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NATIONAL SPELEOLOGICAL SOCIETY

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Firn Caves in the Volcanic Craters of Mount Rainier, Washington

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ABSTRACT

Sub-ice fumaroles and warm air currents form and maintain over 2 km of cave passages beneath ice filling the summit craters of Mount Rainier. Passage size increases from 1970 to 1973 indicate recent, minor, heat flow increases. Large heat flow decreases would allow plastic flowage to close passages and large increases would produce enlargement, collapse, and large crater lakes. Complete melting of summit ice would produce about 1.1 billion liters of water in the west crater and 7.4 billion liters in the east crater, creating a serious potential geological hazard.

The caves are called firn caves, because ice density ranges from 0.55 to 0.81 gm/cm^3. Ice temperatures as low as $-10^\circ C$ partly account for low densities, high viscosity ice, and open passages as deep as 100 m below the east crater snow surface.

Subsiding ice replaces walls and ceilings that are melted back 2.0 to 3.5 m/year. Subsidence, melting, and snowfall are in dynamic equilibrium. Thus, cave dimensions and the snow surface within the crater remain relatively constant. The discovery of a climber's glove and the debris from a 1959 expedition shows that subsiding ice reaches deeper cave areas in a few decades.

Entrance passages lead down the crater slope and perimeter passages parallel elevation contours. An east crater perimeter passage 50 m below the surface in its central part is 915 m long and winds three-fourths of the way around the crater. Narrow passages lead downward from the main perimeter passage in each crater to large grottos. A 40 m long lake in the west crater grotto is the highest crater lake in North America.

INTRODUCTION

General Statement

A long-term study of North America's highest and the world's largest known cave system developed in a volcanic crater was initiated at Mount Rainier (Fig. 1) in 1970 and carried on through 1973. The main objectives were to understand its speleogenesis, geomorphic features, and potential use as a volcanic activity indicator. The geological processes operating in the summit caves produce unique geomorphic features and conditions that enable small geothermal heat release changes to be detected.

The caves were first entered by Hazard Stevens and P. B. Van Trump in 1870, on the first documented summit climb (Stevens, 1876). An overnight stay, during which they huddled over a hot fumarole in one of the caves, no doubt saved their lives. Initial cave explorations by Stevens and Van Trump, as well as later observations by Russell (1898) and Flett (1912), were confined to areas just inside the entrances. Thousands have climbed the mountain since and some have recently explored parts of the extensive cave system (Mitchell, 1969; Molenaar, 1971, p. 185-187; Nelson, 1971; and Whittaker, 1970), but no thorough exploration occurred before our 1970 expedition. The resulting report (Kiver and Mumma, 1971a) contains a preliminary description of deeper areas of the east crater caves. The main conclusions of subsequent expeditions, except for 1973, have been presented in a number of abstracts (Kiver and Mumma, 1971b; Mumma and Kiver, 1971; Kiver and Steele, 1971; Kiver and Lokey, 1973; Lokey and others, 1972; and Kiver and others, 1973) and short reports (Kiver and Mumma, 1971a; 1971c; Kiver and Steele, 1971; Miller, 1974; Miller and others, 1974). They are being brought together, along with detailed descriptions of cave features, for the first time in this report. Documentation of present cave passage locations and conditions will assist in detecting major changes in the cave system and in volcanic heat release by the dormant volcano.

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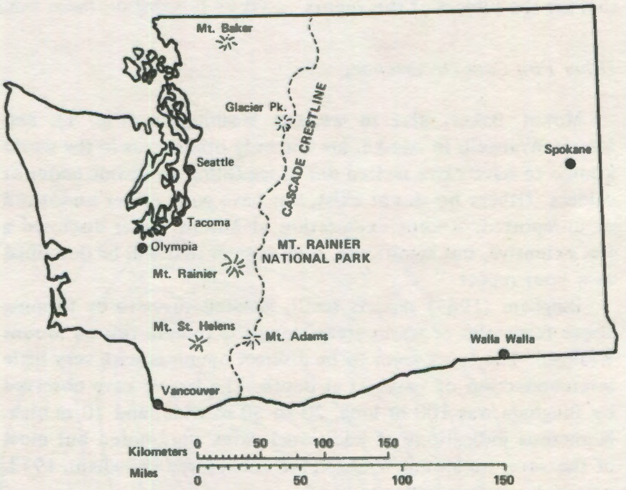

Fig. 1. Washington state index map. (Art work by Fred Munich)

Geologic and Geomorphic Setting

Mount Rainier towers 2,700 m above the surrounding peaks of the Cascade Mountains. Tertiary clastic and volcanic rocks are intruded by a Pliocene granodiorite batholith which, in turn, is intruded and overlain by Late Pleistocene-Holocene igneous rocks. Numerous eruptions of pyroclastic material and (mostly) andesitic lavas built the massive cone during recent geologic time. The Pleistocene volcano had been modified considerably by landslides and glaciers by the time the present uneroded slopes of the summit cone (Fig. 2) were constructed during the last few hundred to two thousand years (Crandell and Mullineaux, 1967). Mount Rainier is the best studied of the chain of volcanoes that extends from southern Canada to northern California along the Cascade Mountain crest. Excellent geologic summaries can be found in Fiske and others (1963) and in Crandell (1969a and 1969b).

The two overlapping craters at the top of the Holocene cone were caused by an easterly shift of eruptive activity during the last major summit eruption. The more recently formed east crater is 420 m in diameter and is shaped like a bowl tipped to the east (Fig. 3). The low side of the bowl on the east has an elevation of 4,328 m and the high side (Columbia Crest, the highest point on the mountain) an elevation of 4,393 m. Where the bedrock of the east crater rim is clear of snow, it stands 5 to 35 m above the surface of the estimated 120 m thick ice-and-snow plug that fills the crater (Fig. 4). A closed depression 160 by 285 m in diameter and 9 m deep on the east side (Fig. 5) lies above the deepest part of the crater floor and the most recently active vent of the volcano.

The west crater is 250 by 310 m in diameter and is truncated on its east side by the rim of the east crater. A small arc of the crater rim stands 3 m above the northwest edge of the relatively flat surface of the 48 m-thick snow-and-ice plug that fills the crater (figs. 4 and 5). A small, prominent depression about 50 m in diameter and 2 m deep occurs in the ice surface toward the north end of the crater. Bedrock areas where geothermal heat completely melts the snow and ice cover during the summer are visible only from the air or by climbers at the summit. Where hot areas are covered by thicker ice, melting produces the caves that are the subject of this report.

Other Firn Cave Occurrences

Mount Baker, also in western Washington (Fig. 1), and Mount Wrangell, in Alaska, are the only other areas in the world known to have caves melted out of ice filling a volcanic crater or caldera. Others no doubt exist, but have gone either unnoticed or unreported. Recent exploration at Mount Baker disclosed a less extensive, but significant, cave system that will be described in a later report.

Bingham (1967) reports small, isolated caves to be forming above fumaroles or warm areas along the caldera rim on Mount Wrangell. The caves seem to be discreet openings with very little interconnection of passages at depth. The largest cave observed by Bingham was 100 m long, 20 to 30 m wide, and 10 m high. Numerous indications of additional caves were noted but most of the caves on Mount Wrangell are unexplored (Bingham, 1972, personal communication).

Evolution of the Cave System

The development of the present cave system postdates the last summit eruption, when heat would have melted most or all of the crater ice fill. At least 14 small eruptions reportedly occurred between 1820 and 1894 (Hopson and others, 1962) and an ash layer dating from 1820 to 1854 occurs near the Emmons Glacier (Mullineaux and others, 1969). If one or more of these eruptions involved a summit eruption, then the consequent heat changes would have melted most or all of the ice in the summit craters and, perhaps, would have formed a crater lake.

Subsequent to the eruption, the craters filled with snow and the formation of the modern caves began. Cavities forming above fumaroles and areas of warm ground were elongated and became enlarged into grottos as warmed air rose upslope toward the surface. Many of the grottos coalesced, gradually creating the present large system of interconnected caves. Intersection of caves with the surface allowed rising air masses to escape more readily. These caves, some of which descend to depths of 76 m, became the entrance passages. Even freer air circulation was promoted when more than one entrance passage intersected the nearly horizontal perimeter passage forming above a line of suspected ring fractures on the crater floor. Cold air descends in one or more of the entrance passages to replace the warm air escaping out of other entrances.* The system gradually became enlarged to include over 2.2 km of cave passages and attained a state of approximate equilibrium with heat release. The development of the cave system was well under way, if not complete, by the late Nineteenth Century, when Stevens (1876) and Russell (1898) described the crater and cave entrances.

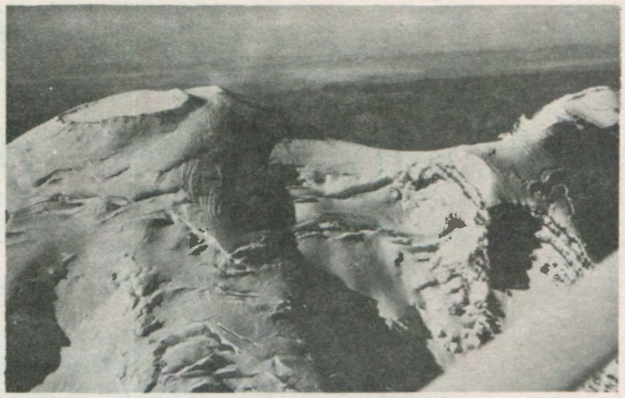

Fig. 2. View southwest across the summit of Mount Rainier, showing uneroded slopes of recent summit cone on left and truncated strata of older cone on right. (National Park Service photo)

DESCRIPTION OF SUMMIT CAVES

"Firn" Caves

The uniqueness of a cave formed by melting ice or snow in an area of high heatflow justifies the application to it of a special term. Non-genetic names considered include ice and firn caves; genetic names include steam, ablation, and geothermal caves.

The most desirable genetic classification, in our opinion, would divide caves of this type into 1) *geothermal ablation caves*, where the air temperature is at least partly dependent on a geothermal heat source, and 2) *atmospheric ablation caves*, where elevated air temperature depends on transfer of heat by atmospheric or running water sources. Thus, the summit caves associated with fumaroles and warm ground temperatures would be geothermal ablation caves, while the Paradise Ice Caves, located in a stagnant glacier at lower elevations on the mountain (Anderson and Halliday, 1969), would be atmospheric ablation caves. The term, "steam cave," has been used informally by mountain climbers and Park Service personnel for many years, but it lacks preciseness of description because bedrock caves can also contain steam.

Non-genetic terms emphasize the material in which the caves are formed and are used extensively in this paper. The term, "ice cave," would be useful, but is often used by speleologists with reference to bedrock or talus caves in which ice remains

*** For example, the flow of air in the east crater in August 1973 was down in the passages along the north rim and up along the south rim passages.**

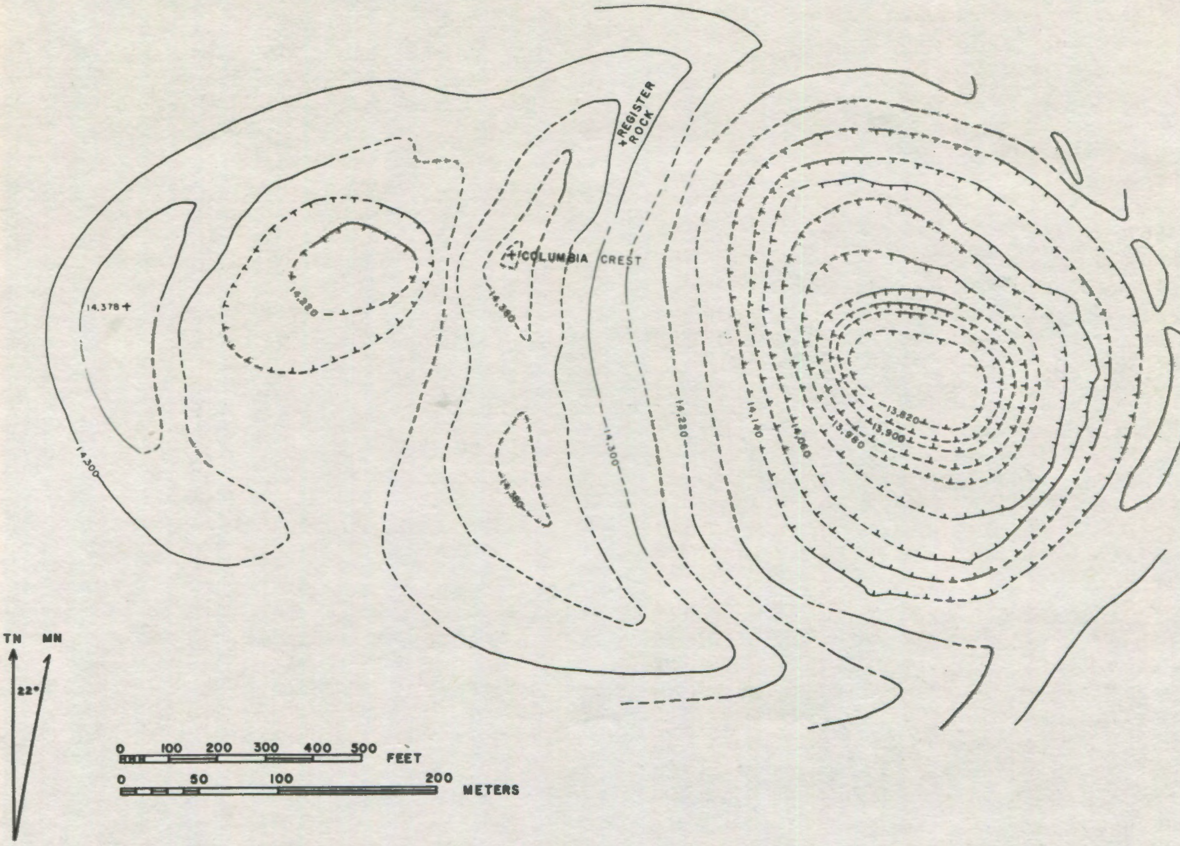

Fig. 3. Sub-ice contour map of summit (contour interval 40 ft).

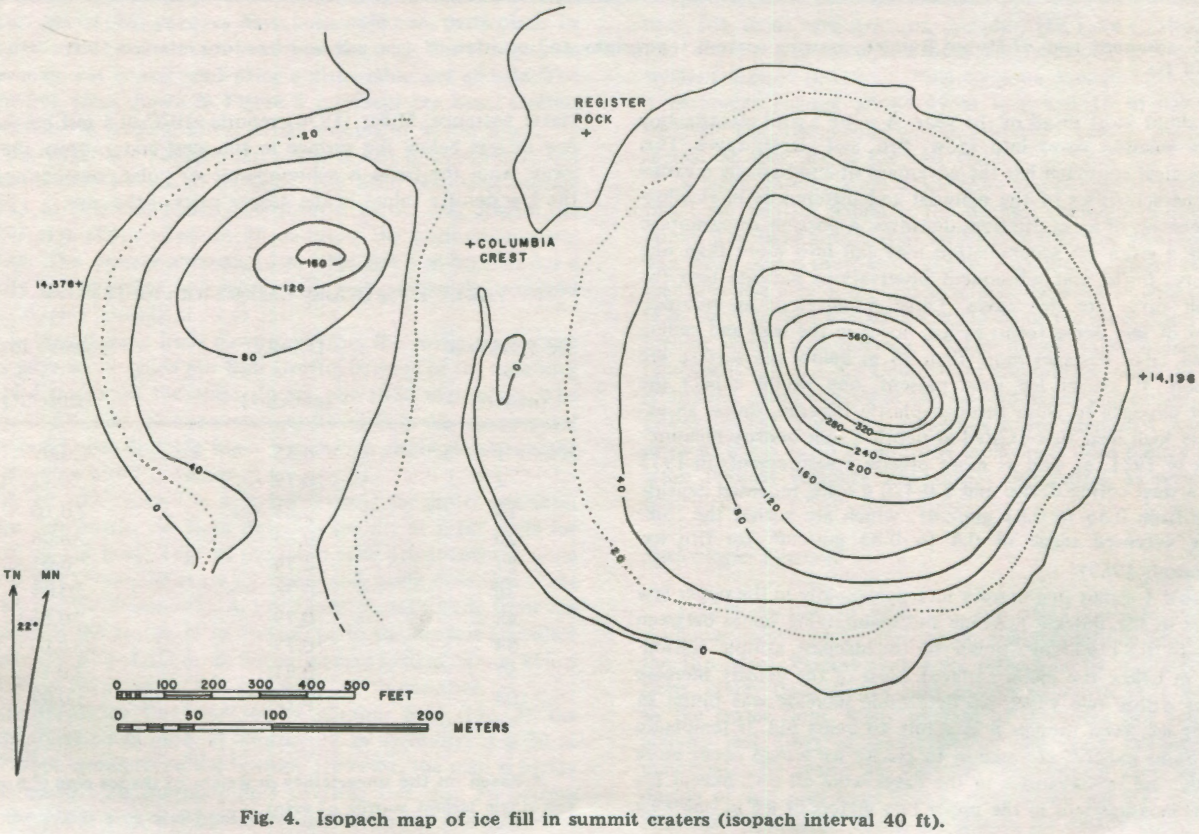

Fig. 4. Isopach map of ice fill in summit craters (isopach interval 40 ft).

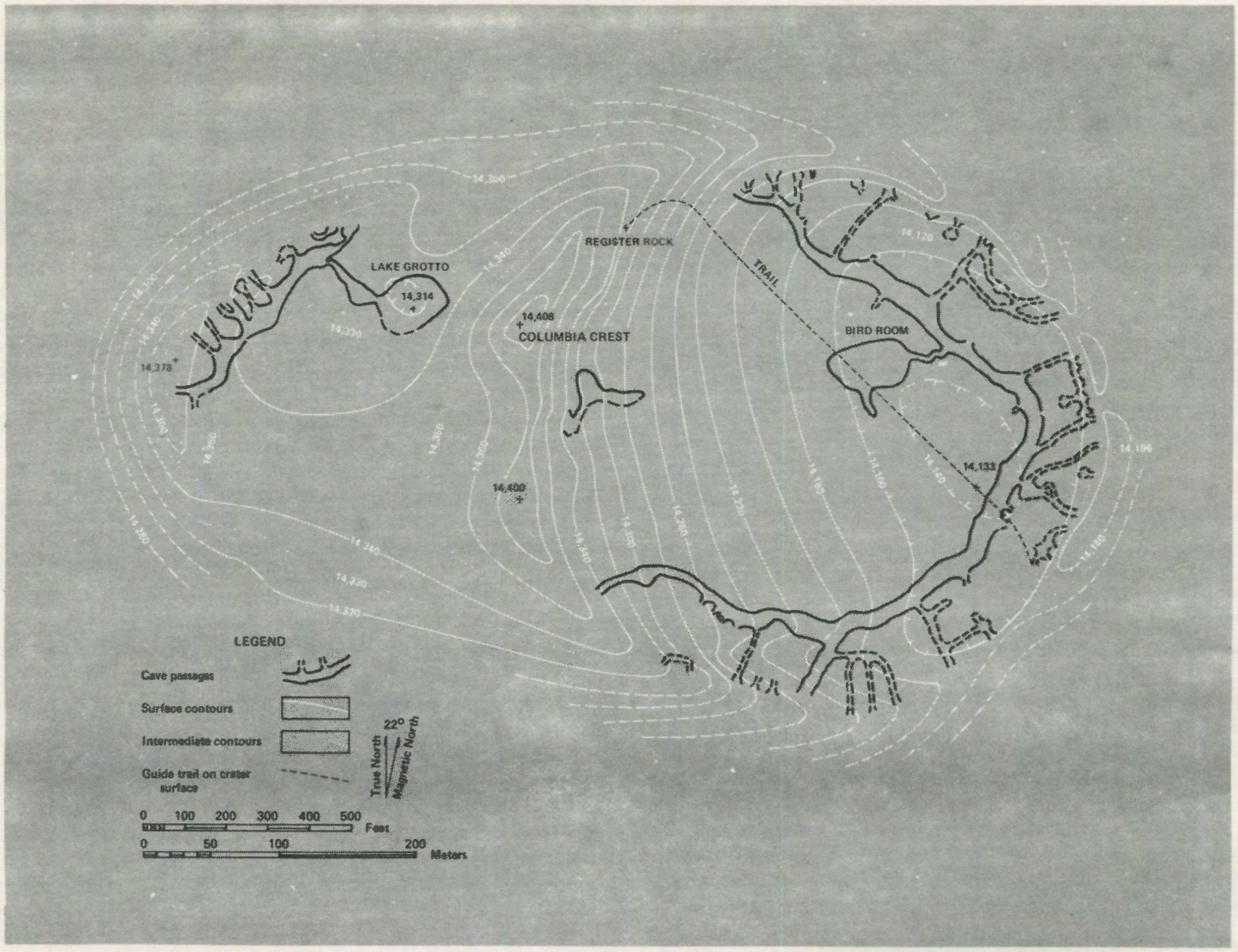

Fig. 5. Summit map of Mount Rainier, showing summit topography and position of cave passages (contour intervals 10 ft and 20 ft).

throughout most or all of the year. A more useful classification divides ablation caves into snow, firn, and glacier caves. This glaciological approach has the advantage of emphasizing measurable characteristics of the material and differences in rheological behavior of ice at different densities. A possible disadvantage is that a given ice accumulation may fall into more than one category. Initial and subsequent observations indicate that the summit caves are firn caves. Large numbers of air bubbles, typical of less dense forms of ice, occur in the wall and ceiling ice and large passages more than 35 m below the surface are common. If glacier ice were present, one would expect the deeper passages to close through plastic flowage, unless an extremely high heat flux existed at depth. Crude density measurements in 1971, as well as more precise measurements in 1973 using a steel coring device and a 0-450 g scale, recorded density values from 0.55 to 0.81 gm/cm³, which are within the commonly accepted range of 0.4 to 0.85 gm/cm³ for firn ice (Patterson, 1969).

Table 1 shows that density increases rapidly in the upper few meters of ice, quickly reaching and maintaining values between 0.61 to 0.81 gm/cm³ down to the deepest sample station (104 m below the crater surface). Most of the density increase occurs within two years. No detectable increase was found in deeper ice, even though it is about 40 years old. If temperate conditions existed, conversion to glacier ice would occur more rapidly. Ice temperatures in the caves were all 0°C, except for a -5°C measurement in the upper two meters of ice in one west crater entrance. Miller (1970) reports -10°C in a test pit dug a few meters below the surface in the west crater. Thus, the ice away from the caves is subtemperate or polar, accounting for the low density values in the deeper parts of the cave.

TABLE 1. Ice Density Changes With Ice Thickness.

Ice Thickness (meters)	Density (gm/cm³)	Estimated Error* (gm/cm³)
1	0.55	±0.09
2	0.79	±0.11
13	0.61	±0.10
20	0.81	±0.06
27	0.79	±0.07
36	0.76	±0.09
45	0.79	±0.062
54	0.77	±0.07
57	0.81	±0.14
104	0.75	±0.09

* Based on the uncertainty in weight of the ice core (±5 gm) the single largest source of error.

Types of Cave Passages

Three major types of cave passages occur in the summit craters. *Perimeter passages* are long passages that tend to parallel the contours of the crater sides. *Entrance passages* are oriented perpendicular to contour lines. The third type is enlarged areas called *grottoes*, or *rooms*. Two very large grottoes have been discovered so far, the Bird Grotto in the east crater and the Lake Grotto in the west crater.

Cave passages are produced and maintained by warm air and steam moving along the contact between the crater floor and the overlying ice fill (Fig. 6). Upward movement of warm air produces and maintains the entrance passages, lateral movements create perimeter passages, and restricted movement of a warm-air mass causes grottoes. The large grottoes are the deepest accessible portions of the caves in both craters and each has only one narrow passage connecting it to the remainder of the cave system. Thus, a restricted exchange of rising warm air and descending cool air occurs and the trapped warm air melts out rooms as much as 54 m long and 41 m wide. Air temperatures are 4 to 5°C in the deep grottoes, slightly less in perimeter passages, and at or near 0°C in entrance passages. The atmosphere in grottoes is clear when first entered but, because of restricted circulation, becomes misty from the condensation of breath after a few minutes. Perimeter and entrance passages are clear, except for areas where fumarole activity is very high.

Cave Size and Extent

The extent of the caves was determined by mapping the system with a metal tape and a tripod-mounted Brunton compass. The resulting map (Fig. 5) shows over 2.2 km of surveyed and approximately located passages (the latter indicated by dashed lines). Passage locations are estimated to be known within three meters (vertically and horizontally) of their true position in the west crater and within ten meters in the east crater. Other, unmapped, passages have been explored, particularly in the east crater, but are not shown on the map. The topographic base map was constructed using a plane table and alidade. The snow-free areas shown in Figure 5 represent the usual summit conditions in August, after partial melting of winter snow.

The larger and younger east crater is 420 m in diameter and contains over 1.8 km of passages. The main perimeter passage is 915 m long and winds three-fourths of the way around the crater at a depth of about 70 m below the northernmost entrance. This passage averages 15 m wide and 6 m high. It has a gently arched ceiling, a downslope ice wall, and a debris-covered floor sloping laterally at 30 to 35°.

A small passage leads downward from the perimeter passage to a large room called the Bird Grotto, because of the discovery of bird bones on the floor during our 1970 expedition. The room is 54 m by 36 m in area and 21 m high. Air temperature here is 4°C and dripping water is common although no running or standing water is visible.

A flat area about 7 m wide occurs near the grotto entrance. From this bench, the floor descends sharply at a 38° angle for 21 m to the lower edge of the room. The debris-covered slope in places exceeds the angle of repose and easily dislodged rocks make descent hazardous. A low, wide passage leads from the bottom of the grotto 25 m downslope to the deepest accessible point (elevation 4,215 m) in the entire cave system, 104 m below the ice surface and 123 m below the north entrance.

Extrapolating from the sub-ice contour map (Fig. 3), the crater floor could descend another 25 m vertically, if a 30 to 35° slope continues to the bottom. However, the map is highly generalized with control points covering less than 50 percent of the crater area and limited to the crater rim and cavernous areas, so the estimate of maximum crater depth can only be approximate.

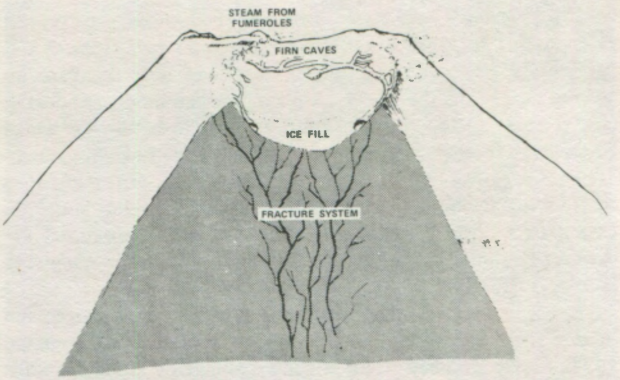

Fig. 6. Diagrammatic sketch of Mount Rainier, showing caves within crater ice fill. (Art work by Fred Munich)

About 16 entrance passages descend from the edge of the crater along a 30 to 35° slope into the perimeter passage. The entrance passages have gently arched ceilings and, in contrast to perimeter passages, two ice walls rather than one (Fig. 7). They are 3 to 8 m wide and 2 to 4 m high. Most entrance passages are unsurveyed. Some of the higher perimeter passages leading from these unsurveyed entrance passages are not completely explored.

Along the west edge of the east crater are a number of unmapped entrance passages and a high perimeter passage. The entrance passage just east of Columbia Crest descends an estimated 70 m over some very steep pitches, requiring the use of rope. This passage is unique in that it is the only place in the crater where running water can be observed. The small stream continues beyond the point where diminishing passage height prevents further exploration.

The west crater is 200 by 325 m in diameter. It contains only 305 m of cave passages, including the Lake Grotto. Thermal activity and cave entrances are restricted to the northwestern edge of the crater. Nine entrance passages lead down to a perimeter passage about 49 m long and 15 m below the surface. The dimensions of the perimeter passage are not as regular as they are in the east crater.

A small passage leads down from the perimeter passage to the Lake Grotto, which is the deepest chamber in the west crater. The Lake Grotto is 52 by 41 m in diameter and 7 m high. The floor slopes southwestward to a small lake, 40 by 10 m in diameter and 5 m deep, that is terminated by a vertical ice wall (Fig. 8). The lake nestled on the floor of the west crater is the highest crater lake (elevation 4,329 m) in North America. Its temperature is 0.5°C. Underwater exploration failed to reveal additional passages beyond the ice wall (Kiver and Lokey, 1973; Lokey, 1973). There are numerous active fumaroles that emit gases as warm as 50°C. Moisture condensed from human respiration reduces visibility and causes difficulties, as previously described.

Speleologic Features

The interaction of heat, ice, and other geologic processes produces many unique speleologic and geomorphic features in the firn caves. Passage shapes are influenced by fumarole locations, crater floor slopes, and passage orientations in relation to the crater floor. Many sections of the caves lack fumarole activity and areas of warm ground and must, therefore, be maintained by warm air currents. The longest cave segment devoid of apparent local heat sources is the last 200 m of the perimeter passage on the south side of the east crater.

The cave walls have a fluted or scalloped appearance typical of snow and ice walls shielded from the sun's rays (Figs. 7 and 8). Fluting also occurs in the glacier caves near the Paradise Visitor Center on the lower flanks of the mountain (Anderson and Halliday, 1969) and in caves formed in large snowbanks that persist through one or more summer seasons. Flutes resemble a variety of suncup (a type of ablation depression) that forms on the surface of some perennial snowmasses, but must be caused by a slightly different process. Suncups require dry air to form (Manning, 1967, p. 401), a condition seldom occurring in the firn caves.

Broad, dome-like melt structures (Fig. 9), much larger and deeper than flutes, are called steam cups (Kiver and Mumma, 1971a, p. 322). They occur on low ceilings or along ice walls and are caused by the close proximity of fumaroles. There are about 12 steam cups in the east crater and one in the west crater, the last at the location where Stevens and Van Trump probably spent their chilly night in 1870. Steam cups are one of the features that would be particularly useful in detecting changes in heat release at the summit.

The cave floors are covered by hydrothermally altered clay of unknown composition and by bouldery talus derived from the crater floor and rim. Bedrock exposures are rare. Debris is particularly abundant in the east crater perimeter passage, where longer and larger entrance passages funnel considerable material to the lower levels. Boulders 30 cm in diameter are common but few reach 2 m. The debris is stable at the bottom of the perimeter passage, but is at the angle of repose in most other places. Numerous boulders were accidentally dislodged during our study, but no natural boulder falls were observed.

Fig. 7. View south across subterranean crater lake towards ice wall. (author)

Ice-barred talus cones form in the east crater perimeter passage at the lower ends of the long, descending entrance passages and a debris ridge is often present against or near the downslope ice wall. Some debris ridges have ice cores. These probably evolved from ice-barred talus cones, as suggested in Figure 10. Thick debris accumulations at the toes of talus fans insulate the base of the ice wall and cause ice ledges to develop, as the ice wall ablates more rapidly above than at its base. The uneven melting of the ice ledge, combined with slumping, then produces ice-cored debris ridges (Fig. 10c). Other debris ridges, without ice cores, may form when momentum of rolling debris is great enough to create an ice-barred debris ridge as suggested in Figure 10a.

A unique arrangement of passages and air current movements has caused an ice floor (Fig. 11) to form 180 m from the southwest entrance of the perimeter passage in the east crater. At this location, just west of its intersection with a large entrance

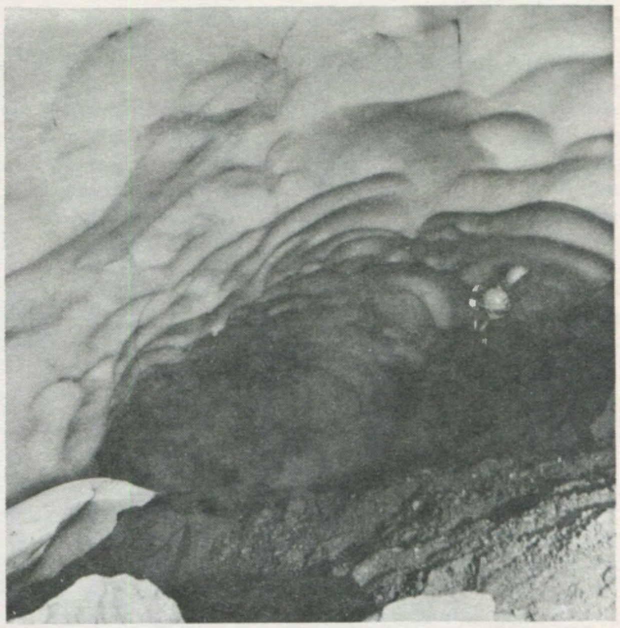

Fig. 8. View down an east crater entrance passage. (Truman Sherk photo)

passage (Fig. 5), the 25 m-wide perimeter passage narrows to 10 m. Most of the clockwise-moving warm air circulating through the perimeter passage turns and escapes out this entrance passage. Some of the warm air, however, continues straight ahead and enters the smaller segment of perimeter passage. In doing so, it impinges against the ice wall just west of the perimeter passage-entrance passage junction, causing the ice wall to retreat and form a large ice ledge. The ledge is 4 m wide on the east. It becomes less than 1 m wide on the west before disappearing from view underneath the rock debris covering the floor. The perimeter passage west of this junction lacks fumarole activity and is maintained as an open passage by the same air currents that maintain the ice ledge.

The ice ledge has not changed appreciably since 1970 (when it was first observed), excepting at its western end, where a small, ice-cored debris ridge about 30 cm high was present on the ice ledge in 1973. No other large ice ledges occur in deeper areas of the cave, but some temporary ice floors form near cave entrances where collapse, snow drifts, and uneven enlargement occur.

Fig. 9. Steam cup in east crater perimeter passage. (author)

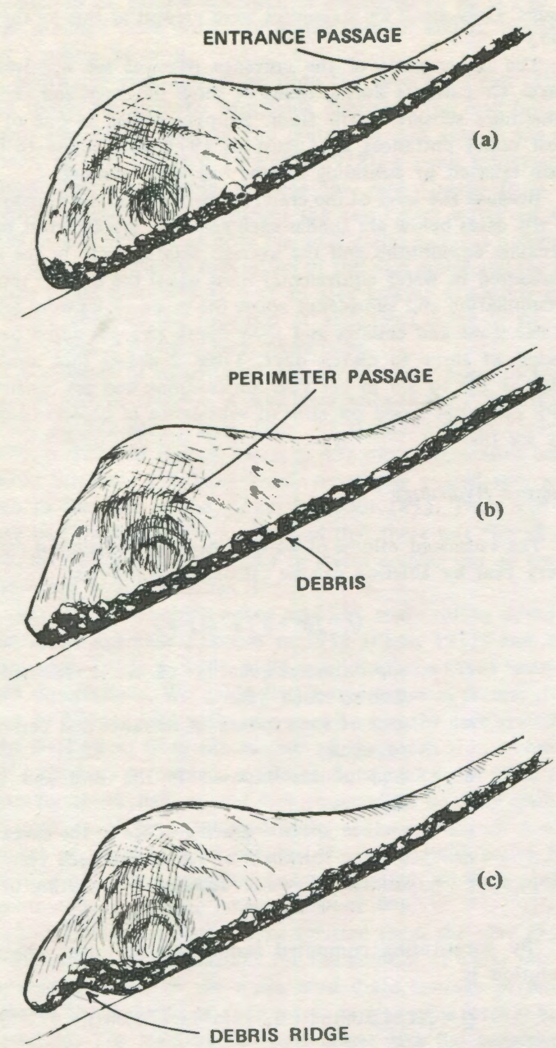

Fig. 10. Suggested debris ridge genesis: a) Ice-barred talus cone and debris ridge forms. b) Wall retreat and ice ledge formation. c) Wall retreat, ice ledge disintegration, and formation of ice-cored debris ridge. (Art work by Fred Munich)

Cave collapse features are very rare, except near the rapidly changing entrances, where walls and ceilings weakened by ablation occasionally fall. Only two instances of falling ice masses deeper in the caves were noted in 4 years of observation. An ice block weighing approximately 80 kg was discovered in the east crater in 1970 about 40 m from the southwest entrance. It had melted away by August, 1971. An elongate ice block discovered just west of the Lake Grotto entrance in the west crater in 1972 had not melted appreciably by 1973. A crevasse in this area probably determined the elongate shape of the block.

Ice speleothems are rare, except near entrances where meltwater freezes into stalactites and columns. Ground temperatures are usually too high to allow stalagmite formation. Air temperatures are a constant 4 to 5°C in the large grottoes and in the deep east crater perimeter passage, but fluctuate considerably with outside conditions near the entrances and in the higher perimeter passages. Winter cave temperature conditions are unknown because access is prevented by drifted snow and the closure of some entrances by subsiding ice.

Freezing conditions during early summer and fall extend down some entrance passages almost into the perimeter passage.

Fig. 11. Ice ledge exposed in east crater perimeter passage. (author)

A large snowdrift often plugs the first 60 m of the northwest end of the perimeter passage in the east crater. Drifted snow and ice coatings on the floor occasionally extend down entrance passages nearly to the perimeter passage.

In the west crater cave, near the Lake Grotto entrance passage, there is often a northeast-trending row of ice stalactites and columns (Fig. 12). The alignment of the ice speleothems and the fallen elongate ice block described previously suggests localization of both features by a crevasse. The ice speleothems appear rapidly when a number of warm days during the summer intensify surface ablation and allow water to percolate through the crevasses. Small crevasses, a few centimeters wide, occur at the surface above the ice speleothem location but are usually hidden by the previous winter's snowfall.

Fig. 12. Ice speleothems in west crater passage. (National Park Service photo)

DYNAMIC EQUILIBRIUM IN THE CAVE SYSTEMS

General Statement

With very few exceptions, the size and appearance of the cave passages remain constant from year to year, in spite of numerous indications that the cave walls and ceilings are continually melting. A balance must, therefore, exist between the amount of heat released and the rate at which the plastic flow of ice tends to close the cave openings. The surface of the subsiding ice fill in the crater also appears to remain relatively constant from year to year, suggesting that snowfall at the summit is more than adequate, even in a light snowfall winter like 1972-73, to fill the craters to a normal snow level. Spring and early summer climbers report snow-filled craters and sealed cave entrances every year (personal communication, Lokey, 1972; and Molenaar, 1973). Ironically, after the below average snowfall of the 1972-73 winter, the cave entrances were more difficult to enter in August, 1973, than in August of 1971 and 1972 following winters of record snowfalls. The number of clear, warm, summer days or other factors are more important than winter snowfall in determining the level of the crater ice surface and when the entrances open.

The surface of the subsiding crater ice varied 12 m (vertically) over a period of 103 years. Van Trump (in Haines, 1962, p. 65-66) noted that the west crater cave where he and Stevens had spent the night in August, 1870 was 9 m shorter in August, 1883 because of a lower ice surface. Stevens (in Haines, 1962, p. 202) observed in August, 1905 that the crater ice level was about 12 m lower than it had been during his previous climb in 1870. The ice level in the west crater is now similar to the conditions described in 1870, with the snow-free northwest edge of the crater rim normally standing about 3 m above the ice surface in August.

Subsidence resulting from subterranean melting accounts for numerous features. Small surface crevasses, seldom large enough for human entry, form at the crater surface due to tension created by relatively rapid ice subsidence over areas underlain by caves. The closed depressions in the crater surfaces described previously occur directly above cave passages and are also subsidence features. Moxham (1970, p. 85) noted the existence of the smaller depression in the west crater on airphotos and suggested that this might be a small volcanic crater. However, this depression is centered directly above the Lake Grotto and must be due to subsidence.

Ablation and Subsidence Rates

Objects on the crater surface directly above the caves move downward with the subsiding ice and, in a relatively short time, appear in the caves below. In the east crater Bird Room, the remains of a Greater Yellowlegs(?) (*Totanus melanoleucus*) were discovered on the cave floor in 1970. A similar bird was discovered on the surface of the crater ice. Identification of species is difficult because both specimens lack heads. In 1974, a *Spatula clypeata* or Shoveler duck was found near the Bird Room (Jack Snavely, personal communication). In the ceiling ice of the same room, a machine-made, red, woolen glove was discovered. Based on subsidence rates determined by ablation measurements, it is estimated that the glove was dropped by a climber crossing the crater 30 to 40 years ago and subsided at a rate of 3 m/year through the 100 m thickness of ice.

In the west crater Lake Grotto, a number of metal rods, cables, and other hardware were found on the floor in 1971, along with a part of *The Chicago American* from July 17, 1959. The materials were probably left at the crater surface by the Project Crater expedition of 1959-60 (Miller, 1970). They moved through the 41-meter-thick ceiling at a minimum rate of 3 meters per year.

Other objects, including bottles and cans, were present on the floors of some of the east crater caves and, most likely, subsided through the ceiling ice. Most of this debris was cleaned up in 1971 by Project Crater personnel.

Although subsidence and cave ceiling melting rates balance on a yearly basis, one may exceed the other during different seasons. In the west crater, bent and cracked columns of ice near the entrance to the Lake Grotto (Fig. 12) indicate more subsidence or less cave ablation at this locality during the summer. Ice columns formed during the late summer in 1971 and 1972 disappeared during the winter, when the cave entrances were sealed by snow. Access by cold winds is eliminated at this time and the consequently warmer(?) cave air melts the speleothems. Only small ice stalactites were present at this locality in 1973.

The upper parts of the entrance passages are deprived of warm air currents during times of snow blockage and ceilings sometimes subside to the floor. Supplies placed in one of the west crater entrances in August of 1971 were found to have been crushed by subsiding ice the following summer.

Because the level of the crater ice surface and the dimensions of the caves below are similar each year, the system must be in dynamic equilibrium and the average ablation rate in the cave (measured in water equivalents) must equal the average rate of accumulation and subsidence above the caves. Holes were drilled in the walls and ceilings and their depth changes noted over a period of three to eleven days. Table 2 shows that ablation rates of 2.0 to 3.5 m per year are common and are consistent with estimates based on rates of subsidence of objects through the ice fill.

Summit Hydrology

The estimated volume of water released in the summit craters every year by ablation can be calculated as follows:

$$Q = [(a + a')\,\bar{s}\ \bar{m}] + r$$

where
- Q = approximate quantity of water released by melting each year.
- a = area of ice exposed in entrance and perimeter passages.
- a' = area of ice exposed in the Bird and Lake Grottos.
- $\bar{s}$ = average specific gravity of ice in the caves.
- $\bar{m}$ = average thickness of ice melted each year.
- r = volume of water released by melting of ice and snow on crater surface.

By substituting computed and estimated values, the final solution is

$$Q = [(124{,}500\text{m}^2 + 5{,}574\text{m}^2)\ (.77\ (3\text{m})] + 47.979\text{m}^3$$

$$Q = 348{,}450\text{m}^3 \text{ of water released each year.}$$

TABLE 2. Cave Wall Ablation Rates

Location	Dates	Elapsed Time	Depth Change	Estimated Yearly Melting
Bird Grotto	8/16-8/27/71	11 days	8.0 cm	2.65 m
Bird Grotto	8/16-8/27/71	11 days	5.0 cm+	1.66 m
Steam Cup	8/16-8/27/71	11 days	5.8 cm	1.92 m
Lake Room	8/12-8/14/73	56 hours	2.2 cm	3.48 m
Lake Room	8/12-8/14/73	56 hours	2.2 cm	3.48 m
Lake Room	8/19-8/27/71	8 days	4.5 cm	2.05 m

The calculated value of Q is a minimum value, because no allowance was made for water added to the hydrological system by the frequent formation of summer hoarfrost or for the cave passages that are undiscovered or unmapped. There may be isolated chambers not joined to the system connecting with the surface and melting in some areas may not be rapid enough to maintain a cavity beneath the subsiding ice. The latter may be particularly true of deeper parts of the east crater, where shear stresses and the resulting tendency for openings to close through plastic flow are very high. Melting by conduction and solar radiation is important during the summer months and, in the calculation of surface ablation, is estimated to be 0.3 m per year. Rainfall does not occur at the summit and is, therefore, not a source of error.

Meltwater generates drips from the cave walls and ceilings and moves through the permeable debris on the cave floor towards the deeper part of the crater. In one area in the east crater, as first reported by Flett (1912), a small subterranean stream descends steeply towards the Bird Grotto from just below Columbia Crest. Whether the east crater water accumulates in a crater lake as does the west crater water is unknown. The size of the closed depression in the east crater and the centering of its deepest part over an area lacking accessible caves strongly suggests the presence of additional caves or areas of rapid ablation where a lake might exist.

A number of unverifiable reports of crater lakes in Indian legends may be credible. A lake warm enough to swim in (Clark, 1953, p. 32) or a "fiery lake" of steaming water (Stevens, 1876, p. 522 and Haines, 1962, p. 42) may have existed before 1870. Saluskin, a Yakima chief, reportedly guided two white men to the base of the mountain about 1855. They told him they had climbed the mountain and that there was "ice all over top, lake in center, and smoke or steam coming out all around like sweat house" (Haines, 1962, p. 17).

Reports of splashing noises made by rocks rolling down the east crater entrance passages in 1912 (Flett, 1912) and 1954 (Molenaar, 1971, p. 186) could not be substantiated from our own observations. We could find no evidence of former shorelines or bedded lake sediments anywhere in the cave system. A lake level 48 m from the surface of the south rim, where the 1954 climbers descended, would produce a lake over 86 m deep covering about half of the east crater. It is unlikely that the ice roof could be maintained under these conditions. A small pocket of water trapped between the ice and the crater floor might explain the splashing noises, although no such pockets occurred from 1970 to 1974.

The volume of water vapor emitted from the cave system seems inadequate to account for the amount of water released by melting. Most of the water must drain through permeable zones in the crater floor and emerge as springs or seeps at lower elevations. The level of the west crater lake has consistently remained within $\pm$ 15 cm of the 1970 level during the three years of observation. Higher levels correspond to long periods of warm weather conditions at the surface and lower levels occur in early summer. The lake bed is sealed with hydrothermally-altered clays, but enough permeability must exist to allow some water to escape.

Firn Caves as Volcanic Activity Indicators

The delicate balance between the caves and geothermal heat release produces a useful indicator of volcanic activity and its potentially serious hazard to human life and property. Occasional monitoring of the caves would detect subtle increases of heat many months before an actual eruption. It is preferable that such inspections be performed by individuals with an intimate knowledge of the cave system, to avoid well-meant "false alarms" describing collapsed cave roofs (Nelson, 1972; Wilkins, 1972) and other phenomena that might indicate heat changes.

Rapid changes in cave-passage size and location could indicate the appearance of new fumaroles, shifts in fumarole location, or significant changes in heat release. Steam cups are particularly useful indicators of fumarole changes, because their appearance or disappearance is more easily detected than are small changes in passage size. No fumarole changes were noted during the study, but a slow enlargement of at least some of the cave passages is occurring. Certain walls in the Bird and Lake grottoes have retreated upslope, slightly increasing the

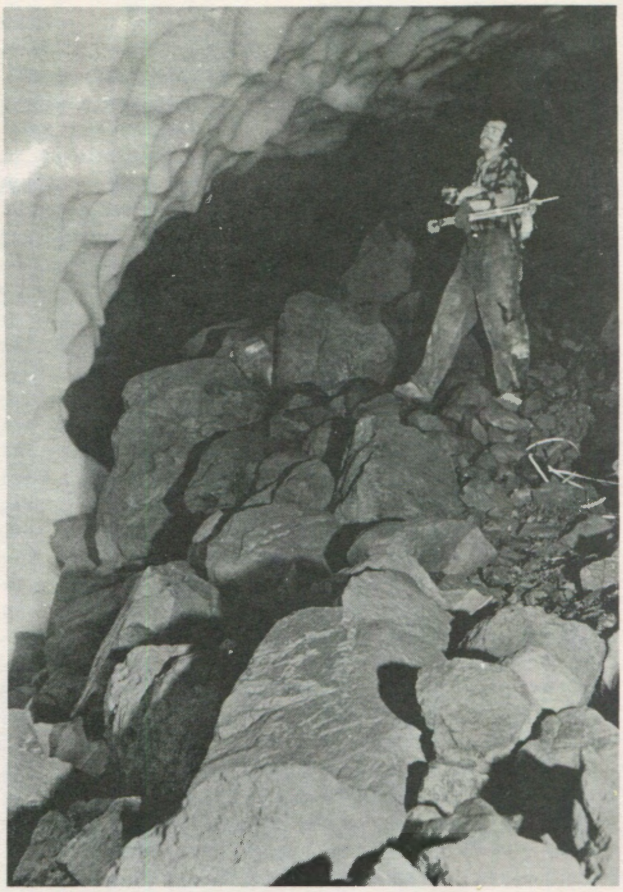

Fig. 13. Moat separating debris ridge and ice wall. (author)

floor areas of the grottoes. This slight enlargement is due to the retention of heat in the grottoes, because of inefficient air circulation and insufficient time for equilibrium to have been achieved. The perimeter passage in the east crater also shows slight changes. Recent closure of one entrance passage (prior to 1970) is suggested by a debris ridge against the ice wall of the perimeter passage where there is no corresponding upslope entrance passage. A few debris ridges that were formerly in contact with the ice wall now have a separating moat as much as one meter wide, indicating widening of the passage (Fig. 13).

The very slow enlargement of the cave passages during the past 4 years suggests that complete equilibrium to recent small increases in heat has not yet been attained. The observed changes are not considered significant at this time, because small increases and decreases in heat release may be a normal condition. Future observations will clarify the meaning of the passage enlargement.

If the volcano were beginning another eruptive phase, the first indications would probably be changes in the firn cave systems, followed or accompanied by an increase in microseismic activity. The passages would be enlarged rapidly and sulphurous gases, along with increased steam, would be emitted. As the buildup continued, the increased heat release would be apparent from increased steam emissions and from changes in crater ice topography visible from the summit or from an airplane. Large steam emissions would also be visible from viewpoints on the flanks or at the base of the mountain.

The firn caves would collapse, as the crater ice melted, and crater lakes would appear in the two craters. The west crater lake would contain a maximum of 1.1 billion liters of water and the east crater lake about 7.4 billion liters. This volume of water perched on top of the unstable mountain would create a potentially serious hazard (Crandell and Mullineaux, 1967). The hydrothermally altered material at the summit would be particularly susceptible to debris flow. The release of large quantities of warm or hot water could cause surging, breakup, or other unusual behavior of the numerous glaciers on the mountain flanks. Earthquake intensity would increase and would be felt by the people nearby. Seismographs, sensitive tiltmeters, and aerial observations would be the most useful means of monitoring volcanic activity at this stage.

CONCLUSIONS

Snow falling into the two craters on top of Mount Rainier changes into firn ice (density 0.4 to 0.81 gm/cm^3) within two years. Maximum ice thickness is 48 m in the west crater and estimated to be 120 m in the east crater. Following a small summit eruption between 1820 and 1854, caves began forming above fumaroles located along ring fractures in the crater floor. They gradually enlarged laterally and towards the surface, producing a relatively horizontal perimeter passage connected to the surface by a number of ascending entrance passages. Total passage length, including the large grottoes, exceeds 2 km. The Lake Grotto in the bottom of the west crater contains the highest crater lake in North America. The deepest accessible point is in the east crater Bird Grotto, 104 m below the crater surface and estimated to be 16 m above the crater bottom.

In glaciers, where denser forms of ice exist, crevasses and other openings close rapidly below depths of 25 to 30 m. The extreme depth of some of the firn cave passages on Mount Rainier (104 m) is possible because of the combination of high heat flow, the presence of firn rather than of glacier ice, and the polar temperature of the ice.

The crater ice is subsiding at rates of 2.0 to 3.5 m per year to replace ice melted in the caves. The sizes and locations of the cave passages are determined by the subsidence rate and the local rate of heat flow. The system is in approximate equilibrium, although slow enlargement of Lake and Bird grottoes in recent years is probably due to a slight increase in heat release.

The delicate balance between subsidence and ablation provides a sensitive indicator of geothermal heat release at the summit and, consequently, an indication of the activity of the volcano. Periodic inspections of the cave system might give considerable advance warning of an impending eruption. Geologically recent volcanic activity, historic accounts of restlessness
ically recent volcanic activity, historic accounts of restlessness,
and present seismic and geothermal activity demand that Mount Rainier be considered an active volcano and that adequate contingency plans be made to protect human life and property in the area.

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The Origin of Maze Caves

Arthur N. Palmer *

ABSTRACT

A maze cave consists of a network or irregular pattern of solution passages containing numerous closed loops of contemporaneous origin. Examination of field data indicates two common settings under which nearly all maze development occurs: (1) where soluble rock receives diffuse groundwater recharge from the overlying surface or through an adjacent formation; and (2) where ground water in a cavernous region undergoes great variations in discharge and in hydraulic head, owing to floodwater recharge. In case 1, water is supplied uniformly to all major fractures within the cavernous zone, so that each one experiences comparable rates of solution. This type of recharge generally occurs in karst aquifers capped by permeable but insoluble rock, or in isolated hills of soluble rock. In case 2, ponding occurs behind constrictions during peak flow in active stream passages, resulting in the rapid development of blind fissures and diversion mazes. In either case, mechanical joint enlargement by tectonic forces or by removal of overburden favors the development of joint-controlled network mazes. Other settings commonly associated with maze caves, such as artesian karst aquifers, appear to have only an indirect influence on maze development.

INTRODUCTION

Among the numerous patterns displayed by caves in soluble rock, two dominant geometries can be recognized: a *branchwork pattern*, which is formed by tubular or canyon-like conduits that intersect as tributaries in the downflow direction, and a *maze pattern*, which consists of a labyrinth of intersecting passages of rather uniform character that form closed loops. Although branching conduits seem well suited to the general interpretation of caves as flow paths for ground water from points of recharge to springs at lower elevations, the origin and hydrologic function of maze caves present an interpretive problem that has received little previous attention.

The purpose of this paper is to assemble as much as possible of the known information related to maze caves and to categorize them into genetic types on the basis of passage patterns and geohydrologic setting. Few pertinent hydrologic and chemical measurements have been made in maze caves, however, so the genetic hypotheses presented here should be considered only tentative models to guide further investigation.

CHARACTERISTICS OF MAZE CAVES

The labyrinthine character of a true maze cave is the result of simultaneous, rather than sequential, enlargement of joints and partings by solution. However intricate, a cave cannot be considered genetically to be a true maze cave if its closed loops are formed by the interconnection of passages representing different stages of development, as in the case where ground water is diverted to progressively lower levels.

A maze cave can generally be classified as one of the following types: (1) A *network maze* consists of an angular grid of intersecting fissures that is formed by solutional widening of nearly all major joints to roughly the same size within a given area of soluble rock. This pattern is commonly restricted to areas of vertical or near-vertical joint orientation. (2) An *anastomotic maze* is typically formed of curvilinear tubes of circular or elliptical cross section that intersect in a random or braided configuration. Passages may form a two-dimensional pattern that is confined to favorable partings or geologic horizons, or a three-dimensional pattern that follows no single geologic structure. Joint control of local passage segments is common, but rarely exerts a dominant influence upon the overall passage pattern. (3) A *spongework maze* consists of interconnected, nontubular solution cavities of varied size and irregular geometry arranged in an apparently random, three-dimensional pattern.

Although most maze caves fall decisively into a single category, continuous gradations can occur among these three extreme types. Some mazes possess characteristics of all three. Typical cave patterns are compared in Fig. 1.

The distinction between branchwork and maze characteristics is often rendered difficult by complicating factors. Owing to limited groundwater recharge, many caves show a rudimentary pattern, perhaps consisting of merely a single passage, in which only the *tendency* toward branching or maze geometry is present. Passages in larger caves may exhibit a complex morphology because of variations in geologic and hydrologic conditions within the drainage basin, changes in hydrology with time, or the superposition of more than one stage of development.

BRANCHING VERSUS MAZE TENDENCIES IN CAVE DEVELOPMENT

On the basis of personal field experience and the speleological literature, it is estimated that at least 75 percent of the known caves are composed of essentially branchwork elements. Therefore, an understanding of the origin of maze caves is perhaps best obtained by examining the conditions that produce a branchwork cave pattern, and by recognizing the circumstances under which this pattern would be suppressed in favor of closed loops.

The hydraulic gradient that determines the flow of water through limestone is much more dependent upon variations in channel size than upon variations in discharge.* Under phreatic conditions, given two adjacent flow paths in limestone (Fig. 2), the one possessing the larger initial fracture width will normally maintain a much lower hydraulic gradient, even though it transmits more flow. If both paths discharge into a nearby river, the

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* For laminar flow, hydraulic gradient is proportional to Q/r^5, where Q = discharge and r = effective channel radius; for turbulent flow, hydraulic gradient is proportional to Q^2/r^5. Variations in r will produce a much greater effect upon the hydraulic gradient than variations of the same percentage in Q.

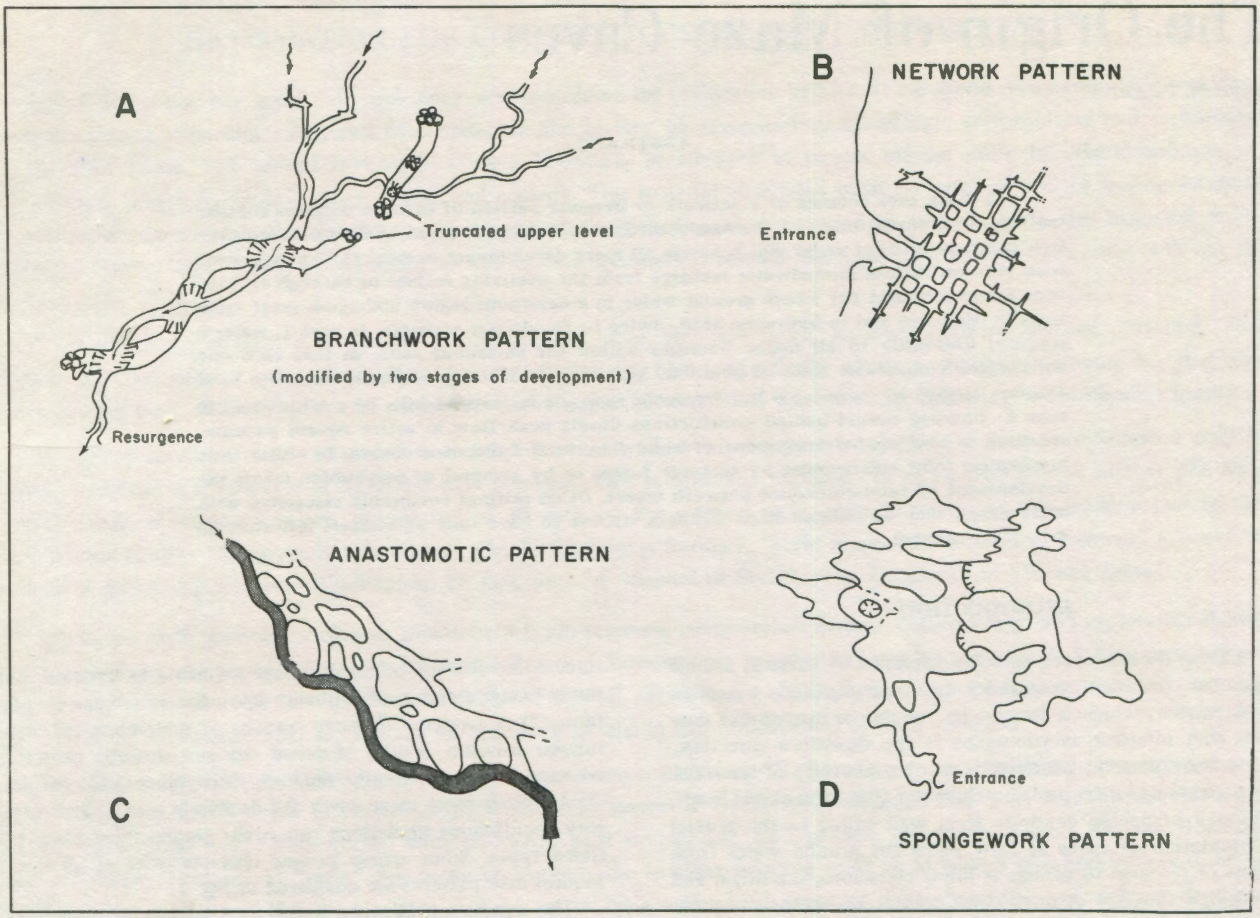

Fig. 1. Representative examples of cave patterns discussed in the text. In contrast to the three maze patterns (B, C, and D), the closed loops in the branchwork cave are the result of sequential stages of development at different levels. Joints influence the passage trend in the branchwork cave, but do not affect the basic passage interrelationships.

wider, more efficient opening will contain water at hydraulic heads only slightly greater than that of the river, while water in the less efficient openings will be required to maintain higher heads. Consequently, the water in nearby inefficient channels possesses a steeper gradient toward the wide channel than toward the more remote river. As flow is directed toward the most efficient path of ground-water flow, a branching pattern of tributaries and master channels tends to develop early in the history of a cave. In addition, with numerous potential flow routes competing for space within the limestone, random intersections are common, allowing different flow paths to join as tributaries to successively larger channels in a branchwork pattern. The latter mechanism is probably significant in perched or vadose systems, where neighboring channels tend to be hydraulically independent. In either case, only the paths of flow possessing the greatest discharge evolve into major passages of a cave system, for slow, diffuse flow in limestone generally approaches saturation over distances that are too short to allow significant passage enlargement (see pertinent discussions by Weyl, 1958, p. 175; Thrailkill, 1968, p. 29-31; Shuster and White, 1971, p. 126-127). The development of branchwork caves has recently been simulated by Ewers (1973) using salt and plaster models.

The origin of maze caves requires a suppression of the branching tendency. Solution by slow-moving ground water under phreatic conditions has been cited most frequently as the responsible agent (e.g., W. M. Davis, 1930, p. 557; Bretz, 1942, p. 720), although if this idea is correct, relict maze caves should be common in all well-fractured limestones that have been subjected to phreatic conditions in the past. Mazes are the exception, however, even in the majority of highly jointed limestones. Furthermore, there are many well-documented examples of single long cave passages, showing no maze geometry, for which a phreatic origin is indicated by bedrock floor and ceiling segments that rise in elevation in the downstream direction (as in Deike, 1967, p. 83; Ford, 1971, p. 85; and Appleton, 1974, p. 35).

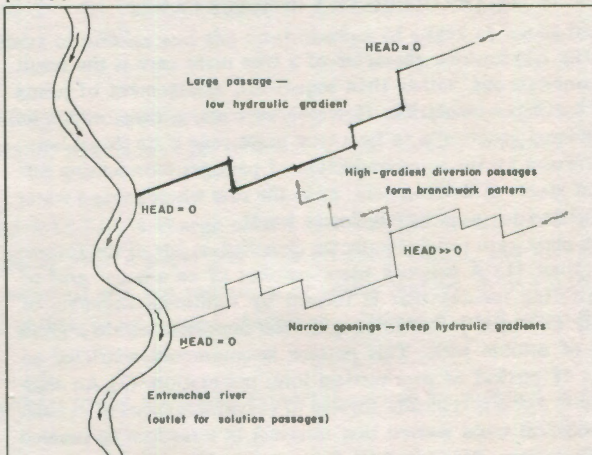

Fig. 2. One of several mechanisms that account for branchwork cave patterns. Water in narrow, inefficient conduits possesses a steeper hydraulic gradient toward nearby large passages than toward more remote spring outlets. Diversion takes place along the most efficient available paths, forming tributaries to the larger conduits. For a maze cave to form, this tendency must be suppressed.

Curl (1971 and 1974) has demonstrated analytically that where a cave passage branches and reconnects over a short distance (in which there is no significant change in concentration of dissolved solute), the alternate flow paths should be competitive in their solutional growth under either laminar or turbulent flow. It follows that the *maze* geometry is the norm, and that unitary conduits represent the unexplained case. Curl recognizes that certain factors not yet considered in his analysis must subdue maze development, though, to account for the preponderance of branchwork caves.

In an attempt to resolve the conflicting ideas on maze origin, this paper approaches the problem from the standpoint of available field data. By grouping the known maze caves according to their physical setting and pattern, it is apparent that nearly all can be attributed to either of two situations that tend to suppress the common branching tendency: (1) where diffuse, solutionally aggressive recharge takes place uniformly into all available fractures in a soluble rock unit, entering from an adjacent insoluble formation or from the overlying land surface; and (2) where floodwater recharge causes the temporal variations in discharge and head in a growing cave system to be so great that no fixed passage configuration is allowed to stabilize with respect to the flow, a situation that is found mainly in caves fed by sinking streams. It will be shown in the following sections that these conditions can be fulfilled by a variety of common geologic settings.

FIELD DATA

Information on approximately 260 maze caves was gathered from field work and from the available literature, mainly representing North America and Europe. Fifty-four were visited personally during the course of research for this paper. The sample was limited to multiple-passage solution caves that consist predominently of closed loops of synchronous origin, or that display an irregular pattern that could be shown objectively to have a non-branching tendency. Maze sections within otherwise predominantly branchwork caves were included. No attempt was made to assign a specific origin to every cave in the sample, but, rather, to group them according to setting and pattern, so that broad trends in cave origin could be obtained. Consequently, only a few typical examples are identified and described in detail in this paper. Nearly all of the remaining caves in the sample are described in the cited publications.

The location and pattern of each cave was studied in relation to its geologic setting, past and present hydrology, and topographic associations. The 159 maze caves in the United States for which this information was obtained have the physiographic distribution shown in Table 1. Although this list is by no means exhaustive, and is disproportionately weighted according to personal field experience and availability of descriptive literature, it is immediately apparent that maze caves exist in a wide variaty of geologic, geomorphic, and climatic settings. Rock types

Table 1. Physiographic distribution of maze caves in the United States that were studied for this paper.

Geomorphic Province	Number of Maze Caves Studied	Most Typical Geometric Types	Representative Example
New England and Adirondacks	4	Anastomotic	Radium Springs Cave, Mass.
Coastal Plain	1	Network	Warren's Cave, Fla.
Valley and Ridge Province	49	Network, Spongework	Breathing Cave, Va.
Appalachian Plateaus	50	Network	Bear Cave, Pa.
Interior Low Plateaus	15	Anastomotic, Network	Parts of Blue Spring Cave, Ind.
Glaciated Central Lowlands (Minnesota to New York)	16	Network	Glen Park Labyrinth, N.Y.
Ozark and Ouachita regions	2	Network, Anastomotic	Duncan Field Cave Okla.
Great Plains	2	Network, Anastomotic	Parts of Powell Cave, Tex.
Rocky Mts. (including Black Hills)	6	Network, Anastomotic	Jewel Cave, S.D.
Basin and Range Province (including Guadalupe Mts.)	8	Spongework	Dry Pot, N.M.
Sierra Nevada and Coast Ranges	6	Anastomotic, Network	White Chief Cave, Calif.

in the total sample encompass nearly every kind of soluble rock, from Precambrian marbles, dense limestones and dolostones, to poorly indurated Tertiary limestones, as well as gypsum. Although roughly 75 percent are situated in rocks dipping less than ten degrees, examples occur in strata of every attitude from essentially horizontal to vertical. Despite such an extensive range of settings, however, maze caves can be easily categorized into several genetic groups that are both predictable and consistent where the local details of geology, topography, and hydrology are known.

CASE I: MAZES FORMED BY DIFFUSE INFILTRATION

In the first case to be considered, diffuse groundwater recharge is supplied uniformly to all major joints or partings within a local area of soluble rock. Each prominent opening in the soluble rock receives comparable amounts of water, regardless of its initial size, so that the typical result is a tight network of intersecting solution fissures. In contrast, many of the normal karst processes such as sinkhole development tend to concentrate recharge at comparatively few points, and for this reason network caves are not commonly found beneath well-developed karst surfaces. Field evidence is given in the following sections for two distinct types of maze development by diffuse recharge: (a) beneath a thin cap of permeable, insoluble rock; and (b) within isolated hills of soluble rock.

Maze Development beneath a Permeable, Insoluble Cap-Rock

A potential major source of groundwater recharge to a cavernous formation is the land surface directly overlying it. The nature of any intervening cap-rock is, therefore, significant to the cave origin, either as a transmitter of infiltrating water or as a hydrologic barrier.

The overlying rock type was determined for 155 maze caves in the total sample. Fifty-nine of these are overlain by insoluble strata and, of this group, all but two exhibit a joint-controlled network pattern. Caves of this type have been interpreted by previous workers to be the result of artesian flow confined beneath the insoluble rock. However, 86 percent of the capped network mazes are closely overlain by sandstone, which is generally recognized by well drillers as the most uniformly permeable and reliable of consolidated sedimentary aquifers. It is reasonable to assume that most sandstones have greater potential as a path for recharge to the caves than as a source of confinement. If this is the case, the sandstone should act as a governor to the flow, supplying rather uniform amounts of water to each major joint in the underlying soluble rock and promoting the development of network caves.

Field information supports this hypothesis. Typical evidence includes the following:

a. The thin, prominently jointed limestones of the Chester Series (Upper Mississippian), particularly those of Indiana and Illinois, contain numerous caves of basically tubular geometry with local network zones. Nearly all the caves are located in ridges overlain by thick clastic units, most commonly sandstone grading upward to shale. However, network development is largely restricted to the peripheral areas of ridges, where the overlying cap-rock is thinnest and composed only of permeable sandstone. Maps of several such caves are shown in Powell (1961) and Bretz and Harris (1961). The fact that recharge takes place through the exposed sandstone into the underlying limestone has been noted by Powell (1966) and by McGrain and Bandy (1954).

b. Several of the largest maze caves of the eastern United States, such as Clarks Cave and Crossroads Cave in Virginia (Douglas, 1964, p. 150-155), are located in the Helderberg limestones of Devonian age where the cavernous strata dip less than ten degrees and are capped by thin sandstone. This limestone group can be traced northward through the folded Appalachians of Pennsylvania to New York State, where it is again exposed over large areas in a gently dipping attitude, but overlain by rather impermeable shales. Even though the northern exposures are more prominently jointed, network caves are few. Those that do occur are irregular in pattern and show conspicuous evidence for a floodwater origin, as will be illustrated in a later section.

c. Most sandstone-capped mazes in the sample are concentrated near the top of the exposed limestone. The cavernous zone in many cases is much thinner than the exposed thickness of limestone above base level. Most caves of this type have numerous upward extensions to the sandstone contact and receive vertical seepage during wet periods.

Figure 3 illustrates the probable history of development for a maze cave beneath a permeable cap-rock, such as sandstone. (Not all caves overlain by permeable rocks are mazes, however, nor do all network caves overlain by such rocks necessarily share this origin.) Initially, the permeability of the sandstone is greater than that of the limestone (Stage 1) and the maximum hydraulic gradients are lateral, toward the nearest surface outlet. Consequently the majority of groundwater flow takes place within the sandstone. The small amount of water that enters the limestone is solutionally aggressive, however, and the resulting joint enlargement eventually causes the permeability of the limestone to exceed that of the sandstone (Stage 2). The limestone becomes an efficient conductor of water toward the surface outlet, so that much water is attracted downward from the overlying sandstone, enlarging the joints into a network cave cave oriented sub-parallel to the sandstone-limestone contact. Continued erosional dissection ultimately causes the cave to drain (Stage 3), although vadose seepage is still contributed through the sandstone, and backflooding may occur from nearby surface streams. Cave development apparently takes place under a continuous spectrum of groundwater conditions, from deep phreatic to vadose, although the relative importance of the various stages may differ according to the geologic setting and erosional history. The local dip of the strata is not of primary significance, except that a low dip is more likely to provide the conditions favoring uniform recharge to the cave passages.

The flow mechanics during stages 1 and 2 are rather complex and are merely summarized here. The recharge rate to each joint in the soluble rock can be expressed as follows:

$$q \propto \frac{-K\,(\Delta h)}{ln\left(\frac{w}{T}\right)}$$

where q = discharge per unit length of fissure along the sandstone contact, K = hydraulic conductivity of the sandstone, Δh = head difference between the water table in the sandstone and the fissure opening in the underlying soluble rock, w = fissure width in the soluble rock, and T = height of water table above the soluble rock. The quantity $ln\left(\frac{w}{T}\right)$ is negative, approaching zero with increasing values of w. The expression for q becomes invalid as w approaches the magnitude of T. The width of the joint opening is the primary factor in determining head within the joint and, therefore, in controlling flow rate. The *direct* influence of joint width is realtively minor, however: a ten-fold increase in width is necessary merely to double the discharge, if all other factors are held constant. (The logarithmic relationship is caused by the convergence of flow in the sandstone as it approaches each joint, a hydraulic inefficiency that is not relieved significantly by an increase in the joint width.)

During Stage 1, when the transmissivity of the fractures in the soluble rock is very low, an increase in joint width by solution or by expansion due to erosional unloading will result in a great increase in flow rate. The wider openings receive disproportionately large amounts of recharge and therefore are enlarged faster by solution. However, the head within the larger openings eventually approaches a minimum value, in which there is virtually zero hydraulic gradient through the soluble rock. Further enlargement of the openings has a negligible effect on the head, and so the recharge rate to these openings becomes rather insensitive to width increase. The smaller joints, meanwhile, are not robbed of significant amounts of flow, and continue to be enlarged until they, too, contain minimal hydraulic heads and receive essentially the same amount of recharge as those joints that were initially larger. Stage 2 is achieved when a sufficient number of joints have been enlarged so that the effective permeability of the soluble rock exceeds that of the sandstone. The total discharge through the soluble rock soon becomes limited by the availability of infiltration from the overlying land surface, and the water table in the sandstone declines in elevation.

Given a sandstone of typical permeability, containing a water table no more than a few tens of meters above the resurgence, and whose basal contact with the soluble rock is a few tens of meters below the resurgence, the transition from Stage 1 to Stage 2 occurs when the major joints have enlarged to widths on the order of one millimeter. On the other hand, joints less than about 0.01 mm in width permit such small flow rates that they probably never enlarge to competitive size. The importance of initial fracture width is evident, and will be discussed at length in a later section.

Many networks occur beneath insoluble rock of low primary porosity that contain prominent joints vertically continuous with those in the cavernous rock. In this case the soil may assume the role of groundwater collector, distributing water to the joints in the caprock, which transmit it downward to the soluble unit. Areas in which soil directly overlies soluble rock do not favor maze development, as unconsolidated material is easily sapped into the solutional fissures, resulting in sinkholes that disrupt the uniformity of recharge.

The geologic setting described in this section is similar to the conceptual "sandwich aquifer" of White (1969), in which network cave development takes place in a thin carbonate unit both overlain and underlain by insoluble strata and where backflooding occurs from a nearby river. This mode of cave origin is more closely related to the floodwater processes described in the following section than it is to diffuse infiltration, for the backflooding is not essential to the origin of networks beneath a permeable cap-rock. Backflooding may contribute a great deal to cave enlargement, however. Considered individually, the small number of network caves capped by impermeable rock, such as thick shale, can nearly all be attributed to this type of recharge.

The variety of setting and morphology that occurs within sandstone-capped maze caves is shown in the following examples:

Clarks Cave, Virginia, is described here in detail as a typical sandstone-capped maze (Fig. 4). In addition, it illustrates many of the ambiguities that plague the interpretation of this type of cave. It is an extensive network of fissure passages in the New Scotland Formation of Devonian age, overlain by 10 to 25 m of prominently jointed sandstone. Its six entrances are located in a steep bluff along the Cowpasture River. The cave is restricted to the upper 15 m of the limestone and is roughly concordant to the dip, which is locally 2.5° to the south-southeast, away from the river. Passages are ungraded, with poorly accordant junctions and highly irregular floors and ceilings. Much of the cave is presently dry, but seepage enters through the sandstone in some areas, especially during the spring season. Detrital sediment is mostly fine grained and shows no evidence of significant free-surface stream flow. Water that enters the cave quickly disappears into sediments and inaccessible fissures. The major passages extend upward to the sandstone in numerous places (Fig. 5), although the contact is generally obscured by secondary mineral growth and by exposure in inaccessibly high or narrow passages.

The proximity of the river invites speculation that it may have been the main source of recharge to the cave. However,

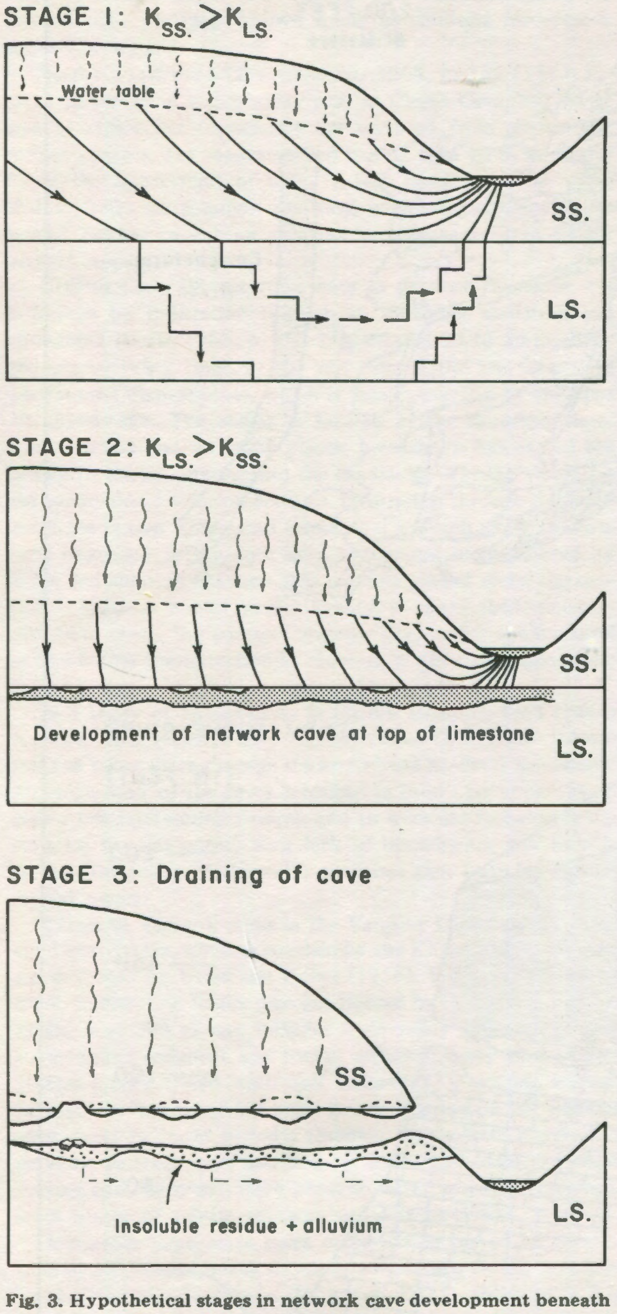

Fig. 3. Hypothetical stages in network cave development beneath a permeable, insoluble cap-rock (in this example, sandstone overlying limestone). In the limestone, the increase in permeability (K) due to solution causes increasing amounts of water to flow through the limestone. The governing effect of the sandstone on the rate of recharge allows all major joints in the limestone to become enlarged at comparable rates.

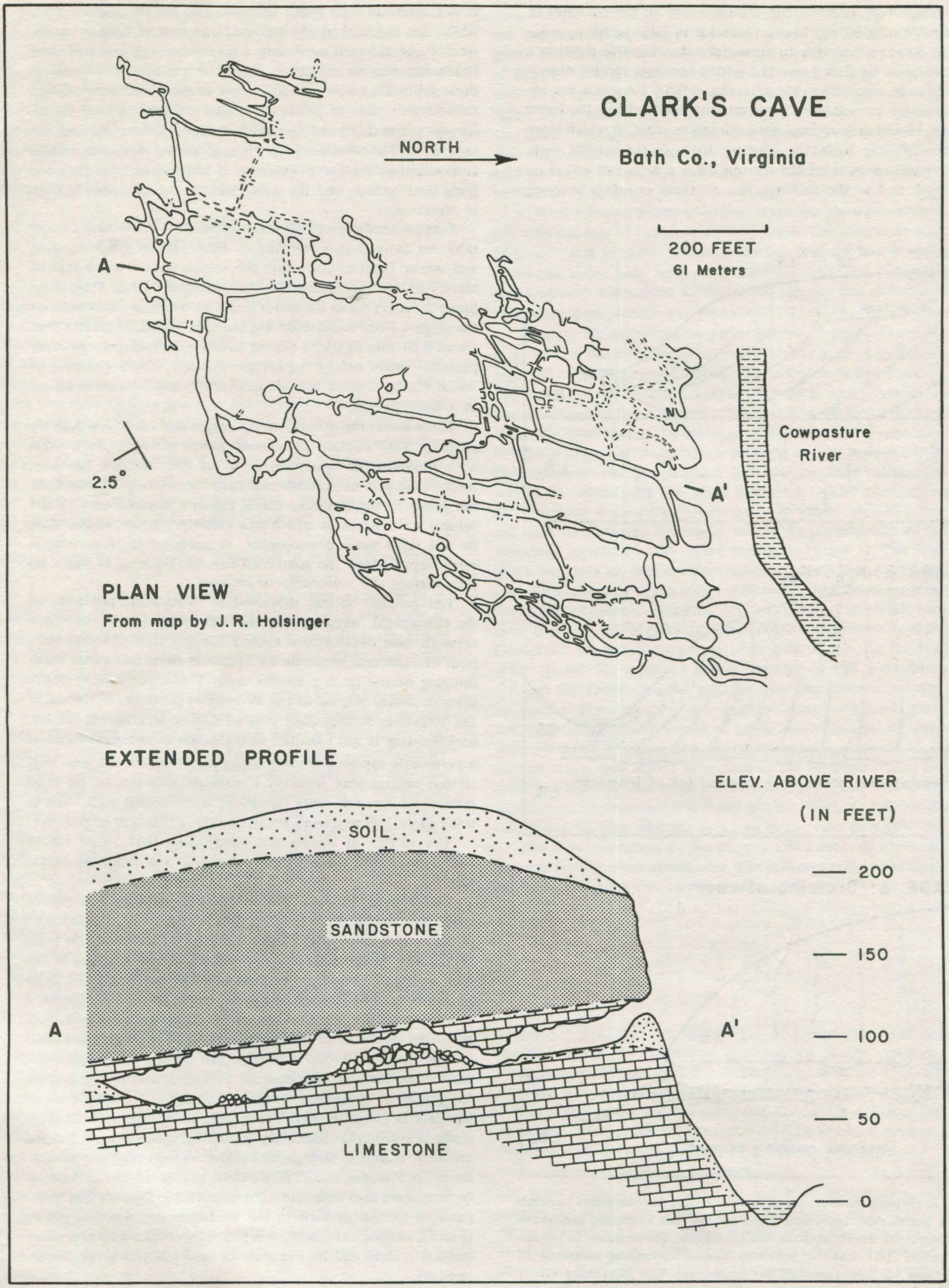

Fig. 4. Map and profile of Clarks Cave, Virginia. Map of cave from Holsinger (1961); geologic profile by A. Palmer, M. Palmer, J. Allen, and P. Bongei, April, 1969.

comparison with caves known to have formed in this manner, such as Glen Park Labyrinth in New York (described later), indicates that flooding from the river must have been rather static, and not the through flow typical of a subterranean meander cut-off. In particular, the walls show no evidence of current scallops, which are the dominant solutional forms in this type of maze. Backflooding must have contributed to cave development, but is not a necessary condition for its network pattern, as certain other nearby mazes in the same geological setting are not near surface streams. It is also unlikely that the cave was formed by artesian flow confined beneath the sandstone cap-rock, for there is little evidence that the passages once continued far beyond their presently known limits as conduits for remote sources of water.

Nearby Crossroads Cave (Douglas, 1964, p. 152-155) is located in the same stratigraphic unit as Clarks Cave, but in an area of rather flat topography far removed from entrenched surface streams. The sandstone cap is only 5 to 10 m thick and forms the upper limit of many fissure passages in the cave. Vadose flutes are common and most intersections of joint-controlled fissures have been enlarged to rudimentary shafts by vertical seepage.

Griffith Cave, Illinois, represents an extreme of vadose enlargement by infiltration through an overlying sandstone cap (Bretz and Harris, 1961, p. 16). It is located 25 to 30 m above the nearby valley floor, at the very top of the Ste. Geneviève Limestone (Mississippian), which is locally overlain by the Spar Mt. Sandstone. The ceiling is located at the sandstone-limestone contact and, in many places, breakdown has caused the ceiling to retreat upward into the sandstone. Vertical flutes are the predominant solution features. Infiltration through the sandstone has caused widespread interstratal solution along the contact, expressed in the cave as a recessional niche floored by lapies and solution channels that extends inward along the contact as much as several meters. Typical of many sandstone-capped maze caves, the passages are now dry, raising the question as to whether the reduction in recharge is due to changes in the local physiography or to variations in climate.

In a study of fissure caves in eastern Missouri, Brod (1964) suggested that recharge into the cavernous Ordovician carbonates has taken place through the *underlying* St. Peter Sandstone. Although most of the caves described in Brod's paper consist of only a few intersecting fissures, and so were not included in the sampling for this paper, their lack of branchwork tendency is perhaps the result of the same processes that form sandstone-capped mazes.

Extensive network caves in the Vanport Limestone of western Pennsylvania, which is overlain by the Kittanning Sandstone, are described by White and Fisher (1958). The caves consist of small, intersecting fissure passages floored by clay and extending as much as 300 m into hillsides from valley walls. No coarse stream-borne sediment was found, although many of the caves possess locally fluted walls and stalactites, indicating seepage through the overlying cap-rock. A clay coating on the walls of some caves indicates periodic flooding by stagnant water. The network patterns were ascribed by White and Fisher to slow-moving and nearly saturated phreatic water, although the dominant source of solvent recharge was not specified.

Numerous large maze caves occur at the top of the Gasper Limestone (Mississippian) in Alabama and are closely overlain by the Hartselle Sandstone (Varnedoe, 1973). Included in this group is well-known Anvil Cave, which has an aggregate length of nearly 20 km. Varnedoe (1964) attributes their origin to artesian flow beneath the sandstone cap-rock. However, it is probable that the sandstone has acted to some extent as a path for recharge, rather than as a confining unit. Elsewhere in the same formation, where the cap-rock includes considerably less-permeable shale, maze caves are generally absent.

Fig. 5. Typical passage in Clarks Cave, Virginia. The contact with the overlying sandstone is at ceiling level (not visible here).

Maze Development Beneath Isolated Hills of Limestone

Roughly 20 percent of the maze caves in the total sample are located within isolated hills of resistant but uncapped limestone. Otherwise they share no common geologic setting or pattern, and in general their hydrologic associations are obscure. Many caves in this group consist of a tight network of fissures similar to those beneath a permeable cap-rock, but which occur within a great variety of rock types and structural settings, in beds that range in attitude from horizontal to vertical. An example is shown in Fig. 6. Equally typical are irregular spongework caves having a plan-view pattern resembling an ink blot. This latter group includes some of the world's most celebrated tourist caves, such as Luray and several others in the Valley of Virginia (Douglas, 1964), the St. Michael's Caves in Gibraltar (Shaw, 1956), Katerloch in Austria, Lewis and Clark Caverns in Montana, and Carlsbad Caverns in New Mexico (Fig. 7). All are characterized by an irregular, non-branching pattern, similar to example D in Fig. 1, and contain unusually profuse travertine. Intersections with the land surface appear to be random and are generally not significant points of recharge or resurgence. Floors and ceilings are ungraded and irregular, although in many such caves the rooms and galleries are concentrated at crude levels. Sediment is sparse and almost entirely fine grained. Solution pockets and spongework occur in the walls and ceilings, but scallops and other indicators of water flow are typically absent, except very locally.

Despite the heterogeneous nature of these caves and the general lack of information regarding their past and present

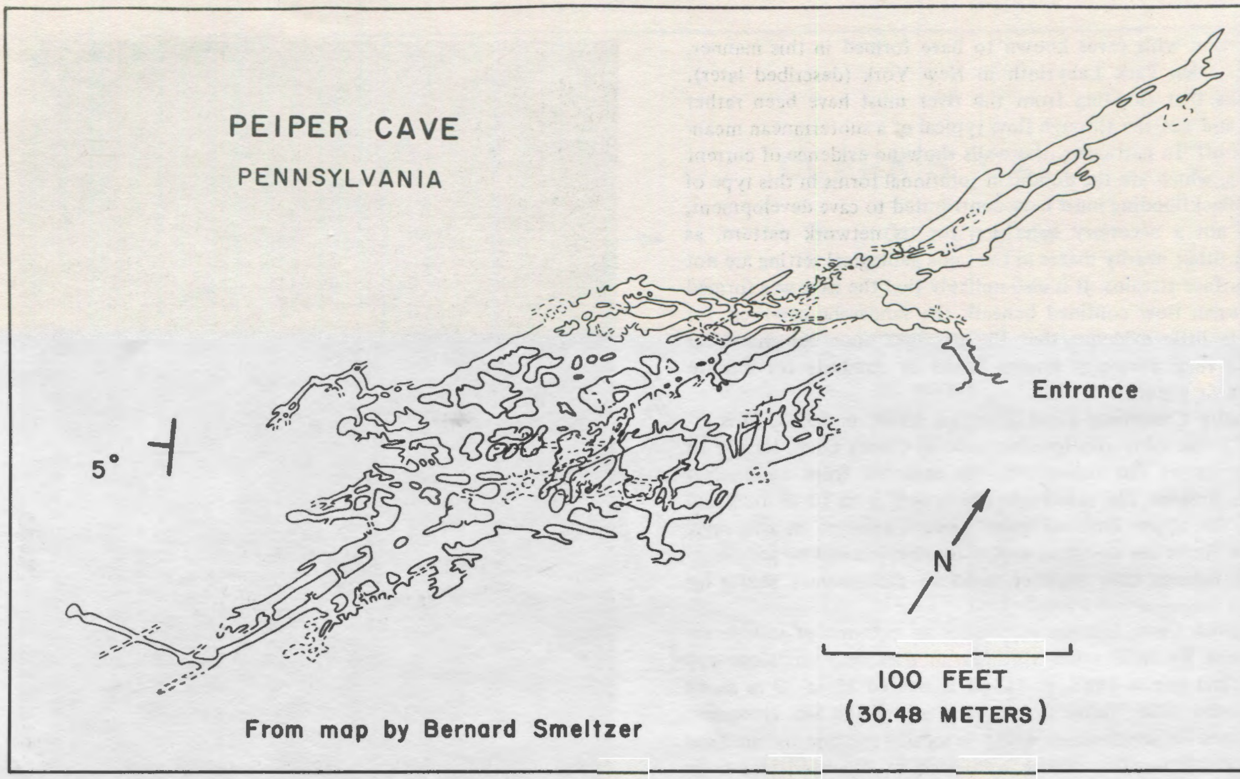

Fig. 6. Peiper Cave, Cumberland County, Pennsylvania. The cave is located in a small knob of limestone and exhibits the irregular, laterally restricted pattern that is characteristic of caves in this type of setting (Map of cave simplified from Smeltzer, 1958; also see Stone, 1953, p. 85-86).

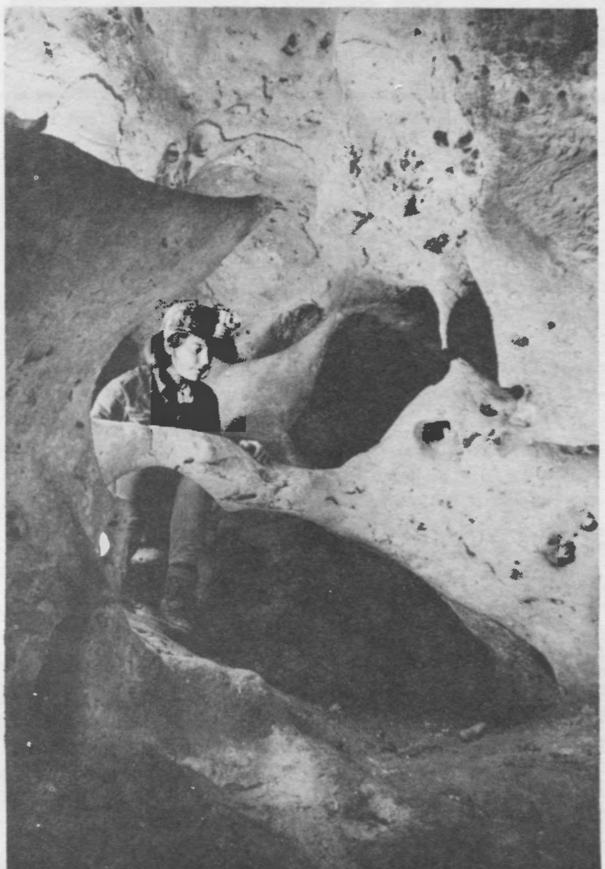

Fig. 7. Spongework in the "Boneyard" of Carlsbad Caverns, New Mexico.

hydrology, the association with isolated hills is so consistent that their maze pattern must be related either directly or indirectly to the local topography. The large volume of dripstone and flowstone in many indicates that diffuse seepage through the overlying land surface, though saturated with respect to calcium carbonate, can be an important source of groundwater recharge.

Maze caves of this type appear to be restricted to hills that have remained topographically higher than their surroundings throughout the development of the present landscape, owing to resistance to weathering, but are rare in erosional outliers of dissected plateaus. Among the most common examples are reefs, ridges of inclined limestone overlain and underlain by shale, and monadnocks located along drainage divides (Fig. 8).

The convex topographic surface of a knob or ridge is unfavorable to the development of concentrated recharge through sinkholes, so that infiltration from the surface remains comparatively diffuse. If infiltrating water is to be effective in developing maze caves, it is necessary to allow the water to retain or to regain its aggressiveness. A possible mechanism could be the process of Mischungskorrosion (Bögli, 1963), which could account for solution in the phreatic zone by the mixing of saturated groundwater recharge with other saturated waters of different chemical content (e.g., where the two solutions possess different CO_2 partial pressures). As the water table drops, calcium carbonate would be precipitated in those parts of the cave ily remain diffuse, with rather static groundwater conditions favoring the development of spongework and network caves.

Evidence to support this hypothesis is difficult to obtain, because most caves in such a topographic setting are too small to allow a valid interpretation of their pattern. However, nearly all the large caves found in isolated hills of limestone have network or spongework patterns and are not truncated segments of tubular conduits. An example can be given for Pennsylvania, a state that is known for small, joint-controlled caves. Of the 272 caves listed for that state by Stone (1953), 141 are described in sufficient detail to distinguish whether they are of tubular or maze geometry. Surprisingly, only 23 percent of this group contained maze sections (as defined in the introduction to this paper), whereas the remainder consisted of basically tubular passages, either as unitary conduits or in branchwork patterns. Twenty-two caves are described as being located above the base, or at the top, of hills or ridges. Nine of these could be assigned a meaningful geometry; most of the remainder consist of isolated fissures. *All nine* are well-developed network caves. The assembling of such figures is rather subjective, but the maze geometry of caves in isolated hills of limestone can be demonstrated throughout the world (for example, see Jennings, 1971, p. 145).

Flooding from nearby surface streams is a factor in the development of many caves in isolated hills that are bordered by rather flat, impermeable areas. As will be shown in the following pages, this type of recharge is also effective in producing maze caves. Peiper Cave, Pennsylvania, is a typical example that has received both diffuse seepage and periodic flood recharge (Fig. 6). As in most caves of this type, the relative importance of the two sources is not easily determined. This situation is also common in tropical karst towers and mogotes bordered by alluvial plains, where aggressive surface water enters at the base of the hills and forms recessional niches and irregular caves (Miotke, 1973; Sunartadirdja and Lehmann, 1960). Spongework patterns are common among the caves found within this setting (see Wilford, 1964).

Perhaps the most enigmatic of karst features, maze caves formed within isolated hills of soluble rock offer a considerable challenge to future geomorphic and geochemical field research.

CASE II: MAZE CAVES FORMED BY FLOODWATER RECHARGE

Nearly all of the remaining maze caves in the sample for this paper, representing approximately 40% of the total, are characterized by periodic flooding in response to intense rainfall or snowmelt, or show evidence that flooding has been significant in the past. Their recharge is rapid and direct, causing sudden short-term increases in groundwater flow that flood all or part of a cave to the ceiling at least several times yearly. A sinking stream is the most common source of floodwater recharge, accumulating its water at the surface and generally contributing it to a cave at a single point. Such conditions are prevalent in regions of diverse geology, where limestone is exposed adjacent to large catchment areas on non-carbonate rocks or sediments. Flooding is also common in caves fed by infiltration from large areas of barren karst, in which the retention of moisture by soil is negligible.

Caves that are developed under these conditions possess many features commonly associated with a "phreatic" origin, including local maze patterns, but instead have formed mainly above the low-flow top of the phreatic zone (Palmer, 1971). Because of the great variation in groundwater discharge, no single cave passage can adjust to all flow conditions. Local breakdown, insoluble beds, or clastic fill deposits form constrictions in the active stream passages that are generally able to transmit low flow, but transmit peak flows so inefficiently that water is easily impounded behind them, flooding parts of the cave under great hydrostatic pressure. Because in closed-conduit turbulent flow the hydraulic gradient is inversely proportional to the fifth power of the passage radius, even minor constrictions can produce head losses several orders of magnitude greater than those in the larger passage segments. As the discharge increases, the pressure difference across each narrow section rises geometrically. Upstream from the constriction, fractures and partings in the limestone are subjected to extremely steep hydraulic gradients, allowing rapid development of blind tubes, spongework, and ceiling pockets, as well as maze-like diversion passages around the constriction. The dynamic and variable flow conditions perpetuate the tendency for severe flooding by promoting collapse and localized sediment accumulation, which restrict the flow during periods of high discharge. In contrast, under long-term phreatic conditions, even though static pressures may be great, the head differences, aggressiveness, and turbulence are minimal in comparison with floodwater conditions. Most diversion passages formed by flooding are ungraded tubes or fissures containing numerous closed loops. Scallops are prevalent in the areas of high velocity. Sediment is patchy and of irregular thickness, and varies in grain size from clay to large cobbles according to local flow conditions. Floodwater features are generally superimposed on a pre-existing cave pattern because the origin of these features demands the presence of passages that are large enough to deliver turbulent, acidic water to the site of a constriction. The importance of fluctuating ground-water levels to the origin of caves has recently been stressed by Powell (1970).

Fig. 8. The Sattelberg, Austria, a hill of Silurian reef limestone in which the well-known tourist cave Katerloch is developed. The hill rises to an altitude of 1140 m and the entrance to the cave (near the skyline in left center) is at 900 m. The cave has a spongework pattern, is heavily decorated with travertine, and extends over a vertical range of 200 m.

The morphology of floodwater mazes is highly varied. Strong bedding-plane control of cave passages generally yields an anastomotic maze, a geometry that is rare in other types of maze caves. Prominent jointing favors the development of networks, though most are rudimentary and of more irregular pattern than those formed by diffuse recharge. Small spongework mazes are formed in massive rocks possessing few secondary openings.

There appears to be little relationship between anastomotic floodwater mazes and the smaller scale bedding-plane anastomoses, which possess a denser pattern than the maze geometry, are not scalloped, and generally decrease in size away from the passage with which they are associated. Although some may have been formed by periodic inundation of bedding-plane partings by flooding of a pre-existing cave, there is strong evidence that most anastomeses are primitive flow routes that define the pattern of subsequent cave passages (Ewers, 1966).

The following paragraphs illustrate the considerable variety of maze development that can occur under floodwater conditions.

Diversion Caused by Breakdown: Blue Spring Cave, Indiana

Few examples of cave morphology show as clear a relationship between cause and effect as the Maze in Blue Spring Cave

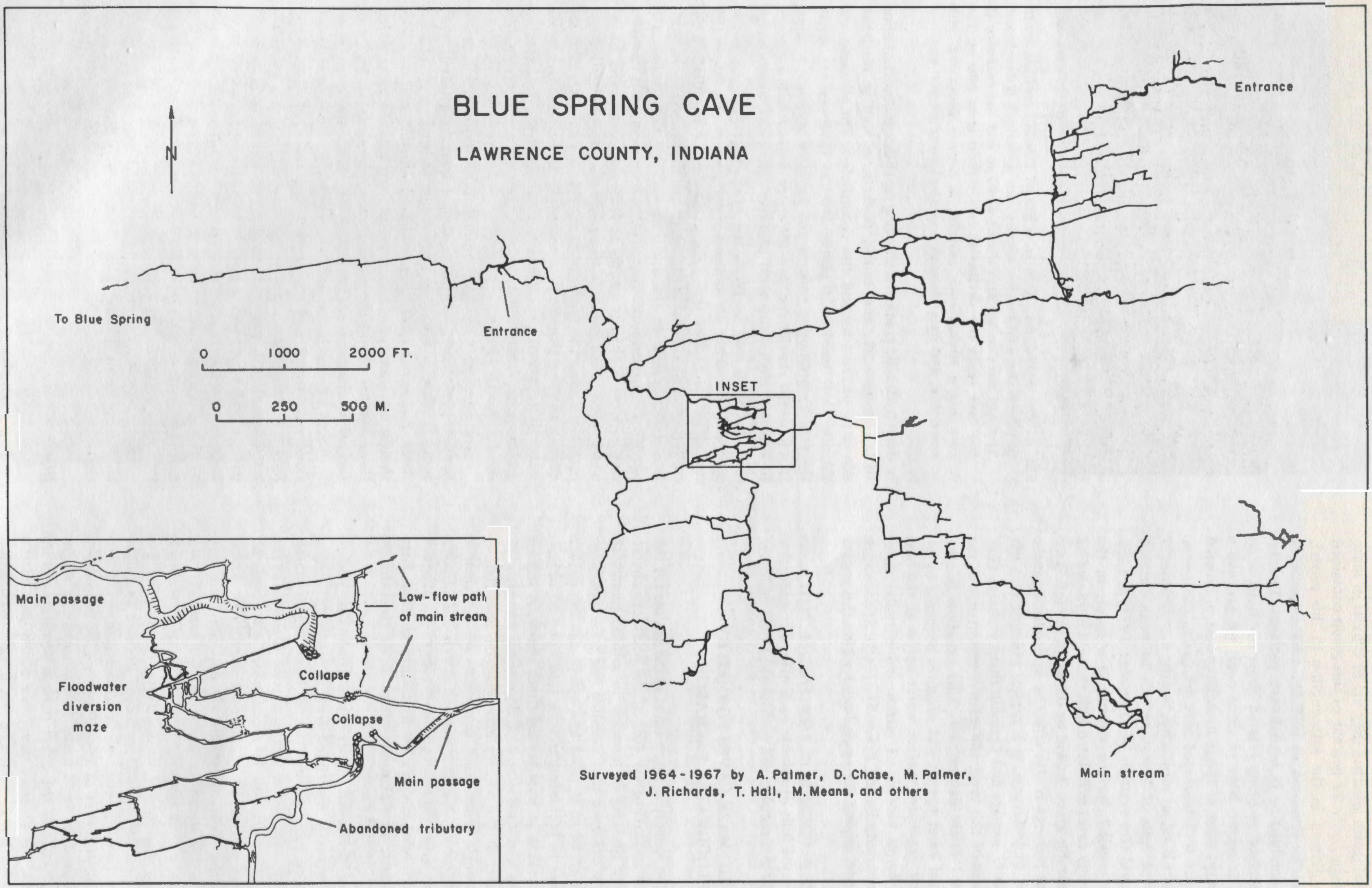

Fig. 9. Map of Blue Spring Cave, Indiana, showing diversion maze around collapse in the main stream passage. Large closed loops in other parts of the cave are formed by sequential diversion to successively lower levels and are not genetically related to mazes. The cave is developed in the Salem and basal St. Louis limestones of Mississippian age (Map from Palmer, 1969).

(Fig. 9). Extensive collapse has occurred in this branchwork cave where a major tributary enters the main stream passage. In response, stream diversion has formed an irregular network that bypasses the collapse and re-enters the main passage at several downstream points. The cave drains more than 10 square km of sinkhole plain upstream from the collapse, with a discharge that varies roughly from 0.02 to 6 cubic meters per second. Although the unobstructed passages in the cave are nearly all capable of transmitting the normal annual peak flow, the collapse material, consisting of breakdown blocks in a dense matrix of gravel and clay (derived mainly from the surface), has such a low permeability that even the low flow has been forced to follow a diversion route. Continued breakdown and solution have created a large compound sinkhole in the overlying surface.

The Maze consists of more than 2 km of fissure passages in the massive Salem Limestone of Mississippian age. Despite its large number of passages, it is still subject to partial filling by periodic high-velocity floodwaters. Its recent development and non-phreatic origin are shown by the predominance of diversion passages at the same level as the present main stream, which occupies an entrenched vadose canyon 5 to 7 m deep in the main passage. Fissures in the Maze extend as many as 10 m both above and below the main stream level. Many of these are more than half filled with water; numerous unexplored leads are entirely water filled. Scallops, natural spans of limestone, blind tubes, and ceiling pockets are common (Fig. 10).

The branchwork passages of the cave possess rather uniform gradients that are strongly concordant with the gentle dip of the limestone. They undergo only small and gradual variations in cross-sectional area throughout their length. Although the prominent jointing in the Salem Limestone imposes a distinct rectilinear pattern on these passages, few exhibit a true fissure geometry. In contrast, more than 90% of the Maze consists of ungraded, joint-controlled fissures containing numerous abrupt changes in cross-sectional area.

Floodwater Development Influenced by Clastic Fill: Big Brush Creek Cave, Utah

Big Brush Creek Cave (Fig. 11) has formed by the subsurface diversion of a large stream flowing off the southern flank of the Uinta Mountains. It drains more than 60 sq. km of predominantly insoluble rocks, mainly Precambrian quartzites and Tertiary conglomerates. The cave roughly follows the contact between the Deseret Limestone and the overlying Humbug Formation (both of Mississippian age), which dips south at 10 to 15 degrees. The deseret consists of interbedded limestones and dolomites, whereas the Humbug is a massive, siliceous limestone breccia. Dye tests indicate that water in the cave resurges more than 8 km from the entrance, at a spring in the thick clastic rocks that overlie the limestone (Green, 1961). Flow into the cave entrance is periodic now, as the sinking stream has recently been diverted to another underground route at a point several hundred meters upstream. The flow is also regulated by a reservoir and diversion ditch upstream from the sink point. However, the discharge during severe floods is still great enough to fill large portions of the cave. During low flow, the main stream does not appear in any of the known passages of the cave.

Two main passages were formed in sequence but, as they grew larger, they were partially filled by stream-borne clastic

Fig. 10. Tributary stream passage entering the Maze in Blue Spring Cave, Indiana. The irregular pattern, continuous spans, and solution pockets are typical of the floodwater diversion routes that by-pass the collapse in the main stream passage.

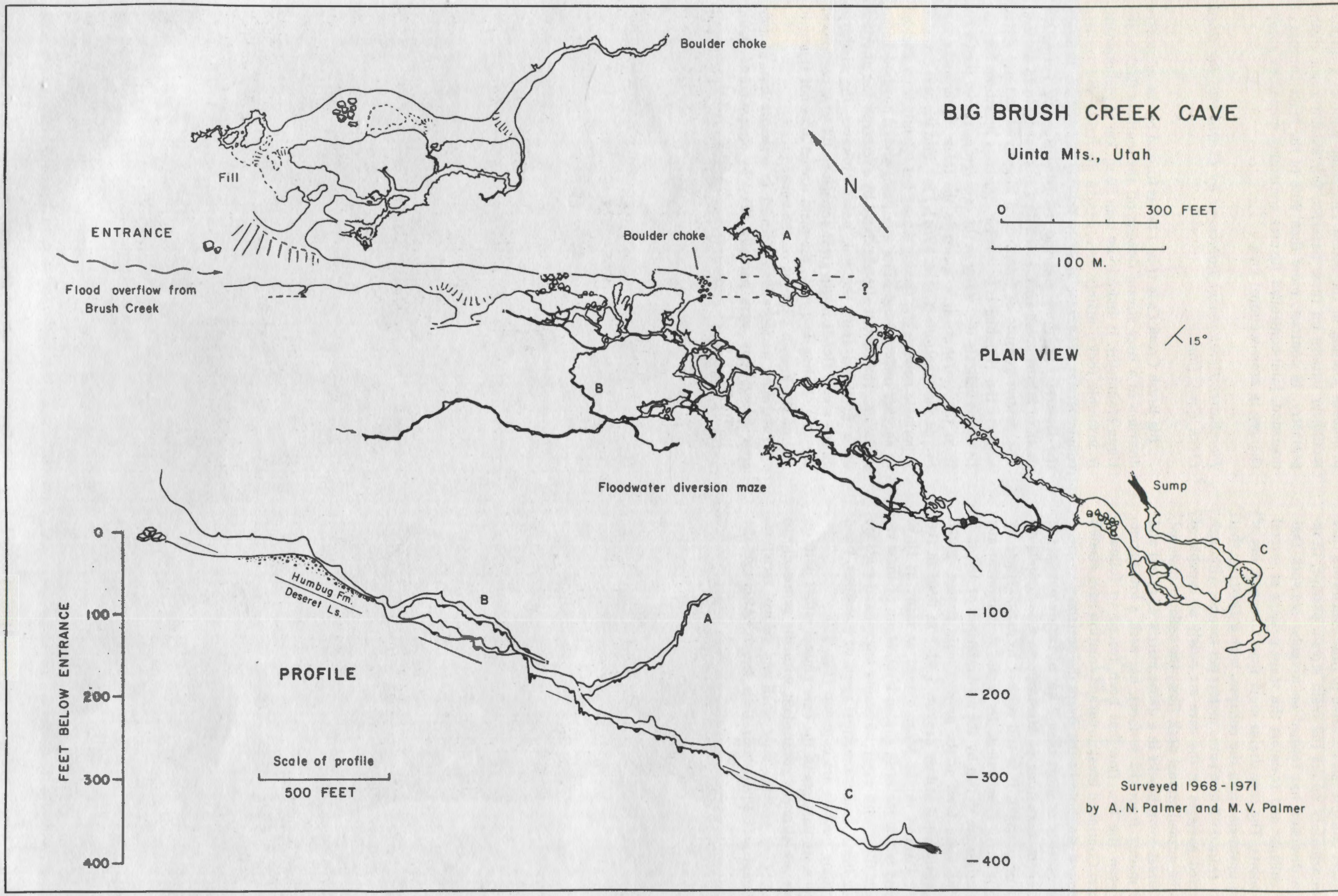

Fig. 11. Map and profile of Big Brush Creek Cave, Utah, showing floodwater diversion routes formed as a result of blocking of the main passages with clastic sediment (Also see Green, 1961).

debris. Quartzite sand, gravel, and boulders as much as half a meter in diameter clog the downstream reaches of these passages. In addition, parts of the cave are nearly filled by vegetal matter, including logs up to 10 m in length. Water has been ponded behind these constrictions, particularly during flood stages, forming a three-dimensional anastomotic diversion maze in the massive, poorly jointed Humbug breccia. The diversion passages are conspicuously ungraded, with intensely scalloped walls and potholed floors. As far as is known, they do not re-join the former main passage, as do those in Blue Spring Cave. Despite considerable differences of solubility between the various rock types in the breccia, local differential solution is not apparent in the areas of high flow velocity, indicating considerable abrasive action. In many areas, however, water ponded under pressure has created intricate spongework (Fig. 12). Thick clay deposits occur in abandoned passages subject only to stagnant backflooding.

Static Flooding behind a Blocked Resurgence: Skull Cave, New York

Skull Cave has developed in highly jointed Silurian-Devonian limestones capped by relatively impermeable limy shale. Its main sources of recharge include two small, intermittent sinking streams that enter the limestone at the contact with the overlying cap-rock and drain southward through the cave, beneath the cap-rock, roughly in the direction of the 2° regional dip. The two streams are confluent only during high flow, when the western stream overflows through a large floodwater channel. The water resurges into a surface valley that has locally breached the shaly cap-rock, but deposits of glacial till in the valley have diverted the flow to a less efficient route. As a result, the stream is ponded in the cave several times each year, filling nearly the entire southern half of the cave. Roughly 3.5 km of highly joint-controlled fissures have been created by aggressive floodwaters forced both laterally and upward from the main stream passage into the surrounding limestone (Fig. 13). Blind fissures have been enlarged upward along joints to maximum heights of 20 m.

The floodwater fissures are not true networks, as they contain few closed loops, yet they are distinctly non-dendritic. Many attain a length of several hundred meters along a single joint, and one can be followed for nearly 600 m in a straight line along several closely spaced joints. The fissures are highly discordant to the limestone bedding, and vary in cross-sectional area by factors of as much as 50 over short distances. Thick deposits of silt and clay mantle the floors and, in many places, the walls, also. The fissures terminate laterally in bedrock, and are not remnants of former flow paths now partially filled with sediment. Solution scallops in the walls of the major fissures, as well as periodically renewed ripple marks in the silt floor, indicate flow on the order of one meter per second *away* from the main stream passage. Scallops increase in length away from the stream and are indistinct in the areas of greatest fissure enlargement, indicating a transition to relatively static water conditions. After flooding, most of the water drains slowly through narrow, joint-controlled shafts and fissures that are partially choked with clay.

Sediment in the floodwater areas has a shielding effect that tends to direct the solutional activity upward. Some passage ceilings have been dissolved upward into the impure, cherty beds at the base of the cap-rock, in which no other significant cave passages are known. The common interpretation that the uppermost levels of a cave are the oldest is not valid in many floodwater situations such as this.

Flooding Caused by Variations in Bedrock Solubility: Onesquethaw Cave, New York

Onesquethaw Cave, in the Onondaga Limestone of Devonian age, intersects thirteen discontinuous beds of chert that in places comprise more than 40 percent of the bedrock thickness. The

Fig. 12. Spongework in the Humbug limestone breccia, Big Brush Creek Cave, Utah, caused by ponding behind constrictions in the main passage. The breccia contains very few joints or bedding planes, so that solution is limited to small fractures and interstices.

differential solubility has resulted in several pronounced constrictions within the main passage. Overflow from a sinking stream periodically fills most of the cave because of ponding upstream from the constrictions, and numerous anastomotic diversion passages have been formed along prominent bedding planes as a result. The origin and geometry of this cave have been treated in detail in a previous paper (Palmer, 1972). As much as 15 cm of solutional and corrasional wall retreat has been measured in the uppermost diversion passages within six years. During a single storm, 4 m of gravel and sand were deposited at a low point in the main passage, causing further constriction of the stream flow. Such figures testify to the dynamic aspects of floodwater cave development.

Maze Development by "Bank Flow" Adjacent to a Surface River: Glen Park Labyrinth, New York

The Glen Park Labyrinth and associated caves consist of several kilometers of strongly joint-controlled maze formed by the insurgence of river water into the limestone during high flow. The caves are located in a low bluff of cherty Trenton Limestone (Ordovician) bordering a steep-gradient reach of the Black River, where it is punctuated by rapids and low waterfalls. The caves nearest river level act as paths for efficient, high-velocity "bank flow" several times a year. Most passages are narrow fissures with scalloped walls (Fig. 14). The scallop sizes indicate water velocities as high as 3 m per second. In general, passage size diminishes away from the river. Sediment consists of thin deposits of silt and clay with a sparse admixture of residual

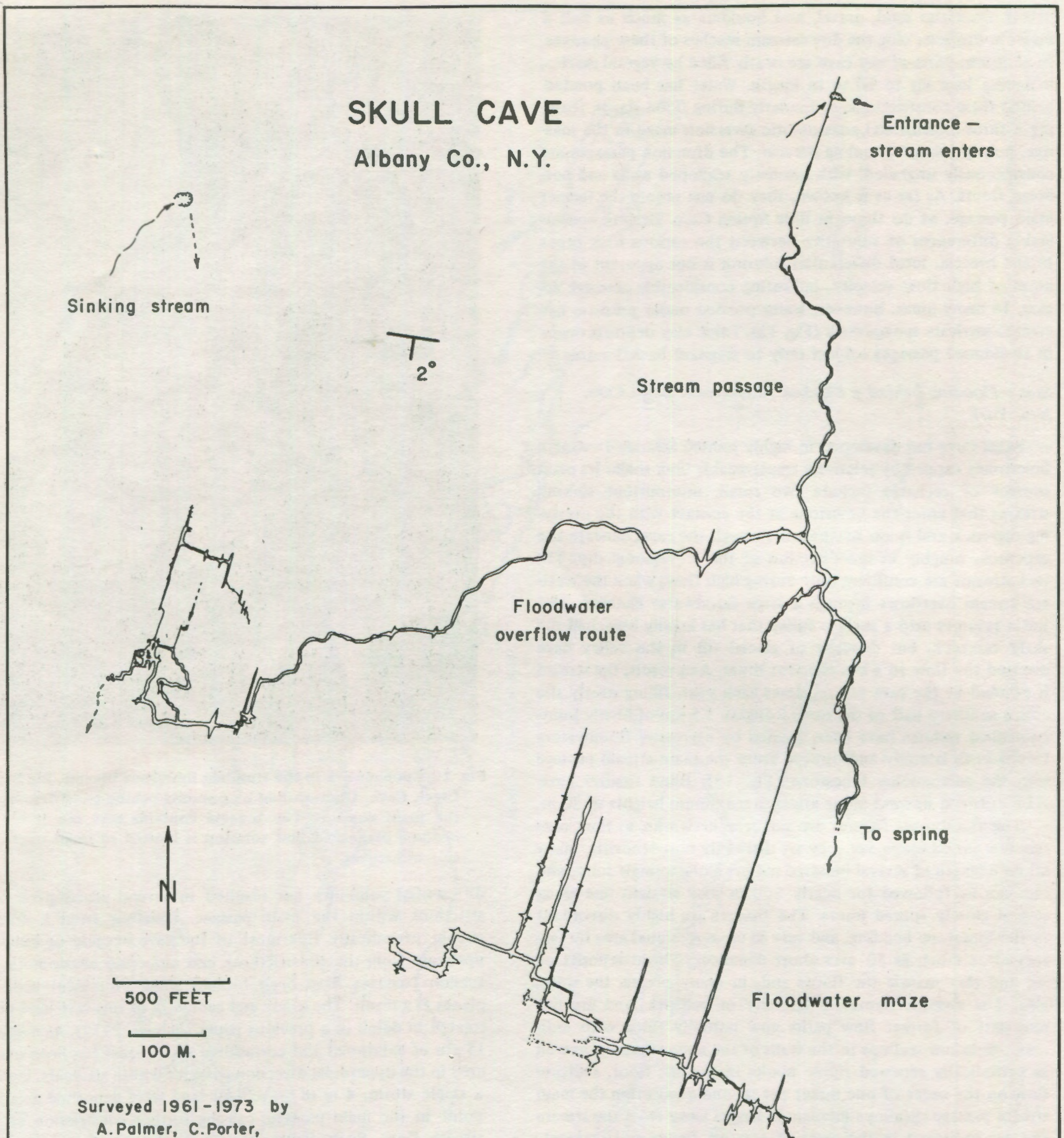

Fig. 13. Map of Skull Cave, New York, showing fissures developed by ponding behind a partially blocked spring. The floodwater passages are confined beneath a shaly cap-rock and have no direct spring outlet (See Palmer, 1962).

chert fragments from the limestone. The caves are nearly devoid of speleothems.

Maze development has been restricted almost entirely to the up-dip edge of the dissected plateau, perhaps because of the greater prominence of joints in those areas. Glacial unloading may have contributed to the well-jointed aspect of the limestone and, therefore, to the strong network pattern of the caves.

Alpine Floodwater Caves

Soluble rock exposed at high altitudes is often subject to rapid and efficient groundwater infiltration, owing to large amounts of precipitation and to sparse soil cover and vegetation. Karren topography predominates at the surface, varying in relief from subdued limestone pavements to a virtually impenetrable terrain of solutionally widened joints with intervening vertical slabs of bedrock. Fluctuations in groundwater discharge are rapid during periods of rainfall or snowmelt, with the result that most alpine caves exhibit local anastomotic maze patterns due to flooding. Excellent examples are found in Europe, including Hölloch in Switzerland (Bögli, 1970) and Eisriesenwelt in Austria (Trimmel, 1966, p. 7). Most caves of this type in North America occur in the Canadian Rockies and in the marble belts of the Sierra Nevada, with isolated examples in the central and northern Rockies of the United States. Typical alpine caves possess ungraded profiles, steep gradients, abundant closed loops, spongework, bedrock spans, and ceiling pockets. Passage patterns commonly resemble those shown in Fig. 11. The orientation of solution scallops indicates many examples of high-velocity flow that has been directed upward through steeply inclined or vertical tubes. Overflows, complex branching, and diversion routes are common, as shown by situations where dye has been traced to several widespread resurgences from a single injection point (Zötl, 1961).

SOME POSSIBLE EFFECTS OF MECHANICAL FRACTURE ENLARGEMENT

Application or release of stress is not only capable of producing fractures in a rock formation, but also of enlarging the fractures to considerable width in local areas. Major sources of stress include tectonic processes, erosional or glacial unloading, gravity sliding, and ice wedging. Many maze-like fissure caves have been formed essentially by these mechanical processes alone (see, for instance, Campbell, 1968). Although these effects are generally obscured in a soluble rock by solution and precipitation, the mechanical enlargement of fractures prior to, or during, solutional cave development may be significant in initiating or maintaining the competitive growth rates among alternate flow paths that are necessary to form a maze.

As mentioned previously, Curl (1971 and 1974) has shown that alternate flow paths between two points in a soluble rock are competitive in solutional growth, regardless of their initial size, provided there is no appreciable change in solute concentration. This analysis has not yet been extended to situations where a change in solute concentration does exist. However, in most incipient caves it is probable that narrower conduits cannot compete in growth rate with larger ones, owing to the rapid decrease in solutional aggressiveness in the narrow conduits with distance of flow. For laminar flow through fractures in soluble rock, the penetration distance (the distance that water can travel before becoming 90% saturated) is proportional to νw^2, where ν = flow velocity and w = fracture width (Weyl, 1958; Wigley, 1971). Because the laminar flow velocity is also proportional to w^2, the distance over which effective solution can take place is extremely sensitive to variations in fracture width. Given the variety of fracture widths in a rock formation, there should normally be a great diversity of dissolved-ion concentrations in the upstream portions of different flow paths. This condition is unfavorable for maze development. But where enlargement of joints is aided by mechanical processes, penetration lengths should increase sufficiently along many alternate flow routes to produce competitive growth.

Field evidence relating mechanical fracture enlargement to the origin of network caves is abundant but ambiguous. Roughly 70% of all the sampled joint-determined networks occur in topographic settings favorable to joint enlargement by erosional unloading: 35% of the total are located within several hundred meters of steep cliffs or escarpments (e.g., Clarks Cave, Fig. 4); 26% are located in isolated, residual hills of soluble rock (e.g., Peiper Cave, Fig. 6); and 9% occur in the vicinity of large collapse dolines, where the release of stress in the surrounding

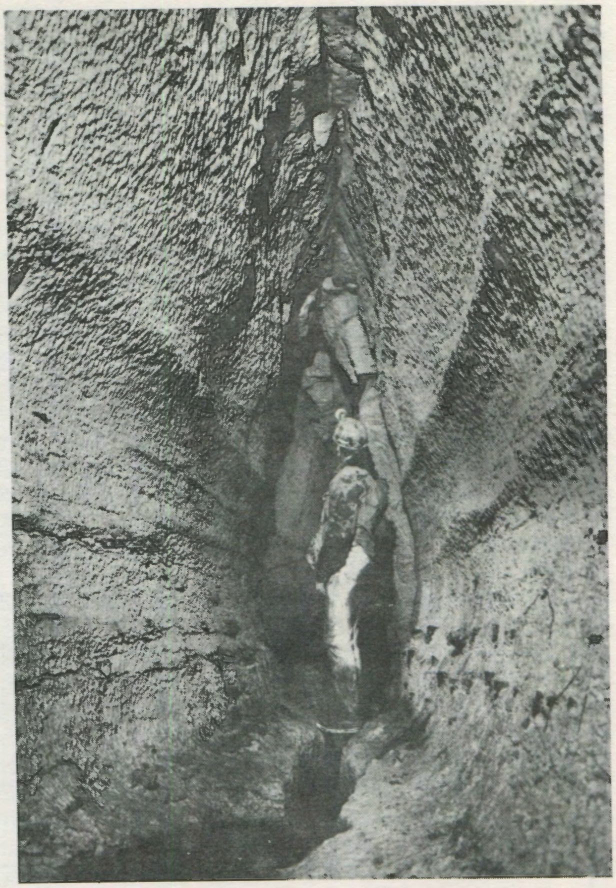

Fig. 14. Glen Park Labyrinth, near Watertown, New York. Solution scallops in the walls indicate the high velocity attained by overflow from nearby Black River.

bedrock may have caused local joint enlargement (e.g., the Maze in Blue Spring Cave, Fig. 9). In addition, roughly 40% of the network caves in the sample are located in regions that have been subjected to Pleistocene continental glaciation, which is probably a significant factor in enlarging joints indirectly through isostatic rebound. In all, only 16% of the sampled network mazes occur in unglaciated regions remote from steep topography where there is no apparent source for more than the ordinary amount of mechanical fracture enlargement. Yet, determining the exact influence of mechanical processes is difficult, for all but a few of the sampled network mazes are as easily accounted for by other factors described elsewhere in this paper. Mechanical enlargement of fractures is probably significant in enhancing network development under the conditions of groundwater recharge described earlier, but alone is probably of limited importance.

Prominent joints are common along the axes of folds in competent bedrock, owing partly to localized tension during deformation. Several well-known network caves are located at the crests of anticlines, notably in the folded Appalachians (Davies, 1958, p. 14, 17; and 1960, p. 6). Although Davies did not interpret their origin, the network pattern of these caves has popularly been attributed to the abundance of favorable flow routes offered by prominent intersecting joints. Yet only 3% of the total sample of maze caves are located along the crests or axes of visually discernable anticlines, so this relationship, even if valid, must be considered of relatively minor importance. At least as many caves are known to intersect folds with no tendency toward network development.

The most frequently cited example of a network cave located along the crest of an anticline is Hamilton Cave, West Virginia. It is especially favorable for study because of its proximity to several non-network caves, as shown in Fig. 15. The anticline in which the cave is located is a broad, plunging flexure in the flank of a much larger fold, which forms a ridge that is locally truncated by a water gap of the South Branch of the Potomac River. Owing to the asymmetry of the anticlinal flexure, the hinge line is located several hundred meters east of the crest, in a position occupied by Trout Cave, which exhibits both rudimentary branchwork and network characteristics. New Trout Cave, essentially a single tube, is located farther down the east flank of the fold, where the dip is 35 to 40°, at an elevation 60 m below that of Hamilton Cave. All three are developed in the cherty New Scotland Formation (Devonian), which is overlain by the Oriskany Sandstone. Both Hamilton Cave and the network portions of Trout Cave are located at or near the anticlinal crest, where the dip is less than 10°, and are near the top of a high bluff of limestone 5 to 15 m below the base of the sandstone. The tubular passages are restricted to the more steeply dipping beds and occupy a position slightly lower in the rock sequence.

Tension along the crest of the anticline has probably been a factor in determining the strong joint control of the networks. Yet, along the hinge line, where the stratal curvature is greatest and tension presumably most intense, the passages are tubular. Thickness of the overlying bedrock is apparently the key factor in selective enlargement of the joints, for, of all the passages, the networks are located closest to the overlying land surface, where joint enlargement by erosional unloading has been most pronounced. Groundwater recharge to the mazes during their development was probably contributed both by diffuse recharge through the overlying surface and by backflooding from the river early in the history of the water gap.

In certain limestones, particularly reefs, the primary pores are so large and profuse that competitive flow paths may occur even before the openings are enlarged by solution. For example, the Capitan reef of Permian age, in which Carlsbad Caverns is located, contains primary pores that range in diameter up to several centimeters, as shown in well cores and man-made tunnels. Though chiefly influenced by large joints, the pattern of Carlsbad Caverns locally reflects the irregular pore structure within the limestone, particularly in zones of spongework (Fig. 7). Diffuse recharge and flow within limestone of this type may be analogous to that of a network cave in which joints have been widened partly by non-solutional means.

"ARTESIAN" NETWORKS: SOME CONTRASTING VIEWS

Artesian flow through certain confined limestone formations has been well documented, particularly in the Mississippian limestones of South Dakota (Swenson, 1968). Most caves in such formations possess network patterns and, through this association, many interpreters of cave origin have attributed their pattern to solution by slow-moving artesian water. This view is accepted by the majority of karst researchers. Despite strong evidence to support this idea, however, several factors other than artesian flow must be given comparable weight on the basis of hydrologic evidence. Although they comprise less than 3% of the total sample for this paper, the caves in question include some of the largest network mazes in the world. For this reason,

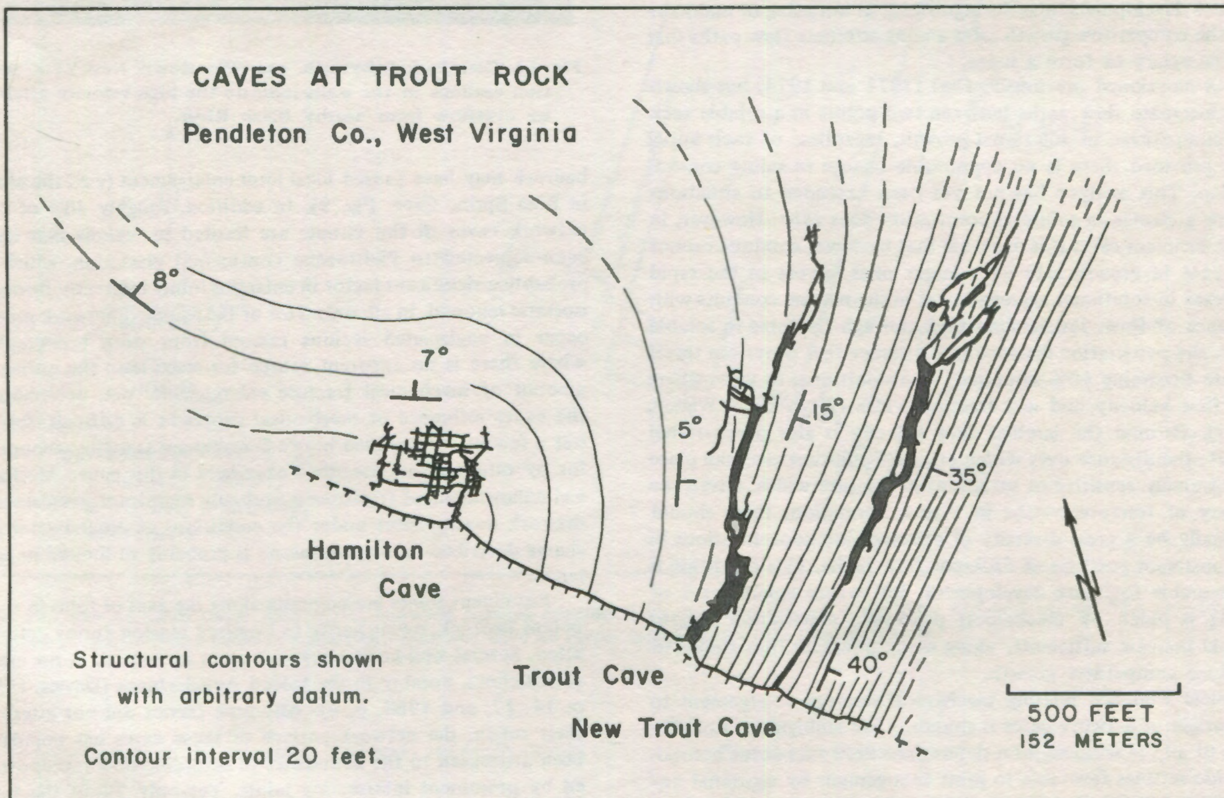

Fig. 15. Structural setting of Hamilton, Trout, and New Trout Caves, West Virginia. Cave maps from Davies, 1958. Structural data by A. Palmer, M. Palmer, J. Allen, and P. Bongei, April, 1969.

it is worthwhile to outline here the major contrasting ideas regarding their origin, even though the issue cannot be resolved at this time.

Proponents of an artesian origin for network caves imply that they are formed by water confined under hydrostatic pressure beneath an impermeable formation. Yet, in the best-documented examples of North America, the existing networks associated with artesian aquifers lie outside the present limits of confinement and are overlain for the most part by thin, relatively permeable sandstone. Examples include the caves of the Black Hills in South Dakota (Tullis and Gries, 1938; Neighbor, 1954; Howard, 1964), the caves at Burnsville Cove in Virginia (Deike, 1960; N.W. Davis, 1971), and the network caves beneath the Hartselle Sandstone in Alabama (Varnedoe, 1964). Where the limestones are covered by considerably less permeable rocks farther down-dip, evidence for network caves is absent, or at best ambiguous. Two well-known examples are discussed in the following paragraphs. Both cave areas are presently undergoing extensive study by speleological groups, so considerably more information pertinent to this discussion should be available in the near future.

The Burnsville Cove system, Virginia, consists of more than 30 km of interrelated caves in Silurian-Devonian limestone. The caves and their geologic setting have been described in detail in the works cited above. Drainage enters the limestone along the entrenched Bullpasture River, which transsects the regional structure. Deike (1960) has presented a strong case favoring an artesian origin for the cave. Artesian flow still exists in parts of the downstream end of the system. Upstream passages along the flank of the syncline, particularly Breathing Cave and the entrance area of Butler Cave, consist of intricate networks overlain by beds of quartzose sandstone. Networks are most prominent in the areas of groundwater recharge around the periphery of the limestone aquifer, but this pattern is subdued or absent in the downstream passages of the system, where artesian confinement should have been more pronounced.

The nature of the recharge that formed the caves was of two possible types, both of which have been demonstrated earlier in this paper to promote maze development: uniform infiltration through the locally thin, well-jointed sandstones overlying the peripheral areas of the system; or floodwater recharge from sinking streams draining from the insoluble highlands to the west. The latter mode of recharge is compatible with the present drainage pattern and was probably of great significance in the past, as indicated by extensive cave deposits of coarse sand and gravel with grain diameters commonly as great as 30 cm. Although episodes of filling with coarse sediment do not necessarily have a direct relationship to the solutional enlargement of a cave, the availability of floodwater recharge could have been the dominant factor influencing the cave origin and pattern.

No example of artesian cave origin has been more widely accepted than the caves of the Black Hills, South Dakota, yet, again, there are factors other than artesian flow that must be recognized as having considerable bearing on their origin. Extensive network caves are developed in the Mississippian Pahasapa Limestone where it is exposed along the flanks of the Black Hills uplift. Of these, the best known are Jewel and Wind caves, which are among the longest caves in the world. More than 80 km of passages have been mapped in Jewel Cave alone. Water enters the limestone where it crops out within the Black Hills and, although some of the flow resurges through the overlying Minnelusa Sandstone at springs along the base of the uplift, the bulk of the flow apparently passes eastward through the limestone for nearly 200 km under artesian conditions, eventually rising into the younger Dakota Sandstone in areas where the intervening formations are thin or absent (Swenson, 1968).

The assignment of an artesian origin to these caves by previous workers has been based on several factors. The caves are not only closely associated with a known artesian aquifer, but the lack of coarse clastic sediment implies solution by slowly percolating water, an idea that is seemingly supported by the lack of well-defined master flow routes in these complex networks. In addition, there is evidence for mineralization at high temperature and pressure within the caves (White and Deike, 1962), although such evidence is unequivocal only for calcite vein fillings that, apparently, pre-date cave development. Fine-grained sediments on the upper surfaces of the protruding veins, covered now by secondary calcite overgrowth, signify that cave development took place after the calcite vein filling, but prior to the secondary overgrowth. There is no clear evidence that the secondary calcite was deposited under high temperature and pressure. There is strong evidence for several episodes of flooding by low-velocity water late in the history of the cave. (See Deal, 1965, for a detailed interpretation of the clastic and secondary mineral deposits of Jewel Cave.) The caves generally lie 30 to 120 m above the present water table and do not receive significant recharge, which makes an interpretation of their origin difficult.

Evaluating their setting in terms of the criteria used earlier in this paper, the major caves of the Black Hills are largely confined to hills of limestone capped by thin sandstone or only recently stripped of their sandstone cover. On a regional scale, cave development is limited to the strike-oriented band of exposed limestone. The overlying sandstone is rather permeable, for groundwater flow passes upward across the sandstone-limestone contact several kilometers to the south of and down-dip from the cave area. Howard (1964, p. 9), Wilber (1962, p. 2), and Deal (1965, p. 69) envision the artesian aquifer as including both the limestone and sandstone. Most solutional openings are located in the upper parts of the exposed limestone, possibly owing to the dolomitic content of the lower strata (Howard, 1964, p. 9).

If the Black Hills network caves truly owe their origin to artesian flow, it would be a necessary corollary that significant cave development has occurred in those down-dip areas of the limestone aquifer that are still under artesian conditions, whether or not such development is continuing today. Hydrologic conditions cast doubt upon this hypothesis, for the measured hydraulic gradients are too steep to allow more than rudimentary joint enlargement to exist. Two examples are given below.

(1) The limestone down-dip from Jewel and Wind caves possesses a hydraulic gradient of roughly 0.005 (Howard, 1964, p. 9). Within the known caves, the average spacing between parallel fissures is roughly 20 m, and most of the fissures have been enlarged by solution over a vertical range of at least ten meters. In some areas, the aggregate height of superimposed passages is as great as 60 m. From Fig. 16, it is clear that even a minor amount of solutional enlargement of joints would require impossibly large amounts of flow to maintain a gradient of 0.005. For example, if solution down-dip from the caves in the present phreatic zone were limited to a vertical range of only ten meters within a section of limestone as wide as Jewel Cave (approximately 1500 m), no more than 16 cm of solutional widening of each joint would be needed to provide a discharge greater than the entire mean runoff from the state of South Dakota (roughly 100 cubic meters per second).

(2) The overall hydraulic gradient in the artesian aquifer east of the Black Hills is 0.001, with a minimum value of 2.3 x 10^{-4}, a condition that led Swenson (1968, p. 179) to conclude that major solution openings exist in the limestone to distances as great as 200 km from the recharge area. However, it is apparent from Fig. 16 that even under such low gradients, only a

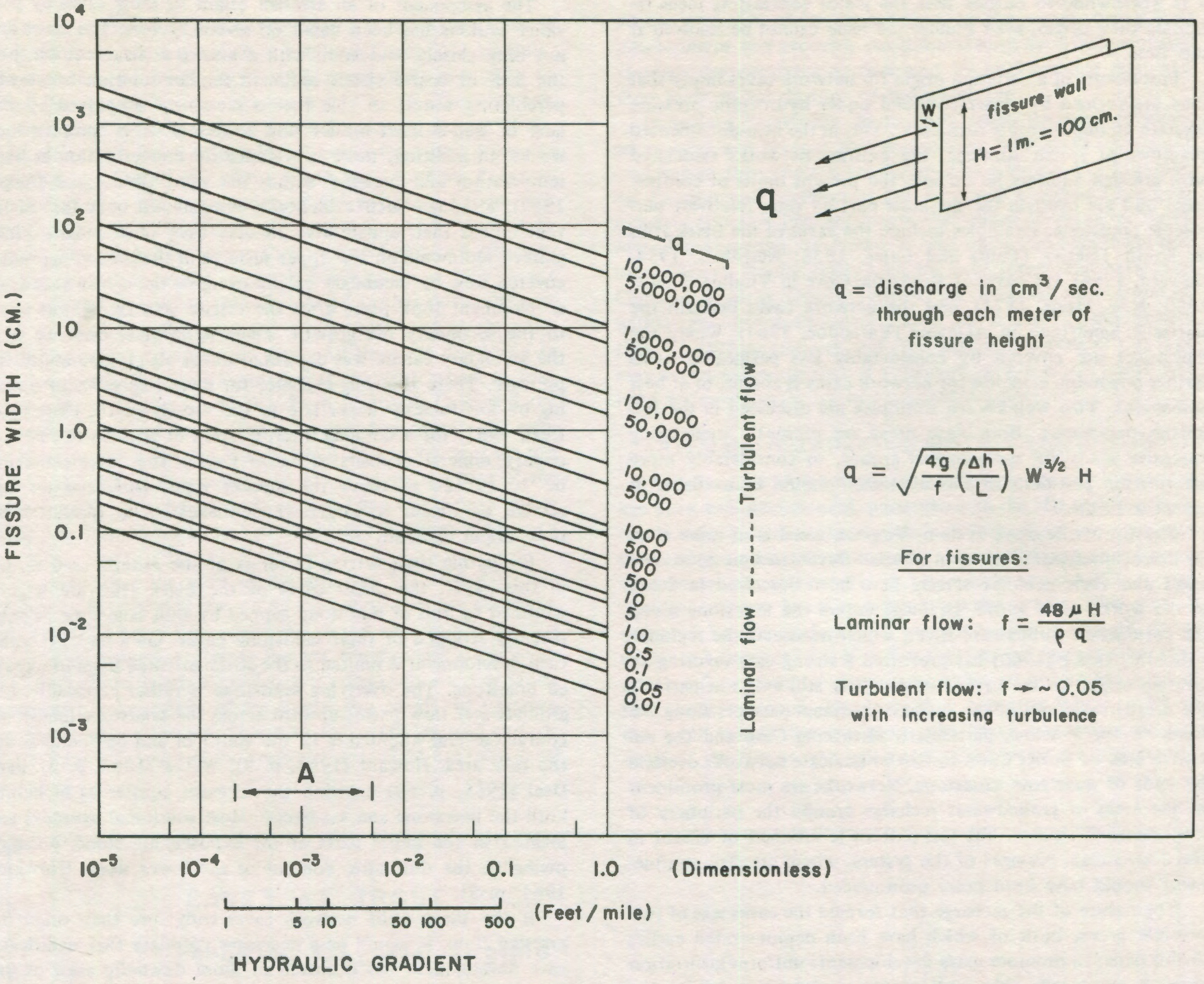

Fig. 16. Relationship between discharge (q), hydraulic gradient $\left(\frac{\Delta h}{L}\right)$, and fissure width (w) for the flow of water through a planar opening under phreatic conditions, where q = gravitational field strength, μ = water viscosity (assumed here at a temperature of 10°C), ρ = water density, and f = friction factor. Values are only approximate under field conditions, because of variations in temperature, uncertainty of the roughness factor under turbulent flow, and irregularities in conduit configuration (For further details see Binder, 1962, p. 109). A = observed range of values for hydraulic gradient in the limestones of the South Dakota artesian aquifer.

few cavern-sized openings would require more recharge than could be supplied along the exposed edges of the aquifer, even if the entire mean runoff of 5.7 cubic meters per second from the Black Hills were to be consumed. For network caves to develop, the flow would necessarily be distributed among thousands of joints over the 220-km north-south width of the aquifer in South Dakota. The effective width of each solutionally enlarged fissure could be no more than a few millimeters. Fissures of this size transmit water readily, but they hardly fit the popular conception of a cave. Considering the gross overestimation of flow, the actual openings are undoubtedly smaller on the average. Any large solutional openings encountered by wells must be of only local extent.

It is reasonable to expect the rate of solutional cave enlargement to be greatest in the recharge areas, where the water is most aggressive and, apparently, this is the case in the Black Hills. Again, it appears that maze development depends strongly upon the nature of groundwater flow in the recharge areas. Water is known to enter the limestone through partially sinking streams, and infiltrating water is capable of passing downward in diffuse form through the surfaces overlying the caves, although the latter mode of recharge is presently very limited. The small amount of solution occurring under artesian conditions may have delineated the ultimate cave patterns but, if so, it is not clear why such a pattern fails to be produced by the minute amounts of deep-phreatic solution preceding the major phases of cave enlargement in every karst area.

In summary, it appears that artesian conditions have had a questionable influence upon the development of known network caves. The association of several large network mazes with artesian flow systems appears to be dependent only upon favorable conditions of recharge in the areas of limestone outcrop.

CONCLUSIONS

From an investigation of their geologic, hydrologic, and topographic settings, it is apparent that maze caves are largely situated in areas where groundwater recharge to the soluble rock occurs either as diffuse infiltration or as concentrated, highly variable floodwaters. Maze development is rare in karst aquifers that receive uniform recharge through limited point sources such as sinkholes.

The specific pattern of a maze cave depends on both its mode of origin and the nature of its host rock type. Network caves, particularly those with a dense pattern of rather uniform passages, are normally the product of diffuse recharge from the overlying land surface. Floodwater activity may also form networks, but normally with a more irregular, less rectilinear passage pattern. Tectonic and isostatic stresses may be significant in accentuating the joint patterns in the bedrock that are necessary for the development of network caves. Anastomotic mazes are almost invariably the product of floodwater activity in rock that lacks prominent joints. With greater joint prominence, this type of maze approaches the character of irregular floodwater networks. Spongework mazes are extensively developed only in massive formations, although local spongework may occur in virtually any soluble rock type. Spongework is the product of ponding upstream from passage constrictions during periodic flooding, or by slow, diffuse flow in limestone containing primary pores large enough to allow competition in growth among numerous flow paths.

The genetic categories into which the known maze caves have been placed may not represent all possible classes of maze origin, and in many cases the exact details of cave development have yet to be determined. However, the overall relationships stated here should provide broad criteria for predicting or explaining the existence of maze caves, and it is hoped that this paper will lend perspective and direction to future, more detailed work.

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