some botryoidal silica is present on gypsum near the entrance of Upper Kane Cave. Siliceous sinter (and fluorite and barite) have been reported in several caves in the Big Horn and Pryor mountains (Bell, 1963, p. A12). The origin of these minerals is not certain. They may have been deposited while the caves were active; if so, they may be primary deposits.

Gypsum. In these caves, gypsum is a primary, not a secondary, mineral deposit. Some primary gypsum may re-crystallize after the cave is formed and therefore could be considered secondary.

OTHER REPLACEMENT-SOLUTION CAVES

Replacement-solution caves are fairly common in the Big Horn Basin. Near Cody, Wyoming, in the Rattlesnake Mountain Anticline, several caves containing bulbous masses of sulfur were discovered during the drilling of a tunnel (Love and Good, 1970, p. B21). According to local residents, there is a cavern containing gypsum crusts near Owl Creek, in the Warm Springs Anticline near Thermopolis.

More replacement-solution caves are located in the mountains along the north and east edges of the basin. Mystery Cave in the Pryor Mountains and Big Horn Caverns in the Big Horn Mountains have many replacement-solution features. The walls of both caves are covered with gypsum crusts in many places; thick gypsum mounds are present in Big Horn Caverns. Both caves consist of several large, gently-sloping linear passages with spongeworks of crawlways developed along the sides of the main passages. The main passages do not gradually narrow up gradient, but have blind terminations. No surface karst development is evident near these caves. The caves contain no flow features, but do contain phreatic features such as spongework and continuous rock spans. Both these caves are apparently replacement-solution caves.

In Nevada, at least one probable replacement-solution cave has been visited by the author: Whipple Cave in eastern Nevada. The limestone near this cave shows no evidence of karst development. The cave has a long linear main passage with a blind termination. Extensive collapse and dripstone deposition in the main passage have buried most of the gypsum crusts in this cave; however, some of the crust is still preserved along the sides of the passage and in side passages. Some phreatic features such as boxwork are evident and flow features are apparently absent.

Two possible replacement-solution caves have been visited by the author in New Mexico. These are Carlsbad Caverns and nearby Cottonwood Cave. The main passage in each of these caves is a long, linear passage with a blind termination. Gypsum crusts and mounds are common in both caves, as are phreatic features. Sulfur is present in cracks in the ceiling of Cottonwood Cave. Flow features appear to be absent in these caves.

It seems that replacement-solution caves are not a phenomenon restricted to the Big Horn Basin, but are found in several western states. Undoubtedly there are many more caves in western states that may be formed at least in part by replacement-solution.

It was noted that in a part of Lower Kane Cave there is no gypsum crust, because it is apparently being dissolved and removed by seepage water. Seepage into caves is much more prevalent in humid areas, such as the eastern United States. If replacement-solution is assisting in enlarging caves in humid areas, either by the hydrogen sulfide process proposed in this paper or by oxidation of metal sulfides as proposed by Morehouse (1968, p. 7), no gypsum crusts would exist as evidence of this process except in unusually dry passages. In Surprise Cave, New York, gypsum grows on the wall of one passage in the fall when the cave is driest and is washed away in the spring. Only chemical analyses of the cave waters would indicate conclusively whether sulfates are being produced in other caves. Unfortunately, in most published analyses of cave waters, either sulfate is not determined or an insensitive method of analysis was used. Thus, it is often not possible to determine from the literature if replacement-solution is contributing to the enlargement of other caves. Certainly, it is not advisable to assume that caves are formed only by solution of limestone as a carbonate until all possibilities have been tested. In many cases, several solution processes may be operating simultaneously.

CONCLUSIONS

The replacement-solution process can explain some features found in certain caves. The process provides an explanation for cave gypsum deposits and crusts and for rare cave sulfur deposits. Phreatic cavern features may not invariably indicate that a cave was once water-filled, as such features apparently can form in air-filled replacement-solution caves. The lack of solution features in some caves is a consequence of the fact that in replacement-solution caves, the walls are not in contact with water while the cavern is growing. Blind passage terminations may develop in caves fed from below.

Another interesting aspect of this study is the minerals in the cave springs. The deposition of uranium and vanadium in these caves located near to commercial uranium cave deposits certainly suggests that the uranium in the mines may have been concentrated in a manner similar to present-day deposition of these metals in the active caves.

Many of the metals in the active caves are found in commercial quantities in other caves. Transport and deposition in these other caves may have been by a similar process. Certainly caves in mines should be studied, as ore deposition may in some cases have been controlled by caves. However, one must be aware that in other cases oxidation of ore forms caves.

Areas with caves and little or no karst development may exist because the water that formed the caves came from below. The presence of karst development, of course, does not mean artesian cavern development is unimportant. Both processes may be important.

Further study is needed on many aspects of replacement-solution. More caves need to be studied, so the variety of features formed is better known. More needs to be known about the physical development of these caves: How they start and how they grow. Little is presently known about the slots that feed the caves, how they form, and their extent. In the future, research on cavern development should look at sulfate and carbonate in cave water, as many caves may have complex origins.

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The conclusions in this paper are my own and are not necessarily subscribed to by those mentioned above.

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APPENDIX

Analyses of ground water in the Big Horn Basin.

The analysis number on the table corresponds to the numbers in the Stiff diagrams in Fig. 1b. The source of data and the aquifer are indicated in Figure 1b and are not repeated here. *Average of four analyses. **Average of two analyses.

Number	Township, Range, Section	Concentrations in ppm							Reaction Value in epm						Depth in Meters
		Na	K	Ca	Mg	Cl	HCO_3	SO_4	Na + K	Ca	Mg	Cl	HCO_3	SO_4	
1.	T58N-R99-100W	-1652-		383	105	1000	730	2855	72	19.2	8.7	28	12.0	59	2017–2027
2.	T58N-R98W	-11-		207	69	Tr.	390	483	0.48	10.3	5.7	Tr.	6.4	10.1	966–1019
3.	T59N-R95W-26	-21-		172	55	4	249	473	0.93	8.6	4.5	0.11	4.1	9.8	0
4.	T57N-R99W-29	290	130	359	92	378	220	1510	15.9	17.9	7.6	10.7	3.6	31	2536
5.	T56N-R97W	-48-		352	70	27	270	971	2.1	17.6	5.8	0.76	4.4	20	1463–1465
6.	T54N-R94W-35	6	1	60	26	5.5	207	105	0.29	3.0	2.1	0.16	3.4	2.2	0
7.	T56N-R92W-31	6.2	2.1	52	17	1.4	205	35	0.32	2.6	1.40	0.04	3.4	0.73	0
8.	T58N-R95W-26	11	4	104	36	12.5	229	223	0.58	5.2	3.0	0.35	3.8	4.6	0
9.	T54N-R94W- 4	-8-		30	44	6	190	104	0.33	1.50	3.6	0.17	3.1	2.2	185–187
10.	T54N-R94W- 9	-198-		277	150	400	1120	246	8.6	13.8	12.3	11.3	18.3	5.1	167–172
11.	T54N-R94W-35	-24-		92	36	14	273	179	1.04	4.6	3.0	0.39	4.5	3.7	0
12.	T54N-R94W-35	-25-		62	24	8	244	92	1.08	3.1	1.97	0.23	4.0	1.91	0
13.	T52N-R102- 3	31.5	16	342	68	18.5	940.5	415	1.78	17.1	5.6	0.52	15.4	8.6	0
14.	T51N-R100	-406-		770	208	196	1780	1848	17.6	38	17.2	5.5	29	38	1330–1333
15.	T51N-R100	-273-		238	139	196	420	1093	11.7	11.8	11.5	5.5	6.9	23	1269–1309
16.	T51N-R93W-24	176	1.9	20	19	5.1	437	96	7.7	1.00	1.56	0.14	7.6	2.0	1191
17.	T50N-R91W-10	-593-		77	12	32	185	1284	26	3.8	0.99	0.90	3.0	27	1143–1271
18.*	T49N-R91W- 2	-226-		243	18.5	17	92.5	1033	9.8	12.2	1.52	0.48	1.51	21	722–763
19.*	T49N-R91W-12	-8.5-		144	41	20	230	318	0.37	7.2	3.4	0.56	3.8	6.6	700–719
20.	T49N-R88W-29	100	5.6	589	42	48	167	1560	4.5	29	3.5	1.35	2.7	32	0
21.	T49N-R89W-28	-15-		45	14	2	235	7	0.66	2.3	1.15	0.06	3.9	0.15	216–288
22.	T47N-R89W- 6	2.0	1.2	45	25	0.8	249	9.6	0.12	2.2	2.1	0.02	4.1	0.20	864
23.**	T47N-R88W- 1	1.85	0.75	37.5	16.5	1.5	195.5	1.0	0.10	1.87	1.36	0.04	3.2	0.02	0
24.	T47N-R90W- 8	2.0	1.1	46	25	1.0	250	10	0.12	2.3	2.1	0.03	4.1	0.21	2030
25.	T47N-R88W-16	2.9	2.0	40	28	1.0	250	17	0.18	2.2	2.3	0.03	4.1	0.35	320
26.	T46N-R98W-28	301	170	519	127	276	988	1480	17.4	26	10.6	7.8	16.2	31	1829
27.	T43N-R95W	262	49	385	76	328	766	769	12.7	19.2	6.3	9.2	12.6	16.0	0
28.	T43N-R93W-28	-130-		517	146.5	239	965	1006	5.6	26	12.1	6.7	15.8	21	—
29.	T42N-R88W-21	6.5	2.2	187	40	2.0	214	449	0.34	9.3	3.3	0.06	3.5	9.4	0
30.	T42N-R95W-25	41	7.4	146	50	39	377	276	1.97	7.3	4.1	1.10	6.2	5.8	0
31.	T42N-R88W-30	75	4.9	506	77	6.5	228	1420	3.4	25	6.3	0.18	3.7	30	0

Lauritzen, Stein-Erik (1981)—Statistical Symmetry Analysis of Scallops: *NSS Bulletin* **43**: 52-55.

STATISTICAL SYMMETRY ANALYSIS OF SCALLOPS

SUMMARY

A graphical method is suggested for measurement of inflection point angles in scallop copies on paper. Statistical analysis of the angle pairs from large numbers of scallops allow a quantitative judgement of the asymmetries of the scallops and, hence, of the flow direction. This is a safer method than judgement of scallop directions by inspection and finger touching.

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IN PHREATIC PASSAGES, the flow direction of water is not necessarily determined by the gradient of the passage or by the present surface topography above the cave. Because flow direction is a useful parameter for deducing former hydraulic gradients and, hence, conditions in the environment outside the cave, there is a need for precise methods to determine former flow directions in caves.

Except for interpretation from texture and grain size in certain cave sediments, the only method known for direct determination of paleocurrent direction and flow velocity in drained (fossil) cave conduits is the study of scallops and flutes (Coleman, 1949; Curl, 1966; Bögli, 1978). Scallops and flutes are formed by the interaction of turbulent moving solvents (or gases) and a soluble (or volatile) solid. Scallops are in some cases found on insoluble rocks, and evidently may therefore also be formed by erosion from sediment-loaded waters (Bock, 1913; Bögli, 1978; Allen, 1972).

Scallop patterns have been analyzed mathematically and simulated under controlled conditions (Curl, 1966, 1974, 1975; Blumberg and Curl, 1974; Goodchild and Ford, 1971; Wigley, 1972). These treatments are based on a solutional mechanism for scallop formation. What role erosion plays in the formation of scallops may be discussed, but for the present it is sufficient to realize that there is agreement in the literature as to the connection between flow direction and asymmetry of the scallop profile.

The ideal profile has invariably the steepest slope on the lee side of the crest (Fig. 1). Usually, the flow direction of the last phase of water-filling in the cave passage is judged by inspection of the scalloped surface, mainly by finger touching or from the size of the shadows when the surface is illuminated at a low angle (Coleman, 1949). This method is in most cases satisfactory, when the asymmetry of the scallops is very distinct. But, there are cases where this method fails because of its subjective character. The purpose of this paper is to introduce a more laborious but "objective" method.

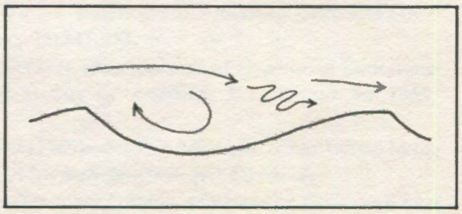
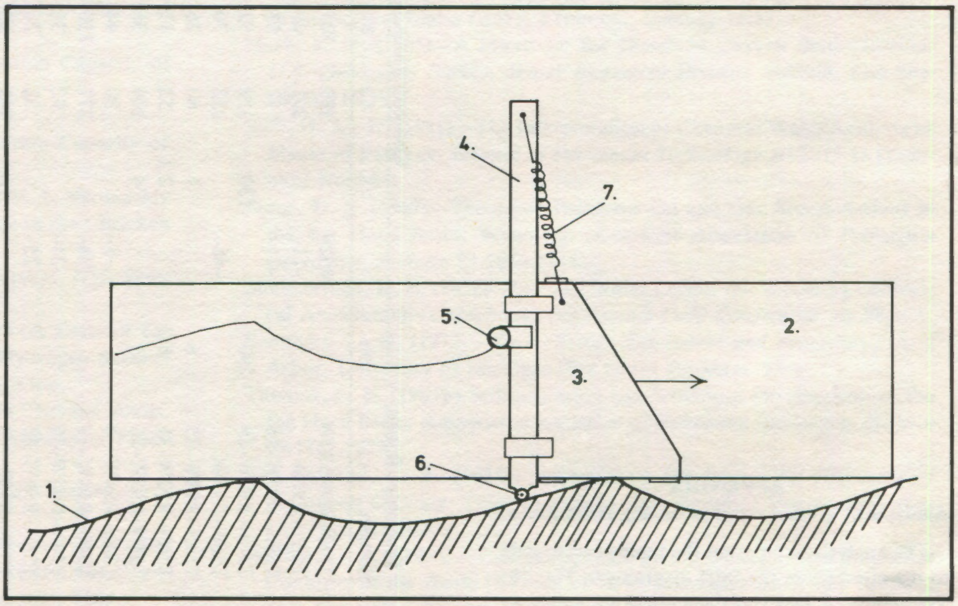


Figure 1. Liquid flow in the vicinity of a scallop. After Curl (1974).

Figure 2. Construction and use of the drafting apparatus for drawing scallop profile to scale. **1.** Scalloped rock surface. **2.** Board covered with drafting foil. **3.** Slide, moving parallel to the edge of the board. **4.** Rod, moving perpendicular to the edge. **5.** Pen. **6.** Small rubber wheel, rolling along the rock surface. **7.** Steel spring.



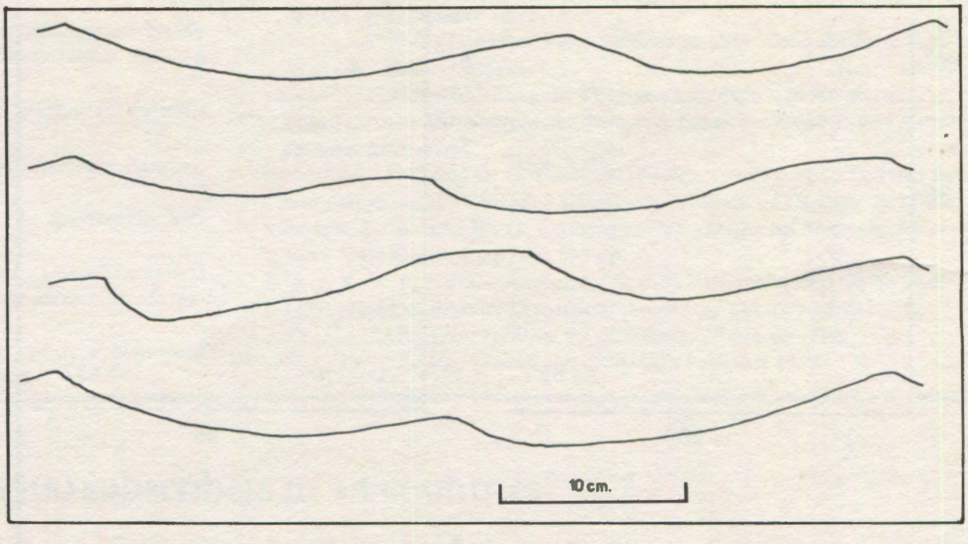


Figure 3. A collection of scallop profiles, as recorded by the apparatus.

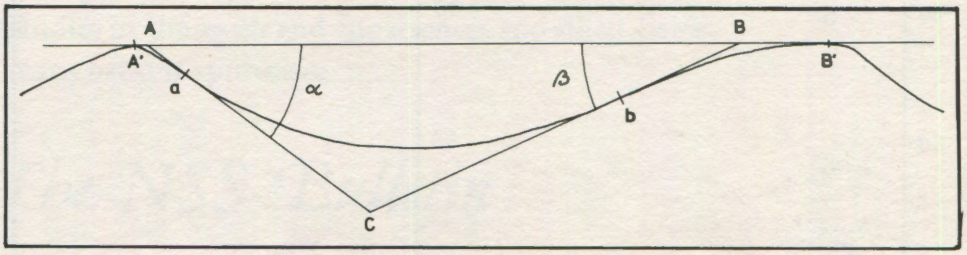


Figure 4. Definition of the inflection-point slope angles in a scallop. A "baseline," A'B', is drawn from crest to crest on the scallop graph. Tangents, AC and BC, are drawn through the two points of inflection (a and b) on the graph. In the triangle ABC,< BAC = α,< ABC = β. These two angles are measured with a protractor.

COPYING THE SCALLOP PROFILES

The basis for further analysis is an accurate copy of the scallop profile on paper. This may be done in several ways, but the simplest and most effective method when a larger sample has to be taken is the device described in Figure 2, designed by the author.

RESULTS AND DISCUSSION

Precision of the Method

The use of the drafting apparatus on scalloped surfaces gives, after some experience, accurate copies of the scallop profiles to scale, Figure 3. The slope angles in a scallop profile may be defined in several ways, and the following method was arbitrarily chosen. The asymmetry of the profiles is determined by measuring the angles between the base line and the tangents through the two points of inflection on the profile graph, Figure 4.

These two angles (α and β) are measured for a large number of scallops and are plotted as pairs on a "symmetry diagram", Figure 5. This gives a visual picture of the distribution of angles in the collection and allow a qualitative judgement of the presence or absence of a possible asymmetry.

The asymmetry, and hence the flow direction, is determined according to which side of the line $\alpha = \beta$ the plot is falling. The line $\alpha = \beta$ represent symmetrical scallop profiles and, hence, no flow direction. The measurements of α and β, however are not accurate and the precision of the method was checked by repeated drafting and angle measurements on the same few scallop profiles. This revealed a standard deviation of the measurements of 0.4 to 1.3°, with a weighted mean of 0.74°. The standard deviation did not show any systematic variation with respect to the angle size. The variance in precision seems to depend more on the quality of the drafts and the original scallops than on the angles themselves. It is evident from the diagram (Fig. 5) that the inherent variance in scallop inflection-point angles is severalfold (≤ 50) greater than the precision of measurement. The method is thus adequate for testing of scallop asymmetry without having the result obscured by the technique.

Statistical Testing for Asymmetry

The significance of the asymmetry in scallop collections may be tested in several ways, but as other quantitative uses of scallop dimensions have all been based on means (Curl, 1974), the testing for asymmetry is also based on this.

The null hypothesis is that $\hat{\alpha} = \hat{\beta}$, or $\hat{\alpha}$ to $\hat{\beta} = 0$, *i.e.*, the scallops are clustered around a point on the symmetry line ($\alpha = \beta$) and represent no flow direction at all. The variance of this difference is given by:

$$\mathrm{Var}\,(\hat{\alpha} - \hat{\beta}) = \delta^2(\hat{\alpha}) + \delta^2(\hat{\beta}) - 2\delta^2(\hat{\alpha}, \hat{\beta})$$

Assuming a normal distribution and the number of measurements $n \gtrsim 30$, then the null hypothesis can be tested by

$$q = [(\hat{\alpha} - \hat{\beta})^2 /\ \mathrm{Var}\,(\hat{\alpha} - \hat{\beta})]^{1/2} \geq 2.57$$

with confidence interval ~ 1%. For smaller data amounts, the use of Student's t-test is recommended (Snedecor and Cochran, 1978). The results from testing of the data in Figure 5 are listed in Table 1.

Practical Examples from Norway

The scallops treated in this discussion are all developed in a dense, homogeneous calcite-marble. Therefore, influence on scallop shape from variations in lithostructure is minimal and can be ignored.

Figure 5a represents data from a phreatic tube passage where inspection also gives a convincing result. The scallop collection is significantly asymmetric and shows no scatter in the opposite direction.

Figures 5b and 5c show measurements in a phreatic tube and its ceiling half-tube, respectively. The two directions are different, demonstrating that water moved in a direction opposite to that in the original tube, when the half-tube was subsequently incised into its roof. This difference in direction cannot be detected with confidence from inspection only. Other observations also suggest that the half-tube formation can only be explained by a sediment filling prior to or contemporary with the half-tube formation (Bretz, 1942). The inversion of the hydraulic gradient during the half-tube formation can be explained by the glacial history of the area (Svartisen, North Norway), when hydraulic gradients in the Pleistocene ice caps were probably superimposed onto the cave system. In this way, an uphill flow could be maintained for some time.

Figure 5d shows a sample of scallop profiles that show no significant direction. Since this profile is located not far from *5b* and *c* (same cave system), it is possible that this passage also experienced a reversed flow at some time. The question is whether the observed angle scattering is inherent in the scallops themselves (*5d* shows the largest scatter of them all) or if *5d* represents a population of scallops "frozen" in a process of adaptation to a reversed flow regime. This hypothesis may be investigated experimentally to see how scallop asymmetry behaves with a reversal of the flow direction.

Figure 6 shows the respective passage cross-sections.

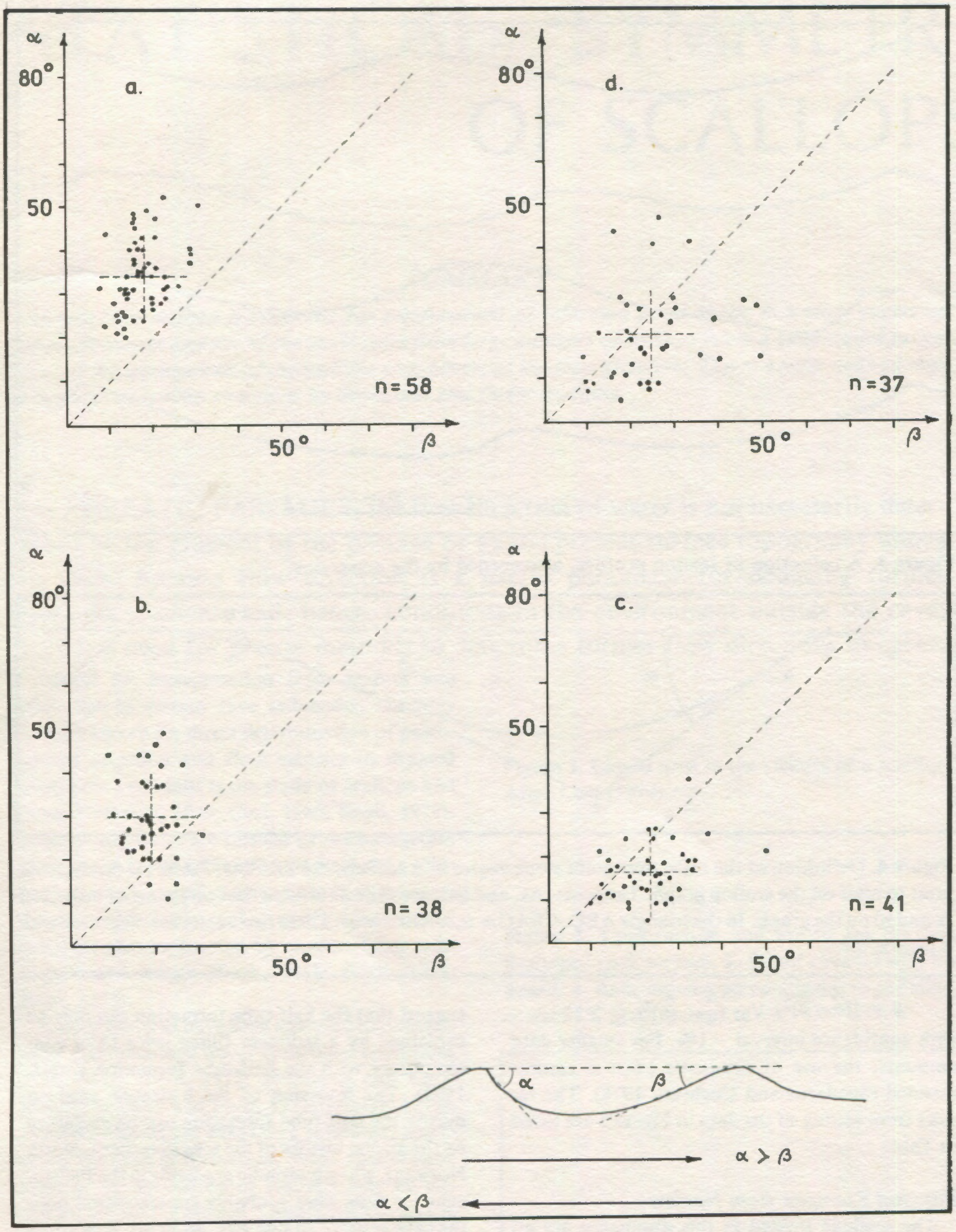


Figure 5. Symmetry diagrams of four different scallop collections from phreatic tube passages. a) Scallop collection where both inspection and angle analysis give a significant direction ($\hat{\alpha} > \hat{\beta}$, flow from left to right on the profile drawn below). b) Tube section with a significant direction, ($\hat{\alpha} > \hat{\beta}$). c) Collection from the same passage profile as b), but in the half-tube that is incised into the roof of the passage. The scallops have a significant direction but opposite to that of the trunk tube ($\hat{\alpha} < \hat{\beta}$). d) Roof and wall scallops that show no significant direction.

Table 1. Results of statistical tests of data from four scallop collections (see Fig. 5).

profiles	5a	5b	5c	5d
number of points	58	38	41	37
$\hat{\alpha}$	34.1	29.8	16.4	19.9
$\hat{\beta}$	18.0	18.8	23.5	24.5
$\sigma[\hat{\alpha}]$	1.115	2.315	0.948	3.053
$\sigma[\hat{\beta}]$	0.443	0.597	1.525	0.772
$\sigma[\hat{\alpha},\hat{\beta}]$	0.195	−0.102	0.443	2.846
q	13.1	4.6	3.7	0.90
significant asymmetry at ≤ 1% level	yes	yes	yes	no

CONCLUSION

Statistical symmetry analysis is superior to field inspection when considering scallop patterns that show variable asymmetry and where a large number of observations is required to obtain a reliable estimate of paleocurrent direction. As applied in the cases mentioned, symmetry analysis can be used in deducing past hydraulic gradients in cave systems.

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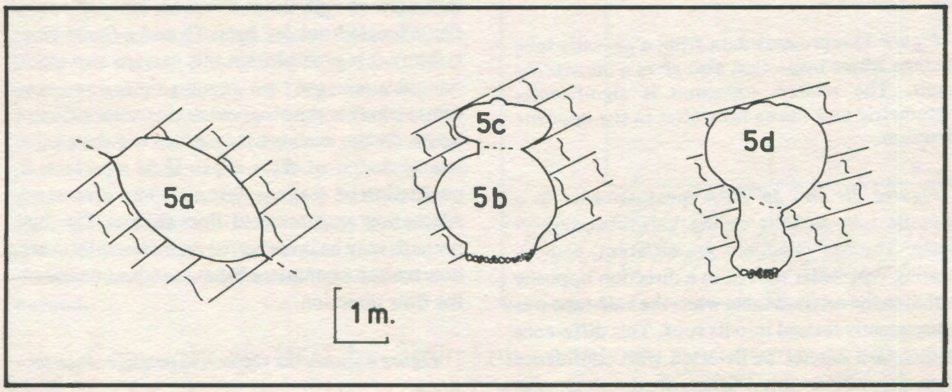

Figure 6. Cross-sections of the cave passages where the scallop profiles were taken.

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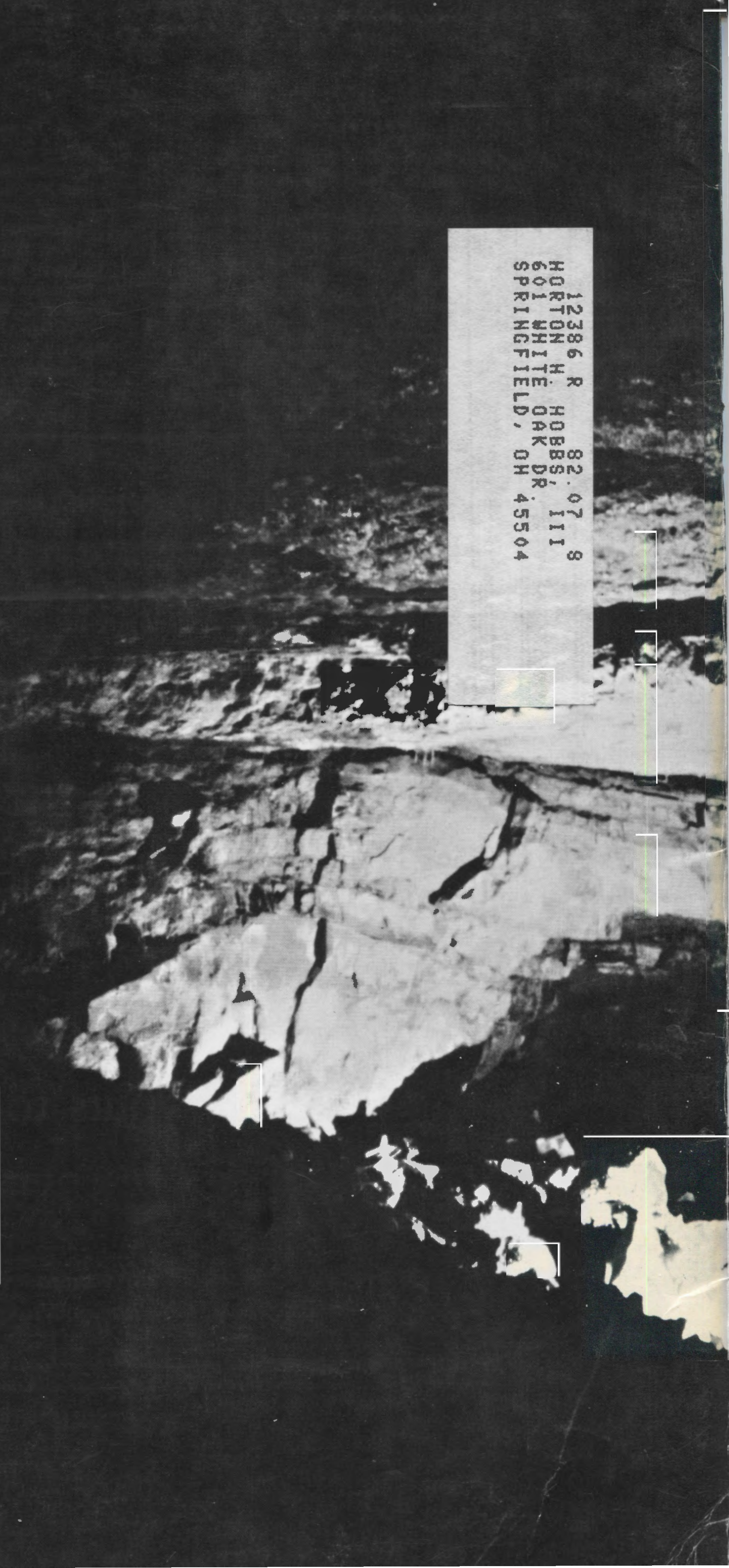

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