BULLETIN

OF THE

NATIONAL SPELEOLOGICAL SOCIETY

VOLUME TWENTY-FIVE PART TWO

Contents

JULY 1963

BULLETIN

of

THE NATIONAL SPELEOLOGICAL SOCIETY

VOLUME 25, PART 2 **JULY 1963**

Published twice a year; Editor William E. Davies, 125 W. Greenway Blvd., Falls Church, Va.; Associate Editors: Nancy G. Rogers, George W. Moore. Assistant Editors: Thomas C. Barr, Jr., Richard Anderson, Kenneth Christiansen, Rane Curl, John V. Thrailkill. Inquiries relating to the publishing of manuscripts in the BULLETIN should be addressed to the editor.

DISCUSSION of papers published in the BULLETIN is invited. Discussions should be 2000 words or less in length with not more than 3 illustrations. Discussions of papers in this issue should be forwarded to the editor by December 31, 1963.

Office Address:

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2318 N. KENMORE ST.
ARLINGTON 1, VIRGINIA

Subscription rate in effect January 1, 1963: $4.00.

The Development of Karst Features

by Alan D. Howard

ABSTRACT——Karst landforms may be considered to be largely composed of three features: cutters, sinks, and caves. These are interrelated and transitional between each other.

Cutters are solutionally enlarged joints. They are normally filled with a residual terra rosa, and form beneath such a soil cover. Cutters are a distinct karst landform, although they are often difficult to recognize because they may be obscured by the soil material. Cutters generally are not remnants of former caves. Where cutters are the dominant karst landform, joint solution generally decreases uniformly downwards. This solution is directly related to present topography in depth, intensity, and directional development. Cutters are normally developed into a crude, dendritic network similar in directional development to the overlying topography. Cutters originate through solution by ground water flowing laterally within the enlarged joints. This is a gravity flow under a gradient. The amount of development of cutters depends upon the climate, the rock properties, such as solubility and grain size, and the geologic-topographic situation.

Cutters are the only karst form present in completely homogeneous soluble rock. Under such conditions they are an equilibrium feature of the landscape, and they have a simple form that changes little with time. Where the rocks are heterogeneous, such as in area of associated soluble and non-soluble rocks, the greater will be the extent of development of sinks and caverns, and the cutter will be less prominent.

Cavernous passages result from solution of the bedrock by through-flowing ground water. Because of the necessity for continuous addition and removal of ground water in order for cavernous channels to form, caves are continuous conduits leading from a source of ground water to a ground water exit. No one theory of cave development or sequence of cave-forming events can be common to all caves. Every cave has a unique history, but nevertheless there are certain broad principles which are true of the development of all caves. Caves appear to fall into two classes, those that arise through solution and abrasion by free-surface ground water or ground water streams flowing with a definite gradient, and those that result by solution by ground water flowing under artesian pressure. Some caves have records of both processes having acted at different times.

Caves result from the presence of a favorable geologic situation (stratigraphy and structure). Both the geology and the geologic effects upon topography are factors which control the development of caverns. In order to completely describe the origin of a cave, the specific geologic-topographic relationship which promoted its development must be specified.

Sinks are enclosed topographic depressions which are collection areas for the diversion of surface water underground. Sinks presuppose caverns, but the reverse is not necessarily true. Most sinks derive their topographic form from the continuous removal of soil material underground, and the sink landform is usually a structure of the soil mantle only, for the bedrock surface does not usually have an associated funnel shape. Only a few sinks originate from the collapse of cavern roofs, and even these are perpetuated by continuous downward removal of increeping soil material.

Caves and sinks are most common in areas of great inhomogeneities of geology. In contrast to the rather static and monotonous topographic forms prevailing in areas of homogeneous rock, the topography in areas of great diver-

sity of geology is generally varied and typically in a state of flux. Such conditions, which are the result of the influences of various lithologies and structures upon the topography, are often conducive to the development of caves and sinks when soluble rocks are present.

Because caves result from the effects of stratigraphy and structure upon topography, neither uplift nor peneplanation is called upon as a direct causal agent for the development of caves.

Cutters are also found in areas of diverse geology, but are present in inverse proportion to the degree of diversity. In such areas cutters will have various forms, and these may approach that of caves.

Karst in this paper includes all landforms on or above soluble rocks which are attributable to solution by ground and surface water. Most landforms on soluble rock are a hybrid of solutional processes and other erosional processes, and landforms on rocks not ordinarily considered soluble (pseudokarst) may resemble those of the true karst type.

Karst landscapes may be considered composed mainly of three forms: cutters (karren), sinks, and caves. These are interrelated and transitional between each other. Of these features, cutters have not previously been thought to be a major karst form, largely because of the lack of expression of these bedrock solutional forms on the surface topography and the general lack of exposure of cutters.

Cutters——In many places on limestone a peculiar bedrock topography is found where the soil overburden has been removed through erosional processes or by man. This topography is typified by an exaggerated valley and ridge appearance, the positive features termed *pinnacles* (karren) and the negative features called *cutters* (Smith and Whitlatch, 1940, pp. 46-47).

An easily accessible example of this solutional form is located just south of Bedford, Indiana on Highway U. S. 50. Limestone is exposed in the roadcut and on the surface adjacent to the cut where overburden has washed away from the limestone exposing an area of well-developed cutters (fig. 1). The negative areas of the surface are greatly enlarged joints, and the positive areas are intervening bedrock blocks. Because the joints are vertical, the cutters are also vertical. The greatest enlargement is along a set of joints that are spaced about 5 to 10 feet apart and strike at approximately right angles to the road. The cutters on the main joints extend downwards below the level of the road cut and are filled with clay. One or more secondary series of joints, closer together and less prominent, are hosts to secondary furrows which lead at a steep gradient from the higher blocks into the main cutters. They appear to be tributaries to the main channels. It is apparent that the main cutters are not funnel-shaped forms attributed to sinks in limestone terrain, for, although they have a steep funnel form in cross section, they extend linearly in the direction of the joints, and are of a smaller scale than sink features. These solution channels are normally filled with red clay, the usual residual *terra rosa.*

The surfaces of the limestone protuberances are characteristically rounded and streamlined, with conical peaks; their surface features are characteristic of solutional origin, with sharp projections and bowl-shaped depressions. Where only the pinnacles protrude above the soil cover and the trenches are filled with clay, the isolated and usually numerous bedrock exposures are termed *karren.*

In a number of limestone quarries near Bloomington, Indiana, quarrying of the Salem limestone exposes excellent sections across the limestone. These sections are well defined because the quarrying is by sawing rather than by blasting. Here enlarged joints similar to the ones at the roadcut were observed (fig. 2). The top parts of the pinnacles were not exposed but the section continued down enough to expose the lowest part of the enlarged joints. The cutters extended to variable depths beneath the top of the quarry and almost all decreased in

Figure 1
Pinnacle and cutter bedrock topography, southern Indiana

Figure 2
Joints enlarged by solution (cutters) in a quarry face in southern Indiana.

width rather uniformly downwards: all disappeared by pinching-out before reaching the bottoms of the quarries about 60 feet below. Almost all openings extended to the top of the cut, and expanded upwards. Only a few, small roofed cavities were noted, and these were usually beneath normal cutters, along the same joint. The cutters were filled with clay as in the roadcut. The Salem limestone in this area contains few sinks and the cutters in the quarry were not funnel-shaped bedrock depressions ascribed to limestone sinks.

Cutters when cleared of clay closely resemble cavern passages because of their development along joints and the solutional features they display. The possibility that they are cavern remnants which have lost their roofs by collapse should be considered. However, several factors make this unlikely:

1. In these localities and at others, there is no evidence that extensive roof collapse has occurred in pre-existing cavities. There are no buried flowstone deposits, no relics of speleothems, no blocks of roof limestone in cave fill, and no coarse clastics within the fills as one might expect from stream deposits within caverns.

2. Pinnacles and enlarged joints appear to be too numerous to be ascribed to former caves.

3. The tops of the pinnacles have been greatly affected by solution and are not fresh bedrock that one might expect to find as part of an earlier cavern roof.

4. Usually the joint openings expand upwards, showing no tendency to close over the cutter.

5. No caverns of significent size or numbers were noted below the cutters, which, in contrast, were very numerous near the surface.

All the cutters and pinnacles that I have seen have been formed under a soil cover. *Lapies,* a solution form which develops on subaerially exposed limestone and which generally occurs as small scale ridges and hollows, has often been mistakenly equated with cutters (Thornbury, 1954, p. 319-320) but it is a distinct form and is often developed on pinnacles that have been bared by removal of the soil mantle.

Different solubility of bedrock gives rise to secondary features on cutters and pinnacles. Chemically resistant beds generally form protuberances into the cutters, and weak beds are cut back. Figure 3 shows the effects of resistant beds at various positions within a cutter. The last form (c) appears to be the most abundant where there is a prominent resistant bed, for, being most resistant to chemical weathering, it persists longer at the surface of the bedrock than less resistant beds. Figure 4 shows a commonly-observed profile through cutters. It should be noted that such cutters are pene-caves, that is, they are nearly roofed over. In fact, it is observed that a resistant bed may entirely roof over a cutter for short stretches forming natural bridges or short caves. Also it was noted that solution is somewhat restricted beneath such a resistant cap, allowing only stunted cutters to form, because the downward movement of solvents in such cases is hindered. Similarly solution increased correspondingly above such a resistant layer, and short caves may also form by abnormal solution above the resistant layer. The number of these, however, is probably smaller than the number found beneath resistant beds.

An example of solution retardation by resistant beds is on Highway 31 north of Columbia, Tennessee. On the east side of this roadcut, cutters, being nearly perpendicular to the roadcut, are exposed in cross section (fig. 5). On the north end of the section are a number of apparently normal cutters, with bottoms not visible. On the extreme south end of the cut a relatively insoluble layer is present above the soluble, cutter-forming unit, and therefore no cutters have developed in this zone and the contact between bedrock and the mantle is essentially flat. Intermediate conditions are present in the center of the section where one cutter has a significant pinching in of its upper parts as compared to the usual widening. Where the resistant unit is thicker, the top is closed over to form a clay filled, cave-like opening of indeterminable extension into the hill. Taking the section as a whole, where the upper resistant unit is present, little vertical solution along

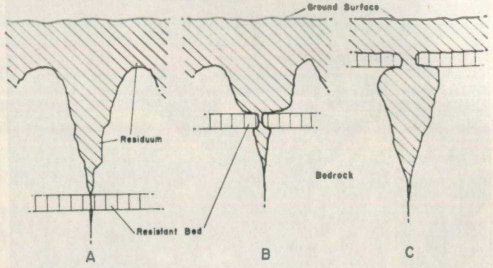

Figure 3
Cutters, showing the effect of a relatively resistant bed at three positions within the cutters.

the joints has occurred, and where absent the joints have widened into cutters.

The phosphate district of central Tennessee affords many excellent exposures of cutter topography because the phosphate-rich mantle has been removed from the limestone bedrock.

Cutters and pinnacles were also observed in tilted and deformed rocks, even where the dips of the rocks approximated 45 degrees. Despite the angle of the bedding, where the rock is essentially homogeneous, the dominant lineaments of the upper parts of the cutters are vertical, as opposed to the possibility of control by bedding or oblique joints (Watson, 1905). In the lower extensions of cutters, control by bedding and inclined fractures becomes more noticeable (e.g., in the marble quarries of eastern Tennessee).

The *pepino hills* of Puerto Rico, the *mogotes* in Cuba, and similar karst forms in other areas have almost universally been interpreted as being old age remnants of a former karst plain (Thornbury, 1954, p. 334). These features are steep-sided towers of limestone which protrude out of flat-floored valleys underlain by unconsolidated residual material. I disagree with the above conventional interpretation, for these towers seem to be more closely related to cutters in origin. They are essentially greatly enlarged joints, and the tops of the pinnacles have been exposed by concomitant removal of the residuum as the joints are deepened by solution. They are very close in origin to the *karren* terrains in the United States, but

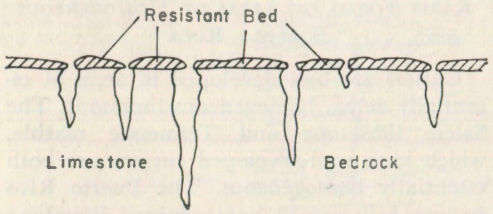

Figure 4
Section through cutters developed under resistant "cap rock".

the wider spacing of the joints and the different climatic conditions of the tropics have resulted in the difference in scale. Monroe (1960) notes that in Puerto Rico the tower forms are present in homogeneous limestone, while sink topography is present where the strata are alternately hard and soft.

It appears unquestionable that solution from which cutters are derived occurs while the cutters are filled with clay. A few interesting studies have been made of the depth of such solution, and the vertical distribution of solution in areas of enlarged joints. D. K. Hamilton (1948) has given a comprehensive treatment of the ground water occurrence in a carbonate rock area near Lexington, Kentucky. He notes that solution occurs along joints and bedding planes, and is directly related to present topography. Beneath both uplands and topographic lows, solutional enlargement was present to a depth not exceeding about 80 feet (fig. 6) and the amount of solution decreased downwards fairly uniformly. He notes that insoluble and impermeable beds inhibit solution beneath them and concentrate solutional effects just above them. Joints which are located beneath topographic lows are enlarged the greatest amount, and least enlargment of joints occurs beneath topographic highs. The solution pattern cannot be correlated with a nearly horizontal, well defined water table and solution development of enlarged joints and bedding planes seems to be patterned into a crude, three dimension connected network, similar in directional development to the overlying topography (fig. 7). Such considerations point to the conclusion that extensive lateral flow of ground water exists within the joints and bedding channels, and that cutters are, in effect, a sub-soil drainage network. The flow of water in cutters would be gravity flow with an assignable gradient. Solution would decrease with depth because of the limited vertical mixing of the ground-water solutions. Walker (1956, figs. 7 and 16) gives similar evidence of the depth of solution. In some localities solution openings have been found at great depth, up to 400 feet or more (Foose, 1953; Moneymaker, 1941). The anomalous openings at depth are probably the result of unusual lithologies and structures promoting deep flow of ground water.

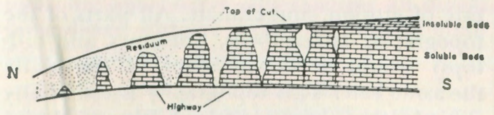

Figure 5
Generalized section of a roadcut north of Columbia, Tennessee, on highway U. S. 31.

Generally the soil-bedrock contact is quite sharp. This would indicate that solution is by surface attack only. The general homogeneity of limestone, its intergranular impermeability, the great volume reduction upon solution, and the rate of solution are probably factors that contribute to the sharp interface. More transitional soil-bedrock contacts are noted in impure limestones and dolomites (Rodgers, 1953, p. 116). Probable reasons for this are the more porous nature of these rocks, allowing more intergranular solution, and the slower rate of reaction, which makes intergranular solution more prominent.

Cutters are best developed in dense, pure limestone with an even, well-developed joint system, and in warm, damp climates. As limestone becomes more impure and more resistant, and the climate becomes less moist, the bedrock-mantle contact is flatter and the soil mantle thinner.

Some granites in arid climates develop a pinnacle topography that is probably closely related in origin to cutters in limestone. Such topography is found in homogeneous granite with even, well developed, coarse

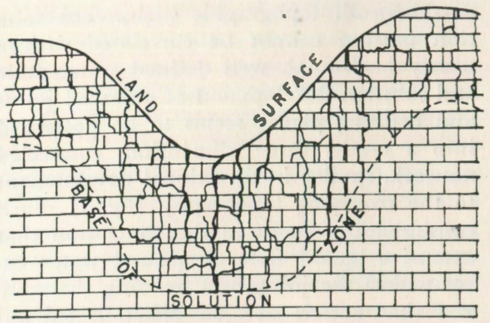

Figure 6

Zone of solution in a valley developed in essentially homogeneous rocks, (Hamilton, 1948, figure 4).

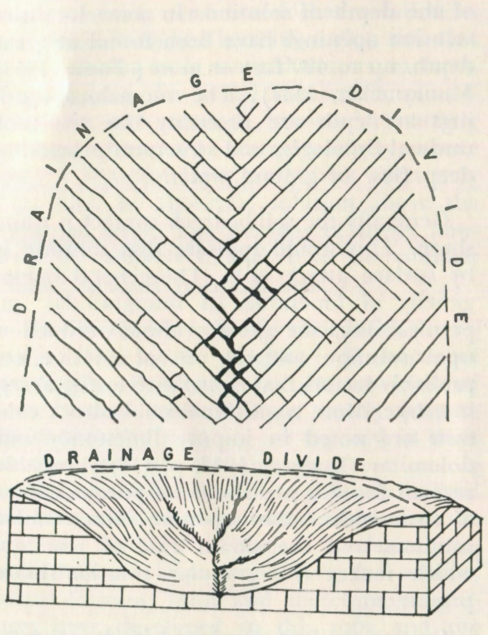

Figure 7

Directional solution along joint planes discordant with drainage, (Hamilton, 1948).

textured joints of vertical orientation. The arid climate that seems necessary for the development of this topography probably results in weathering of the granite by surface attack only, and not by intergranular disintegration.

Karst Forms of Areas of Homogeneous Soluble Rock

Cutters are best developed in areas of essentially dense, homogeneous limestone. The Salem limsetone and Tennessee marble, which have well-developed cutters, are both essentially homogeneous. The Puerto Rico *pepino hills* are in homogeneous limestone and many of the areas in which cutters have been observed in central and western Tennessee are in monotonous topography developed on thick, nearly homogeneous limestones and dolomites.

The form of cutters in homogeneous rock is that of trenches along the joints narrowing downwards rather uniformly. These cutters have sizes and depths of varying orders of magnitude and are so arranged to form a crude sub-soil drainage network. Cutters are also found in areas of variable geology and this will be considered in a later part of this paper.

A critical point to be determined is the topographic elements that might be expected to develop in an area of homogeneous rock. Classical theory assumes that the topography depends mainly upon the stage of landform evolution, and that whether one should expect surface drainage or subterranean drainage through caverns depends also upon the stage of the karst cycle in soluble rocks. Recently, however, the importance of *equilibrium topography* has become apparent as quantitative studies have been made. For example, Strahler (1950) points out that in the mature and later stages of landform evolution the topography of areas of homogeneous rock is typified by a monotonous topography of similarly shaped hills with similar relief, and that all areas have equally well-developed drainage. This equilibrium hill profile would be determined by the physical properties of the rock (solubility, grain size, fracture pattern, etc.), the local rate of erosion, and climate. All parts of the topography (streambeds, hillslopes, and hilltops) would be downwasting at essentially the same rate. This topography is essentially *static* in that there is little shifting of drainage, migration of divides, or changes of hillform during erosion, although relief may decrease or increase as erosion continues.

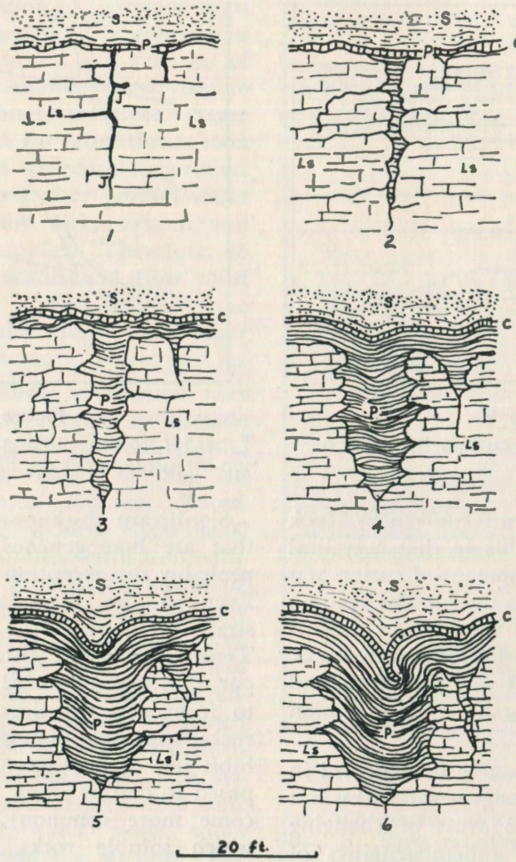

Figure 8

The development of cutters, (Hook, 1915).

S = Soil
Ls = Limestone
C = Clay seam
J = Jointing
P = Phosphate

Because it is an equilibrium topography all disturbances in the topography caused by rapid increase or decrease of rates of erosion should be short-lived and should rapidly reach equilibrium. It also means the direct effects of any initial conditions impressed upon the topography (for example an initial peneplaned surface) would be rapidly eliminated from the topography.

The stage of youth occupies only a short period in even an ideal cycle (less than 5% according to Johnson, 1932), and we should expect that most of the present topography on homogeneous rocks should approximate a static equilibrium topography. Hack (1960) maintains that very little or none of the topography of the eastern United States bears any imprint of cyclical tectonics and, correspondingly, is an equilibrium topography. Equilibrium topography would be continuously present under condtions of constant or slowly varying rates of uplift.

Let us now consider cutters in the framework of the concept that in areas of homogeneous rock the landforms are not changing in form or quality during erosion of the land, and the topographic elements are essentially time-independent. An early concept

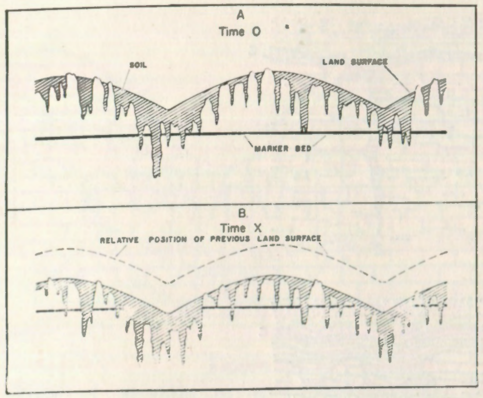

Figure 9
Vertical migration of cutters by erosion.

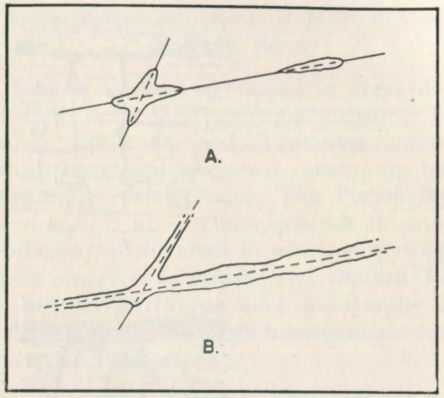

Figure 10
Types of solution along joints; A). Random, non-integrated pockets, B). Continuous channels.

of the development of cutters given by Hook (1915) (fig. 8) is notable in that a cyclical or unidirectional development of cutters was imagined. However, it is the present opinion that such a waxing and waning of cutters is not to be expected in homogeneous rocks, but rather, the amount of solution should remain essentially constant during erosion (fig. 9).

Caves and sinks are not to be expected in extensive areas of homogeneous rocks, but rather are confined to areas of changing topography caused by inhomogeneities of geology. Cutters, on the other hand, are the only karst landform present in areas of homogeneous soluble rock, although they are also present in areas of diverse geology.

In equilibrium topography on homogeneous rocks essentially no drainage gradients exist between a topographic point and any lower topographic point passing through bedrock which are greater than a gradient between the same upper elevation and an equal lower elevation by following a surface route only. Hence subterranean drainage capture cannot be expected. It has already been shown from studies of limestone terrains that are nearly homogeneous that solution decreases regularly with depth, and cutters are not roofed over in homogeneous strata. Even at greater depths there does not appear any reason that one should encounter abnormal cavities or caverns in homogeneous soluble rock.

Significant thicknesses of limestone strata that are homogeneous, or nearly so, are probably the exception rather than the rule. Even in the most nearly homogeneous strata, for example the Salem limestone and Tennessee marble, lithologic variations occur that permit small cavernous openings to form. To the degree to which soluble rocks approach homogeneity they will exhibit few or no cavernous openings (as opposed to cutters along joints, which will become more common). To the degree to which soluble rocks, or combinations of soluble rocks and non-soluble rocks, approach greater inhomogeneity caverns and sinks will form.

Caverns

Caves (caverns) are here defined as enclosed bedrock channels, generally of horizontal extension. All linear openings, from anastomoses to large cavern chambers, are therefore considered to be one continuum.

Cavernous passages result from solution of the bedrock by through-flowing ground water. Some authors have maintained that unintegrated, pocket-like openings often form in limestone by solution, especially in the saturated zone (Davies, 1958, p. 27). Figure 10 demonstrates the difference between an integrated passage and non-integrated pockets. An integrated passage is continuous, although it may enlarge or contract in dimensions. A non-integrated passage is fragmentary, or discontinuous.

Any continued solution of limestone must be accompanied by a relatively continuous addition of fresh solutions and removal of saturated solutions. Ground water with no appreciable flow cannot, therefore, create openings underground, for saturation is soon achieved, and, even if we postulate production of acid by bacteria after ground water is introduced into the rock, oxygen and food must still be supplied. Therefore, to dissolve limestone, a continued flow must be maintained, and, correspondingly, the openings through which the ground water flows must be continuous.

Evidence for the above conclusion may be cited from observation. Reference is made to a quarry near Hershey, Pennsylvania which encountered a 6-inch opening in limestone at a depth of 400 feet (Foose, 1953). This opening yielded 8,000-10,000 gallons of water per minute, but is well within the ground water zone which would be characterized by some as containing only primitive, nonintegrated openings. These, by definition, could not yield more than their capacity of water. The opening encountered must be part of an integrated cavern system, although it probably consists of generally very small openings. This system of openings was able to produce a cone of depression in the water table when water was pumped from the opening. The openings found in the limestone in Pennsylvania are not large when compared to a enterable cave, but are as integrated as their larger brethren.

Because caves must have, or have had, a circulation of ground water through them in order to be formed, they must have or have had a point or points connecting the cave to the surface from which the ground water entering the cave originated, and a corresponding discharge point or points through which the ground water exited from the cave to the surface. One possible exception to this rule are caves in porous limestone. In such a situation, dispersed flow of water through the rock might concentrate in discrete openings, giving no original connection between the cave and the surface. The determination of the entrance and exit points for ground water, and the determination of the general pattern of cir-

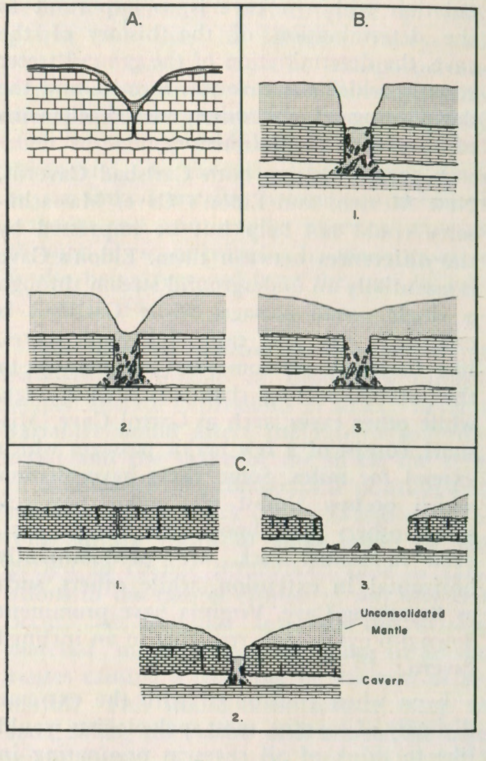

Figure 11
Sink morphology in horizontal rocks: A). Surface topography mirrored on the bedrock surface, B). Sinks resulting from cavern collapse, passing from stages 1 to 3 as surface mantle creep tends to obliterate surface expression of the collapse, and C). Concept advanced in this paper, that surface expression of a sink is not necessarily mirrored in the bedrock. Sink passes through a series of stages 1 to 3, with a much greater length of time in the first stage than in later stages.

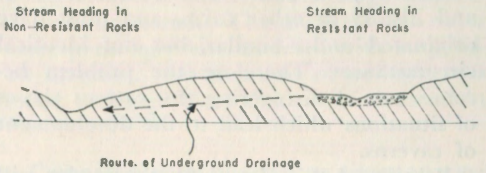

Figure 12
Capture by solution piracy of a stream heading in resistant rocks by a lower stream heading in weakly resistant rocks.

culation within a cave is as important as the determination of the history of the cave, the determination of the ground water zone in which the cave was formed, and the deciphering of the other critical elements of cavern form and history.

A person visiting both Carlsbad Caverns, New Mexico, and Eldon's Cave, Massachusetts, could not help but be impressed by the differences between them. Eldon's Cave is essentially an underground stream through a single, small passage, while Carlsbad is essentially dry and capacious. Some caves, like Carlsbad, are somewhat comparable to the interconnecting chambers of a sponge, while other caves, such as Carrol Cave, Missouri, consist of a few major passages which extend for miles. Some caves have streams which occupy almost all major passages, while others are almost dry. Some caves, even in folded rock, are predominantly horizontal in extension, while others, such as Breathing Cave, Virginia have prominent down-dip extensions, resulting in an inclined cavern.

Even when confronted with the extreme diversity of caverns, most speleologists would like to think of all caves as originating in essentially the same manner (Davies, 1960, p. 29; Bretz, 1956; and Woodward, 1961). I do not believe that it is possible to discover one universally applicable origin of caves unless one speaks in vaguest and most inconsequential terms. Variations in the features and histories of caves must, to a large extent, reflect corresponding variations in the processes operating to form the caves. This approach has been anticipated by Halliday (1960, p. 28) and Lange (1960, p. 29). However, this does not mean that we are forced into an alternative that each cave originates under *completely* unique situations. Many caves are very similar in features and history to other caves, and must have originated under similar, but not identical circumstances. Therefore the problem becomes one of discovering the various classes of situations which lead to the development of caverns.

Because of this viewpoint, the reader will not find in the following pages a complete description of all circumstances leading to the development of caves, but only those situations which the author believes at present to be responsible for the developmnet of some types of caverns. There are, of course, other types of caverns which have not been studied sufficiently to be encompassed within this paper. Although all caverns develop under at least partially unique circumstances, there are a number of principles that are true of all caves. One of these has already been discussed, that no caves can be expected in large areas of homogeneous rock.

The type of ground water flow through discrete bedrock channels which leads to the development of cavernous passages has already been noted as lacking in topography on homogeneous rock. It is my opinion that two types of geologic-topographic situations may lead to the development of caves. The first is the situation where a drainage gradient through soluble rock exists between an upper topographic point at a source of drainage to a lower topographic point which is greater than the topographic gradient. The other situation is where artesian circulation is promoted through soluble rock.

Sinks

Sinks, or sinkholes, may be defined as enclosed topographic depressions resulting directly or indirectly from solution by ground water. Sinks are generally found on otherwise gentle to flat topography, for on significant topographic slopes they are highly unstable. Sinks are collection areas for the diversion of surface water underground. They may have formed primarily from this function, or, like some sinks formed by collapse of cavern roofs, this may be a secondary attribute.

Sinks presuppose caverns, but the reverse is not necessarily true. In areas of almost complete drainage underground through sinks, they are the sole source of ground water responsible for the caverns. However, many caves receive ground water flow through sources other than sinks, and such sinks as may be above them result either from cavern passage collapse or a secondary diversion of drainage from the surface into the cavern.

The simplest and the most prevalent concept of the forms of sinks is that the bedrock surface essentially mimics the surface topography; the bedrock surface is assigned great funnel-shaped depressions (fig. 11A), which form by solution as ground water moves down into cavern passages. Although I have examined many sinks, and quarries in sink areas, I have not seen this type of bedrock depression. The quarries in sink areas , for example near Mitchell, Indiana, show the bedrock surface to be essentially planar. I have seen sinks with nearly vertical bedrock walls, obviously formed by collapse into underlying cavern passages, but these are not of a solution origin.

Coleman and Balchin (1959) conceive of sinks in a different way, envisioning cavern collapse to explain the surface depressions. Their reasoning appears to be as follows: solution depressions in limestone are admitted to exist. They correctly observe, however, that solution, acting downward but slowly, is much slower than the rate of mantle creep, and therefore bedrock solution depressions (such as cutters) have no surface expression, solutional lowering being counterbalanced by inward soil creep. Rather, cavern roof collapse acts to make an initial depression (fig. 11B). After initial collapse of the cavern roof, mantle creep gradually forms a shallower, broader depression, which is finally eradicated by further inward creep, at which time no further surface expression exists above the collapse. However, the quarries in karst areas, mentioned above, show no cave openings large enough to promote roof collapse, even though the quarries are in areas of abundant sinks. In such areas the subterranean drainage is through small joint and bedding openings. In addition, these quarries show no evidence of former collapse of bedrock. Also, passages in many caves pass shallowly beneath surface sinks, and show no collapse. Therefore, although collapse sinks most certainly exist, they do not account for most of the sinks of the sink plains.

The two hypotheses of sink morphology given above neglect process of removal of the unconsolidated mantle material through bedrock openings. Malott (1922, p. 197) points out that in solution valleys all weathered material must be transported underground to surface streams. Where streams disappear underground, they must also carry their suspended material and bed load underground. In caves it is possible at times to see sinks from the bottom up; debris is actively and continually entering caverns through these openings. In many cases it is also being removed from the cave by subterranean streams.

Soil material washed into underground passages is capable of producing funnel-shaped surface depressions of sinks. Several variables would affect the rate of soil material and the shape of the sinks: the type of overburden, the topographic position and nature of the bedrock channels, and climate. The surface forms of sinks might be expected to change throughout time as variations in the rate of erosion of soil material occur. At irregular intervals the mantle material might effectively plug drainage routes causing a lessening of the rate of removal. This would cause mantle creep to start filling the basin, and a perched lake might form. Alternately, these seals would break, and large quantities of material would suddenly wash underground, leaving fresh, steep sinks. Such sudden releases of material underground would be a sufficient cause for the rapid disappearance of lakes or appearance of sinks which are often attributed to cavern collapse. An instructive example of this occurred near Hershey, Pennsylvania (Foose, 1953) where the stone company (previously cited) artificially lowered the water table in a limestone terrain. The resulting cone of depression increased the efficiency of underground drainage. A surface stream was immediately diverted to a subterranean course and a large number of sinks appeared. The underground water correspondingly was muddied with suspended material. No mention of collapse was made, although these sinks were observed in formation.

Compared with quarries or limestone exposures in areas of cutters, those quarries in sink areas in flat lying rocks show contrasting morphology. No cutters were seen in these quarries, rather the mantle-bedrock contact is essentially planar. Surprisingly, perhaps, the total amount of bedrock solution in the sink areas appears to be much less than in the areas of cutters. The solution forms that were seen were stratigraphically determined cavernous channels, most of them rather small (3-4 inches wide and a few feet in height), but occasional larger cave passages are to be found.

My conception of the sink in essentially flat-lying rocks is shown in figure 11C. The flatness of the bedrock surface is caused by the differential resistance of the rock units. The unit immediately below the mantle and above the cavern passage is an especially resistant unit. The less resistant rock overlying this unit has been removed by erosion, and downward weathering has almost halted at the resistant unit. Because the resistant unit has persisted, it has assumed a position as an upland structural plain. Malott (1922, p. 194) points out that the great sink areas in Indiana are found on structural plains. It is assumed that more soluble layers are present beneath the resistant unit. Streams or other concentrations of water on such an upland must exist as perched water bodies by virtue of the very low stream gradients on such a structural plain and its position high above the lowest topographic points of surrounding areas. Hence water on the uplands, where possible, is diverted downwards into the more soluble layer; this results in the formation of an underground drainage network.

The openings through resistant bedrock are small compared to subterranean passageways, and generally become significantly large only at wide intervals, for ground water will concentrate upon enlarging a few larger openings rather than enlarging all joints equally. Hence removal of soil material will occur only at wide intervals, producing the pitted sink plain.

As solution continues, the bedrock channels enlarge, and collapse may occur above the largest cave passages. Such occurrences are relatively rare, for solution must effectively quarry out the collapse block, for rock spans can generally bridge large widths. Figure 11C-2, 3 shows later stages in the evolution of some sinks, where by collapse they become gulfs.

To this point I have discussed only sinks in heterogeneous, nearly flat-lying rocks where the sinks are the sole contribution of ground water to the caves, as is the case for the great sink plains of the eastern United States. The second set of conditions is where the sinks are secondary features dependent upon the presence of cave passages in which the main sources of ground water are from areas other than sink plains. In such cases sinks may form either by collapse of cavern roof or by diversion of surface drainage into the cavern system. Such sinks need not have lithologically determined morphology, and should be relatively rarer than those in the sink areas of flat-lying rocks. Such sinks may form above pre-existing caverns in both horizontal and tilted rocks.

Karst Forms of Areas of Nonhomogeneous Geology

Caves and sinks seem to be most common in areas of great inhomogeneities of geology. Many large caves are found at or near the escarpment of the Cumberland Plateau, which is upheld by a resistant sandstone series, and the caves are formed in limestone layers underlying the sandstone. Wind Cave, South Dakota is found in a limestone member of a series of shales, sandstones, and limestone forming the flanks of the Black Hills Dome. The caves of southern Indiana are found in the areas of rough and varied topography developed on shales, sandstones, and limestones. The caves of the Mammoth Cave region are characteristically associated with the escarpment of the resistant Cypress sandstone, which contrasts greatly in geology with the soluble limestone beneath that contains the caves. Carlsbad Caverns are developed in a variable reef facies, and the effect of the variations on the topography of the area is quite noticeable. In the Appalachians caves of large size are found where the stratigraphy has the greatest variation. Such examples might be extended almost without limit.

In contrast to the rather static conditions prevailing in areas of homogeneous rock,

the topography in areas of great diversity of geology is generally in a state of flux. Divides may be actively migrating; drainage conditions, topographic slopes and profiles, and stream gradients may vary greatly from one location to another; certain areas may have streams "perched" with respect to nearby areas, and stream capture may be common. Such conditions, which are the result of the influences of various lithologies and structures upon the topography, are often conducive to the development of caves and sinks where soluble rocks are present.

Hack (1960), and earlier Gilbert (1877), have shown that pediments are a common feature where drainage from resistant rocks passes onto less resistant rocks. Hack treats as an example an area in Virginia where streams from collection areas in resistant rocks flow onto the less resistant rocks of the Shenandoah Valley. The stream gradients in the resistant rocks are greater than those in the non-resistant rocks, for the non-resistant rocks are more easily eroded. Where the drainage from the resistant rocks enters the area of lower relief of the weaker rocks, a transitional phase of lateral planation, or terrace and flood plain development, is present, and is present only in such areas. Lateral planation occurs because bedload, composed of debris derived from resistant rocks, is more difficult to erode than the weak carbonates and shales. Streams which are developed entirely on less resistant rocks will tend to capture nearby streams which derive drainage from the resistant rocks by virtue of the lower altitude of the former. Such a capture would result in the abrupt lowering of the upper course of the stream heading in resistant rocks, and aggradation of the lower streams, the former resulting in the formation of dissected terraces.

If these weakly resistant rocks are also soluble, the possibility of underground drainage capture is likely. A case of stream capture underground is shown in figure 12. In constructing this figure it is assumed that the two streams flow essentially perpendicular to the plane of the section at this point. The stream headed in resistant rocks, by virtue of being "perched" with

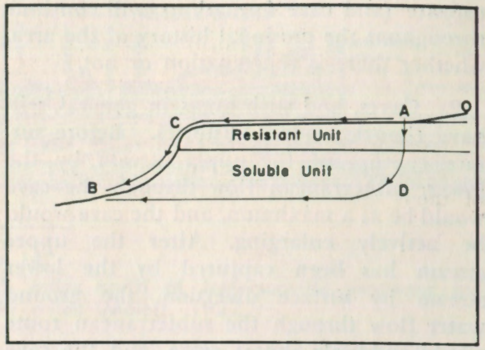

Figure 13

Two possible paths for drainage in a given geo-topographic situation. Path ADB is more stable than path ACB and will replace the surface path as suitable solution channels develop.

respect to the stream headed in the weaker rocks, will tend to loose water to the adjacent lower stream, and the underground route, being more direct, will be favored. It is notable that, when discussing the above type of surface stream capture, Hack used as an example the incipient capture of North River by Mossy Creek. On the topographic map of this area (Parnassus, Virginia Quadrangle) a series of sinks connects the two streams near their closest approach, indicating that limited subsurface drainage capture beneath the gravel cover of North River has probably already occurred.

In the above case the geologic situation was the presence of adjacent areas of resistant and weak, soluble rock, and the resulting topographic situation was the presence of high gradient streams where the drainage heads in the resistant rocks, and lower gradient streams which are developed entirely on the less resistant rocks. This situation promotes underground capture of the higher gradient streams by the lower.

The above case illustrates some principles that can be extended to the study of many types of caves:

1. No regional uplift of the land, nor tectonically influenced rejuvenation of streams, is called upon as a direct causal agent for cave development. The lateral planation and corresponding sporatic stream

capture (and cave formation) will continue throughout the erosional history of the area, whether there is rejuvenation or not.

2. Caves, and such areas in general, will have time-dependent histories. Before surface capture of the upper stream by the lower, subterranean flow through the cave would be at a maximum, and the cave would be actively enlarging. After the upper stream has been captured by the lower stream by surface diversion, the ground water flow through the subterranean route would suddenly almost cease, and the cave would no longer be enlarging. This time-varying history is in direct contrast to the rather static conditions in homogeneous rock areas.

3. Residual landforms are present in the terraces left after surface stream capture, and the dry cave similarly created; both are remnants of processes no longer operating. In homogeneous rock areas there are no such remnants, and in the present case these remnant landforms are not the result of tectonic uplift.

4. Caves have a morphology that is a function of the situation which has caused their development: the gradient of the cave in the above example will be well defined, and would be from the upper to the lower stream (or, after subsequent surface capture, from an upper dissected terrace to a lower stream or filled stream); the levels within the bedrock at which the caves are found in the above case will correlate with terrace levels on the surface; such a cave may be expected to be a simple and linear underground stream passage that is nearly horizontal and structure-crossing; similarly, the cave should be expected to be of shallow phreatic origin; this type of cave might have wall flutings and cave deposits of a pebble conglomerate derived from the material of the upper stream bed load and terraces; one might expect a paucity of sinks above such a cave, for it does not derive its through-flowing water primarily from sinks, although secondary and collapse sinks may be found directly above cave passages; and, because of the sudden decrease in drainage through these caves if surface capture of the upper stream by the lower occurs, a partial aggradation of the cave stream and complementary filling of passages may be present.

Deike (1960) has treated a strikingly different type of cave (Breathing Cave, Virginia) in folded rock. Here a soluble limestone is sandwiched between two insoluble and impermeable units. The topography is developed with respect to the geology such that an upper collection area on the limestone is on the flank of Jack Mountain, which is supported by the resistant rocks above and below the limestone, and a lower ground water exit point is present at a lower altitude where the strata are bent up into a sharp anticline. At an earlier time when the surface of the land was higher, this situation promoted artesian flow through the limestone from the upper area to the lower. Deike demonstrates that such "deep phreatic" artesian flow has resulted in the development of Breathing, and perhaps Butler caves in Virginia (Deike, 1960).

Cavern characteristics apparently associated with this type of development by artesian flow are: passages which follow the dip downwards without diminution; two or more joint systems almost equally dissolved, and a wide zone of solution, thus giving a maze-like pattern to the cave; generally fine-grained cavern sediments; and subterranean streams that are clearly secondary. This type of cave most closely approaches Bretz's (1942) conception of cave morphology.

In this case there is also a time-dependent history of the cave, but in this instance there are two distinct epochs of conditions: the deep phreatic flow of ground water through the cave under artesian pressure, and a later period of subterranean stream invasion after erosion lowers the siphon level of the standing water below that of the cave.

Again, no rejuvenation or particular "stage" of landform development is required in order to account for the presence of the caves. Although Deike shows a previous, higher land surface of low relief, the type of flow leading to the development of the caves would have occured even if the previous relief at the time of artesian flow (when the land surface was at a higher level relative

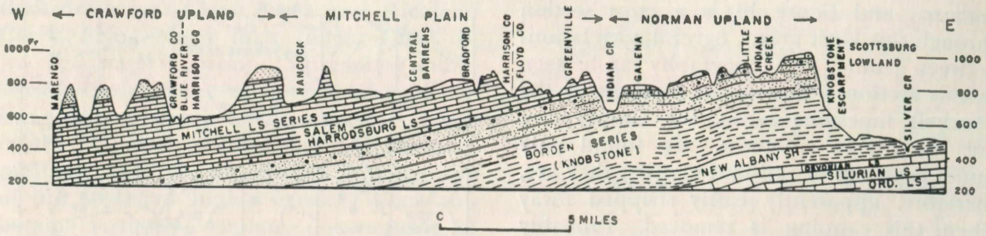

Figure 14
Geologic and topographic profile from about 5 miles north of Jeffersonville westward to Marengo, a distance of 33 miles (Malott, 1922.)

to the structure) was of similar relief as that at present.

Woodward (1961, p. 50) asserts that caves must come about as a relatively sudden event. This may be often true in the case treated previously, but clearly in this case the conditions for artesian flow may persist for millions of years. Thus, depending upon the geologic and topographic circumstances, caves may come about as the result of intensive processes acting for short periods, or as the result of lesser agencies acting for longer times.

Let us now consider the origin of some caves in nearly horizontal rocks. One of the most common situations conducive to the development is that shown in figure 13. The geological situation is that of a soluble unit overlain by a relatively resistant unit. The profile is constructed along a surface drainage route OACB. The stream profile clearly shows the effects of this contrast of geology. The sections of the stream to the right of point A and to the left of point B may be taken to be "normal" stream gradients on uniform rock. However, as the stream has cut down it has encountered the resistant unit which it is unable to erode by usual means at a rate equal to that of the erosion of the normal sections of the stream. Therefore the stream follows along essentially on the top of the resistant unit to the point C. Between point C and B is the nickpoint characteristic of such situations. This develops because it is the only effective way to remove the resistant unit at a rate equivalent to the rate of lowering of the downstream portions. This is accomplished by the intense forces of abrasion and undercutting acting to remove the resistant unit in such a steepening of slope. Therefore the resistant unit is effectively removed by upstream advance of the nickpoint, which is, incidently, not eradicated by further erosion as Davis (1930) imagined. Interstream areas are eroded by a very similar type of escarpment retreat.

Now consider the two points A and B. Point A is anywhere on the flat plain on the resistant unit, and B is a point on the lower area. Drainage will tend to move from A to B by either of two routes, the surface route ACB, and the subterranean route ADB. ADB is the more stable configuration, and will tend to replace ACB, for the subterranean gradient at point A is greater than the gradient of the surface route. Hence sinking creeks and a sinkhole plain, with a corresponding cavern system, will tend to form. Major streams along the upland plain will be diverted into one or more swallow holes, and the drainage system will be fragmented into enclosed sink basins. Sinking streams will form the largest subterranean channels, and lesser sinks, smaller channels, all of which would be organized into a crudely dendritic subterranean drainage system. In cases where the resistant unit over a soluble unit is very insoluble and impermeable, no significant solution can occur beneath such a cap except near the margins of the escarpment, and such features as domepits can be sculptured by vertically moving water from marginal sinks (Pohl, 1955).

Let us consider a number of cases where the above geologic situation is present. Malott (1922) described the karst plains of

Indiana, and figure 14 is a cross section through this karst area. Several correlations between geology and topography can be seen in this section. The upper portions of the Mitchell limestone series, the Gasper and Ste. Genevieve sub-units, are found only under cappings of resistant clastics; they are, therefore, apparently easily stripped away when this capping is removed. Probably because of their homogeneous and weakly resistant properties, correspondingly small outcrops and steep topographic expressions, these limestones, while very soluble, have little subterranean drainage developed in them. The middle of the Mitchell series, which is the top of the St. Louis sub-unit, is a resistant unit of cherty limestone. This unit forms extensive area of outcrop, in an upland structural plain closely following the dip of the strata. For example, the area around Central Barrens is such a structural plain developed upon the cherty unit. It is upon these chert-capped plains that the greatest development of sinks is found, with large areas drained only by sinks. A good portion of the subterranean channels lies just beneath the chert horizon, and most of the well-developed sink topography lies just above the same unit. Most of this underground drainage exits to the west, downdip, in the deep valleys of the Crawford Upland as springs and resurgences. The western-most portions of some underground channels are apparently under artesian conditions, because the channels closely follow the dip of the strata downward to the west, and rise as vertical resurgences, or "negative sinks". The caves further to the east, under the karst plains, are well above the level of the main streams, and are free surface, graded, underground stream passages. The lower part of the St. Louis, the Salem, and the Harrodsburg limestones are of about equivalent resistance and solubility; in these nearly homogeneous but soluble rocks little underground drainage is present except that in cutters. It will be remembered that the Salem limestone is characterized by well-developed cutters. The surface-drained areas on these units extend from Bradford to Indian Creek (fig. 14), and the surface drainage from these areas often passes westward into underground streams where the capping of the cherty unit is present.

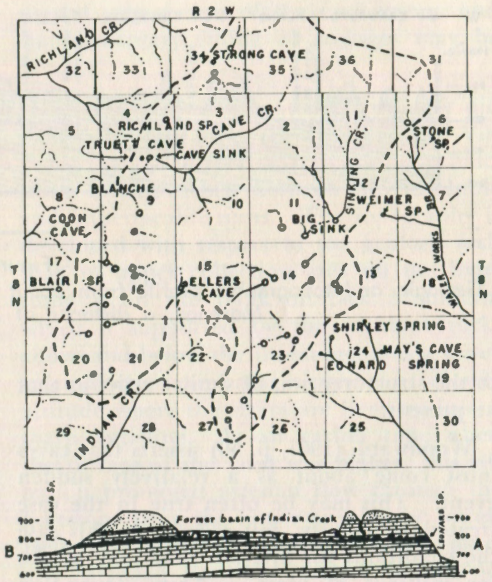

Figure 15
Present drainage conditions near the headwaters of Indian Creek southwest of Bloomington, Monroe County, Indiana. Note that the former Indian Creek basin is 100-150 feet higher that the surrounding streams, and also that this basin floor is developed at approximately the level of the resistant unit of the St. Louis limestone, (Malott, 1922). Dashed line bounds area of subterranean drainage.

The subterranean capture of the headwaters of Indian Creek, near Bloomington, Indiana, affords a striking contrast between the conventional theories of karst interpretation and the present view. Figure 15 shows the present conditions at the former basin of the headwater portions of Indian Creek. This drainage basin is now drained entirely underground, and the subterranean waters exit into the lower 'intrenched" tributaries of Clear Creek and Richland Creek.

According to the conventional cyclical interpretation of the underground capture, at the time of the late Tertiary peneplanation the entire area was drained by surface streams, as illustrated in figure 16. After uplift of the area, Clear Creek, Richland Creek, and Indian Creek began to intrench their drainage. Clear Creek and Richland Creek, being larger than Indian Creek in its headwater portion, intrenched their stream channels by nickpoint advance before Indian Creek could do so. These creeks, by virtue of their lower topographic position, then began subterranean capture of the waters of Indian Creek. When this became virtually complete, downwasting of the Indian Creek basin was effectively halted, and it stands as such a structural sink plain today.

As can be noted in the geologic cross section of figure 15, however, the perched basin of Indian Creek is stratigraphically controlled at the horizon of the resistant cherty unit. This gives a basis for an alternative explanation in terms of the geologic effects of topography.

We can consider figure 15 to represent the drainage conditions at some indeterminate point in the past when no subterranean drainage was present (before the exposure of significant portions of the middle of the Mitchell limestone). However, unlike the previous interpretation, we could assume the topography at this earlier time to be essentially as great in relief as it is at present, for an earlier peneplanation is not a prerequisite. As erosion continued the major streams reached the resistant cherty unit of the Mitchell, the larger creeks, Richland and Clear, reaching it first. Because of the resistance of the unit, erosion was effectively halted for some time beneath each stream, resulting in very low gradient stream segments and partially stripped plains on the cherty unit. In order to accomplish further erosion the streams had to establish a steeper gradient through the resistant unit. This was accomplished largely by the migration of a stratigraphically-determined nickpoint upstream. Richland and Clear Creeks, because of their greater size, accomplished this more rapidly and with a smaller flat transition zone on the resistant rock than did Indian Creek. The migration of the nickpoint upstream on these creeks left Indian Creek perched with respect to the lower creeks, subterranean capture of Indian Creek resulted, and an upland karst plain developed at the level of the resistant cherty unit.

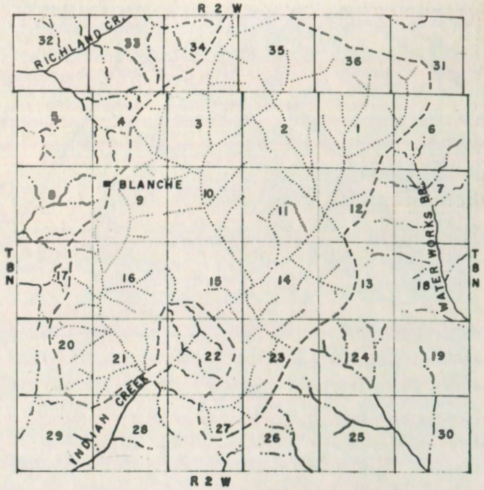

Figure 16
Restoration of surface drainage of the headwater portion of Indian Creek. Such a stage of development is assigned by Malott to the close of the late Tertiary peneplanation. Little underground drainage is assumed to exist at this stage, (Malott, 1922).

In the above interpretation no specific rejuvenation of drainage is called for as a casual agent for the development of underground drainage; the development and migration of the nickpoint is caused by the presence of the resistant unit.

One more case from the karst area of Indiana will be considered (fig. 17). A striking gradation of topographic forms may be seen along this strip of land. In the state park are the headwater portions of Mill Creek, which in this zone is deeply incised beneath the level of the surrounding upland karst plain. This stream ends abruptly at the entrances to caves and at springs, all of which discharge water from underground drainage channels to the surface stream. This underground drainage derives from the extensive karst plain above the incised stream. Progressively from Mill Stream the "intensity" of sink drainage decreases, until, near the far end of the section, surface drainage predominates. Once again this variation in karst forms may be explained by recourse to the geologic controls. Figure 18 is a schematic geologic cross section from

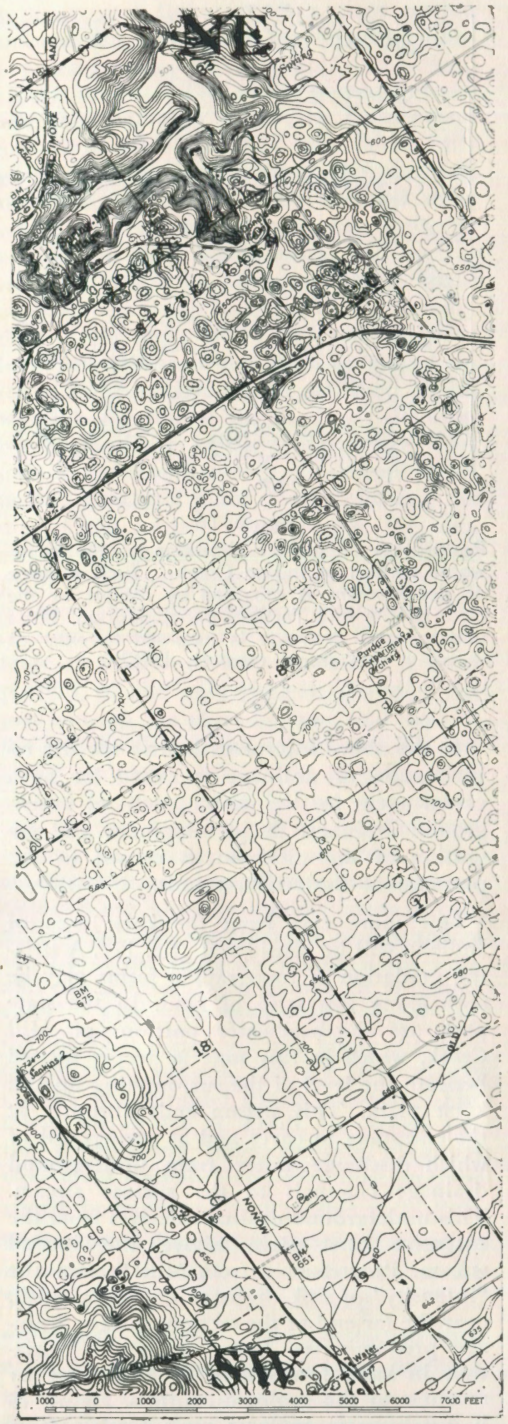

Figure 17
Topographic map, karst area, southern part of Lawrence County, Indiana (U. S. Geol. Surv. Mitchell Quadrangle).

the incised stream to the farthest portions of the strip along an imaginary underground stream passage.

The resistant unit is the cherty part of the Mitchell limestone series, the soluble unit comprises the limestone under the cherty unit, and the weak rocks are the limestones of the upper Mitchell. The sink plain is on top of the cherty unit, the upper Mitchell is in the generally non-sink, low hills to the southwest, and the lower limestone is at the surface only where the incised stream is present. Mill Creek is incised or intrenched because of the presence of the resistant unit, and is slowly eroding headward by nickpoint advance. Because of the intrenchment and the presence of soluble rock underneath the resistant unit, sink drainage is more stable than surface drainage on the upper structural plain. The extent of underground drainage is greatest near the incised stream, where underground gradients are greatest and where sinks and caves have been present for the greatest time, and the amount of underground drainage is least in areas far from this stream, where the underground drainage gradient is much less, and which have had the shortest exposure from beneath the overlying, more homogeneous limestones which are not particularly subject to sink drainage. The areas of intense karst drainage have all the characteristics of what has been termed "old age" karst by the proponents of the karst cycle, for in these areas cave roof collapse and corresponding cave destruction is prominent. However, if the former is true, we must also recognize that the areas farther from Mill Creek must be merely "mature" or even "youthful." But if all stages exist concurrently in adjoining areas this karst cycle must be quite different from the classical conception of the same. It is true, however, that an area of poorly developed sink topography may advance to more well-developed stages as the advance of the intrenched stream steepens the underground gradients, and solutional enlargement of these caverns continues. But the main objection to the term *karst cycle* is the inherent inferences of cyclical landform evolution, a concept which is not necessary in order to describe the origin of caves, or other karst forms.

Although cutters are the only karst feature of areas of homogeneous soluble rock, they are also present in areas of inhomogeneous rock. To the extent that the rock is inhomogeneous, the form and distribution of cutters will depart from the simple, upward expanding cutters of uniform soluble rock. Because cutters are an equilibrium landform in homogeneous soluble rock, wherever a large area of soluble rock is or becomes exposed, cutters will tend to develop, and caves and sinks will be unstable. Transitional forms between cutters and caves may often be found; their closeness to caves or other cutters is dependent upon the degree of inhomogeneity. For only slight lithologic variations, cutters will be prominent, but will have reentrants or overhangs (fig. 4). All areas of predominantly cutter type bedrock surface have some sinks in the surface topography. Such sinks can often be demonstrated to result from collapse of soil material into cutter channels whose original filling has been removed by abnormal amounts of underground drainage because of heavy rainfall. E. H. Walker (1956, p. 24-25) notes that such sinks are most common where the cutters are characterized by having overhangs of more resistant bedrock. In such cases most soil material is held by the horizontal projections and soil removal into the lower openings occurs only at wider intervals, and sinks are the result. If the resistant layer were more prominent, a normal sink plain would have formed, and if less prominent, normal cutters would be found.

In areas of nonhomogeneous rock the development of cutters may approach the "cyclical" evolution of Hook (fig. 8). Where limestone is overlain by impermeable shale, no cutters form, but as the cover is removed from the limestone cutters will start to form.

As another example, let us consider the karst forms developed on Silurian limestone escarpments in northeastern Iowa (fig. 19).

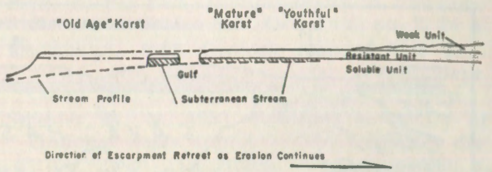

Figure 18
Elements of karst topography where a horizontal resistant unit overlies a soluble unit.

Here the karst forms, while still intermediate between cutters and normal sink plain caves, are closer to the latter. The Silurian dolomite, which overlies an impermeable shale, contains one prominent resistant bed, and the limestone beneath is nearly homogeneous. Where the resistant bed forms a cap over the solutionally enlarged joint in the underlying limestone, high, narrow caves are to be found, as in section B-B' (fig. 19). Above these caves are surface sinks where surface drainage passes vertically through isolated openings in the cap rock, carrying soil material with it, as at E. Some places where such openings through the cap are large and where the soil material brought in is continuously removed by underground streams, as at D, cave entrances will be found. Most often, however, where the resistant cap has been breached, as near cross-section A-A' and C-C', the underlying openings are clogged with soil and debris, and a cutter-like form will be found.

Because cutters are the equilibrium karst landform on homogeneous soluble rocks, where uniform limestones are exposed at the surface, cutters tend to form. As erosion strips away layers of rock such areas of uniform rock will shift. Thus cutters will wax and wane in response to the degree of homogeneity of limestone at the surface, and caves and sinks will be formed and eradicated in response to inhomogeneous geology. The effects of continuing erosion on the distribution of caves and cutters in Indiana is sketched in figure 14.

The rock units above the cherty unit of the Mitchell covered the areas to the east at one time, as shown in this figure, but have now been stripped off. Hence, at one time the type of topography now found in

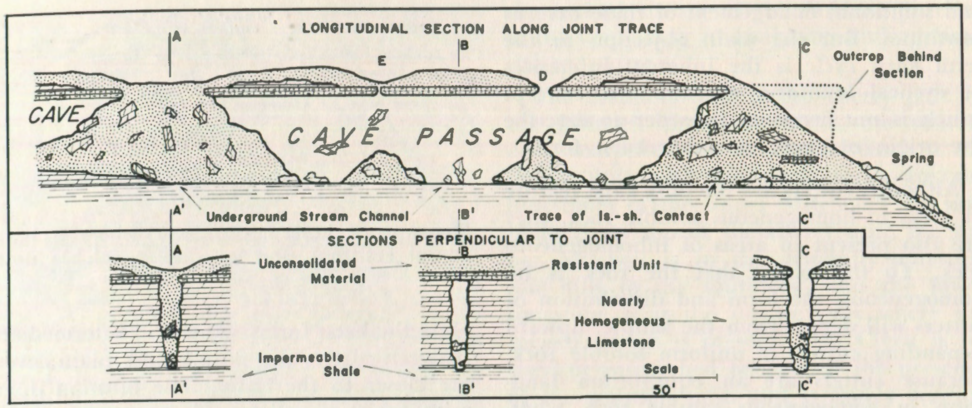

Figure 19

Karst landforms of the Silurian escarpment of northeastern Iowa.

the Crawford upland probably was present where the Mitchell plain is now. Specifically, the area of the present sink plain at one time had no sink and karst features. But when erosive processes had bared the cherty unit of the Mitchell, underground diversion of drainage became more stable than surface drainage which is the condition that now exists. Eventually the resistant layer supporting the sink plain will be removed and more homogeneous limestones underlying the cherty unit will be at the surface. Cutters will replace caves and sinks as the dominant karst form, and topography similar to that near Greenville, with surface drainage, will develop in the present areas of sink drainage. If we choose a point in the Crawford upland, say Marengo, the topographic changes as one progresses eastward from Marengo gives a preview of the successive changes of cave formation and destruction that will eventually prevail in the vicinity of Marengo. Thus by knowing the present relationships between geology and topography, and the resulting karst forms, it is often possible to extrapolate to future or past conditions.

ACKNOWLEDGMENTS—This paper is derived largely from a Senior Honors Essay completed at Yale University in May 1961. In addition to the referenced literature, my conclusions are based upon observations made during caving trips in the eastern United States throughout the past 3 years. In December, 1960, a 5-day trip was made through Indiana, Tennessee, and Virginia for observation of karst landforms.

Acknowledgment is given to Dr. John Sanders, my advisor, and to my companions in caving, especially Stewart Peck, and members of the Yale Speleological Society and the Quint City Grotto, both chapters of the National Speleological Society.

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Canga Caves in the Quadrilátero Ferrífero, Minas Gerais, Brazil

By George C. Simmons*

ABSTRACT—Unusual types of caves have been found in the Quadrilátero Ferrífero of Minas Gerais, Brazil. The caves occur where metamorphosed Precambrian sedimentary rocks have been eroded from beneath a resistant cover of Cenozoic canga. The caves can be divided into two classes, solution caves and corrasion caves, according to the kind of erosion responsible for their formation. This, in turn is dependent upon the lithology of the rocks, the solution caves being in itabirite, and the corrasion caves in phyllite and schist. Many solution caves are multichambered and have linear extents of over 100 meters. The corrasion caves are single chambered and are not known to have lateral dimensions greater than 15 meters.

Leucophosphite, a rare phosphate mineral formed by the action of organic solutions derived from bat guano with iron oxides, was discovered in one of the caves.

Unusual caves, believed unreported in speleologic literature, occur in the Quadrilátero Ferrífero[1] in the State of Minas Gerais, Brazil (fig. 1). The caves are formed under canga, a limonite-cemented, hematite-rich surface debris, some as the result of corrasion, some as the result of solution, and possibly others as a result of corrasion and solution combined.

The smaller, corrasion caves were noted by the writer during 1958 and 1959 while mapping the general geology and iron deposits of certain quadrangles for the Departamento Nacional da Produção Mineral and the U. S. Geological Survey in connection with the "Point Four" Program. Larger solution caves were discovered in 1960, and one was mapped as an example of the solution caves which occur in the Quadrilátero Ferrífero.

CANGA

Canga is the name for limonite-cemented, hematite-rich surface debris which mantles many slopes and some crests and valley bottoms. The debris is derived from hematite-bearing metamorphosed sedimentary rocks. Like all surficial debris, it moves downslope before cementation, and therefore overlies other rocks, many of which have a low iron content. Canga blankets range in thickness from a few millimeters to 30 meters but are most commonly between 1 and 2 meters thick. The blankets are known to extend over areas as large as 500 by 3,500 meters, and these are erosional remnants of blankets which once covered much larger areas. Many masses contain more than 60 percent iron and have been mined for iron ore. Canga is sufficiently strong to stand unsupported where underlying material has been eroded from below.

Canga typically is composed of detritus containing itabirite and compact hematite fragments with minor amounts of other constituents, cemented by limonite. Individual detrital fragments vary from silt to boulder size. The amount of limonite cement also

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1 "Quadrilátero Ferrífero" is the Portuguese term meaning "Iron Quadrangle" which is applied to the part of Minas Gerais that contains a large number of iron (hematite) deposits.

*Publication authorized by the Director, U. S. Geological Survey.

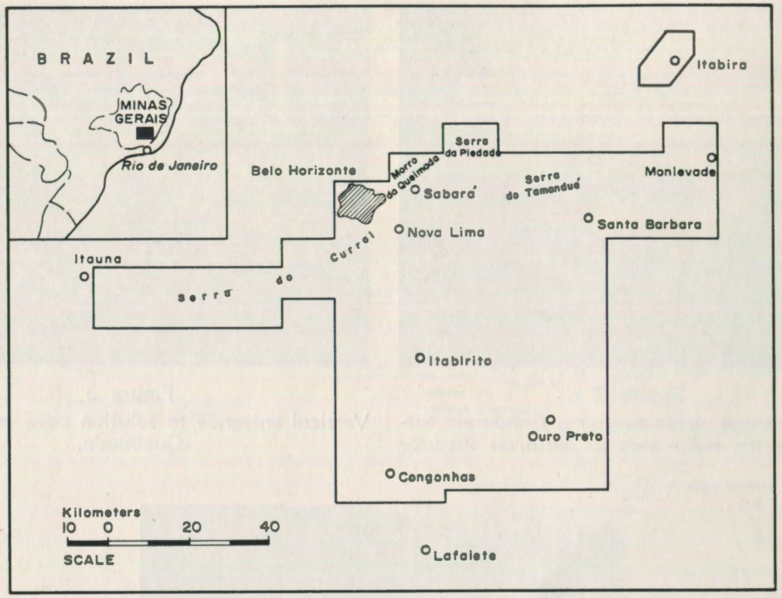

Figure 1
Index map showing the location of the Quadrilátero Ferrifero.

varies widely. At some places it completely fills the interstices but commonly, especially where the detrital fragments are large, the limonite incompletely fills the voids, forming a porous but relatively impermeable rock. Most freshly exposed canga has a yellowish-brown color and botryoidal structure on its upper surface. However, the botryoidal structure breaks down on exposure, and results in the rough surfaced, dark brownish-black rock which is most commonly observed.

During Tertiary and Quaternary time, canga formed as a result of the weathering of Precambrian rocks that contained abundant hematite. There are several hematite-bearing formations in the Quadrilátero Ferrífero, but hematite is most abundant in the Cauê itabirite, and it is canga derived from this formation with which all the caves thus far observed have been associated. The Cauê itabirite is the lower formation in the Itabira group of the Minas series. It is a relatively resistant formation which forms the ridge crests of most of the mountain ranges in the iron region. Extending down from many of these ridges are slopes covered by detritus from this formation. At many places this layer of detritus, especially in its upper part, is cemented by limonite to form canga. Elsewhere, canga is formed in place from itabirite by the precipitation of limonite and the hydration of hematite to limonite concurrent with the solution of quartz, hematite, and dolomite.

ITABIRITE

Though canga forms the ceiling and provides the protective cover which allows extensive solution to take place without intense corrasion, it is itabirite which forms the walls and bedrock floor of the solution caves. Itabirite is a metamorphosed sedimentary rock, much of which has a simple mineralogic composition, consisting of alternating laminae of specular hematite and granular quartz. The laminae are commonly 1 to 5 millimeters thick. Dolomite is locally abundant in itabirite and is concentrated in the quartz-rich layers. Where itabirite is dolomitic, magnetite or one of its oxidation products, maghemite or martite, occur with and sometimes to the exclusion of

Figure 2
Breached canga sheet overlying Gandarela formation on the north side of Serra da Piedade

Figure 3
Vertical entrance to solution cave in Morro da Queimada.

Figure 4
Lateral entrance to solution cave in the Serra do Tamanduá.

hematite. Tremolite has been identified in several places and an amphibole mineral with the radiating structure of tremolite but replaced by goethite is common and widespread. Kaolinite, chlorite, and talc are also common and are associated with dolomitic itabirite.

Exposures of dolomite in itabirite are rare owing to the ease with which carbonate minerals are dissolved under tropical weathering conditions. The former presence of dolomite is often determined by inference rather than direct observation. There are three guides to its former presence which may occur alone or in combination: associated minerals (magnetite, maghemite, martite, kaolinite, chlorite, and talc); porosity; and ocher. All these guides are known at the surface of itabirite which, when explored at depth by mine workings and drill holtes, was found to contain dolomite.

In unweathered itabirite, magnetite occurs with dolomite and is absent where dolomite is not present; it is thought that its abundance in relation to hematite is related to the amount of dolomite in the rock.

Itabirite becomes porous by the solution of dolomite, quartz, and hematite. Where quartz and hematite have been dissolved itabirite breaks down readily, but where dolomite has been dissolved the rock holds its form. The cause for this is unknown. One

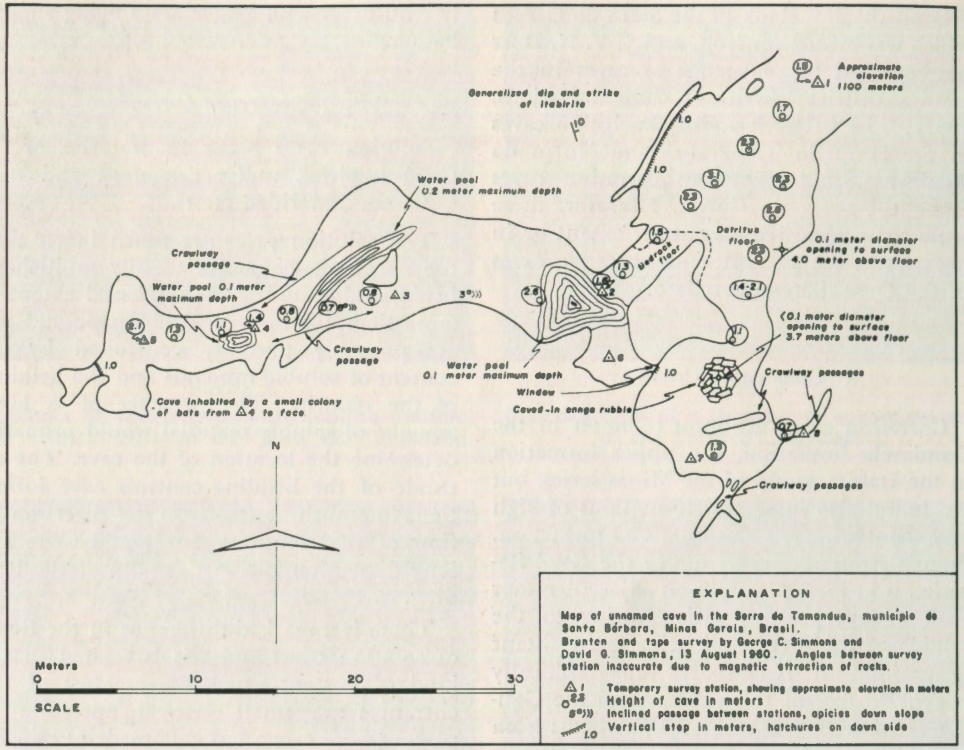

Figure 5
Map of solution cave in the Serra do Tamanduá

possibility is that secondary cement, goethite, forms more rapidly in dolomitic itabirite than in siliceous itabirites.

Ocher, a bright yellowish brown, earthy form of goethite, is a common surface feature of the highly dolomitic itabirite lenses of the Gandarela formation, but also occurs over dolomitic itabiritics in the Cauê itabirite. The reason for its forming from dolomitic itabirites rather than siliceous itabirites is unknown. Perhaps more pore space is developed in dolomitic itabirites so that the iron minerals can be more readily hydrated and there is more interstitial space in which goethite can be deposited.

Cave Distribution and Location

There has been no published study of the canga caves in the Quadrilátero Ferrífero. Most references to the caves are incidental notations in reports concerned with other geologic aspects of the iron region. No attempt has been made to count the number of caves, list their locations, or describe their accessibility.

The caves can occur almost anywhere that canga exists. A series of papers with geologic maps covering the iron region and showing the distribution of canga is currently in preparation. The first paper of the series has been published (Guild, 1957) and many of the other maps are available for study at the Departamento Nacional da Produção Mineral offices in Rio de Janeiro.

A corrasion cave named Casa de Pedra (Stone House) was reported by Guild (1957, p. 71) in the Congonhas district, and he has also reported the existence of caves in the Serra do Batateiro (written communication, 1961). C. H. Maxwell found a large solution cave near Fazenda da Alegria on the east flank of the Serra do Caraça which was mined for iron, and J. V. N. Dorr has reported the existence of caves in the Itabira district (written communications, 1961). The writer has visited solution caves in the Serra do Tamanduá and Morro da Queimada, and numerous corrasion caves along the Serra do Curral. However, there is no comprehensive cave study extant or in progress at the present time on canga caves in the Quadrilátero Ferrífero.

Corrasion Caves

Corrasion caves are most common in the Gandarela formation, the upper formation in the Itabira group of the Minas series, but are found also in several formations of high clay content. The Gandarela formation, though stratigraphically above the Cauê itabirite, is at most places in an adjacent rather than a superposed position owing to the folding of the rocks, and being less resistant to erosion, it is usually topographically lower. Because of this relationship the Gandarela formation is covered by detritus from the Cauê itabirite, and thus also by canga formed from this detritus.

Corrasion caves occur immediately under canga-supported valley rims (fig. 2). Once a stream has breached a layer of canga, erosion of soft underlying phyllites, schists and uncemented debris proceeds rapidly, forming a steep-sided valley. When the valleys are deep enough, weathered rock and uncemented debris slumps from under the canga cap, leaving small grottos. Once formed, the grottos may be further enlarged by the abraiding action of running water.

The caves, with few exceptions, are single chambers, the size of which is limited by the strength of the canga. A room of approximately 15 meters diameter seems to be the maximum size attainable without the collapse of canga, but it is rare to find rooms approaching this size. The chambers are usually twice as wide as they are long, and frequently are slightly constricted at the mouth. The larger caves are often 2 meters high or more, and there is a general correspondence between cave area and cave height, caves with larger areas being slightly higher.

Solution Caves

Solution caves occur in itabirite of the Cauê itabirite, under canga, at and near canga-covered ridge crests.

The solution caves are much larger than the corrasion caves, are usually multichambered, and some have a horizontal extent of more than 100 meters. Their shape and size are probably partially controlled by the content of soluble minerals and the bedding of the itabirite. The presence of a large amount of soluble material would primarily determine the location of the cave. The attitude of the bedding controls cave formation, inasmuch as it affects the direction of flow of the solutions. In several caves the long dimension is down the dip of the bedding.

There is a marked difference in the floors of solution caves and this is related to the kind of entrances they have. Most of the entrances are small vertical apertures at places where canga has collapsed into a void below (fig. 3). Where this is the only kind of opening, the floor of the cave is covered with the undissolved residues from the itabirite. As these residues are mostly hematite, magnetite, and maghemite, they have been mined in some places for iron ore. Where the caves have lateral entrances, a much less common occurrence (fig. 4), streams flow out of the caves during the rainy season, carry away most of the residues, and leave a bare rock floor.

A cave mapped in the Serra do Tamanduá (fig. 5) has many of the features common to solution caves in the Quadrilátero Ferrífero. The ceiling of the cave is nearly everywhere formed by canga, and the angular unconformity between the canga and the underlying itabirite is clearly exposed at the top of chamber walls and pillars (fig. 6).

The cave walls are much smoother than might be expected in a laminated rock of alternating composition, and the rounded form of the chambers is similar to that found in limestone caverns (fig. 7). The cave has a bare rock floor in its further reaches but the

Figure 6
Contact of canga (top) with itabirite (bottom) in solution cave in the Serra do Tamanduá.

Figure 7
Rounded form of a chamber in a solution cave in the Serra do Tamanduá.

Figure 8
Dirt-covered floor (foreground) and rock floor (background) in a solution cave in the Serra do Tamanduá.

Figure 9
Shallow pool in a solution cave in the Serra do Tamanduá.

floor is covered toward the entrance. The covering is composed of breakdown material from the overlying canga and insoluble residues (mostly maghemite) from the itabirite which have been concentrated toward the mouth of the cave by a small stream that flows out of the cave during the rainy season. The rock floor is smooth like the walls, and is almost free of detritus (fig. 8). Some depressions in the rock floor contain water throughout most of the year, forming pools (fig. 9). The longest of the pools is more than 10 meters long, but only 20 centimeters in maximum depth.

The approximate volumetric composition of the porous itabirite forming the walls and floor of the unnamed cave mapped in the Serra do Tamanduá (fig. 5) has been determined from grain counts on four thin sections and eight samples prepared in a Franz separator. It is:

Maghemite	40%
Leucophosphite	20%
Goethite	19%
Voids	12%
Magnetite	5%
Quartz	4%
Total	100%

By combining the percentages of leucophosphite, goethite, and voids it was determined that nearly 50 percent of the wallrock was dissoved. The leucophosphite, a rare secondary mineral discussed below has been deposited in pore spaces. Most of the goethite was also introduced, though a small amount was formed by the hydration of maghemite and magnetite in place.

By combining the percentages of magnetite and its oxidation product, maghemite, and as a small amount of the goethite was formed by the hydration of these minerals, the original magnetite content is calculated as having been more than 45 percent. No hematite was found in any of the samples.

High porosity, high magnetite, maghemite and martite content, and low hematite content are characteristics of weathered dolomitic itabirite. As the wallrock of the Tamanduá cave had a porosity of nearly 50 percent, a combined maghemite and magnetite content of more than 45 percent, and no hematite, there can be little doubt that the cave was formed in what was originally a highly dolomitic itabirite.

Leucophosphite

During the brief study of these caves little attention was paid to any of the biologic features other than to note that the larger solution caves served as the residences of small bat colonies. Guano, though present, was not recorded. This was unfortunate because in the mapped solution cave the guano was probably important in the formation of a rare mineral, leucophosphite, which fills much of the pore space in the cave wallrock.

Leucophosphite, approximately $K_2 (Fe, Al)_7 (PO_4)_4 (OH)_{11} {*} 6H_2O$, has been tentatively identified from its X-ray diffraction pattern by M. H. Falabella of the Brasil Comissão Energia Nuclear (written communication, 1961). The mineral is light yellowish brown and fills much of the pore space of the wallrock. It also occurs on the cave walls and floor as a thin veneer mixed with goethite. Leucophosphite was previously known from only two other localities: Ninghanboun Hills, Lake Weelhamby, Southwest Australia (Simpson, 1932) and Bomi Hill, Liberia (Axelrod, Carron, Milton, and Thayer, 1952). In Australia the formation of leucophosphite has been attributed to the action of solutions derived from bird guano on serpentine. In Africa the formation of leucophosphite has been attributed to the action of solutions derived from bat guano in caves in massive iron ore with iron oxides both in and outside the cave. The new occurrence in Brazil is very similar to the occurrence in Africa and it seems highly probable that the Brazil leucophosphite was formed in a similar manner by the action of organic solutions derived from bat guano with the iron oxides, maghemite, magnetite, and goethite.

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The Caves of Mont Hoyo, Eastern Congo Republic

by C. D. Ollier and J. F. Harrop

Abstract——Mont Hoyo is 25 miles southwest of Irimu in the Eastern Province of the Congo Republic. The caves are in Precambrian limestone and shale that have not been metamorphosed nor strongly folded. Several caves in Mont Hoyo contain large chambers and extensive passages. They are similar in form to caves in temperate areas. Cave fills of red clay, now mostly removed by stream action, and of guano are extensive. The caves in Mont Hoyo are developed on several levels. All contain evidence of origin under phreatic conditions, probably previous to the development of the fault escarpment bounding the Mont Hoyo horst.

Both limestones and caves are uncommon in Central Africa, and large, well developed caves such as those described here are extremely rare, and therefore worthy of description. Most detailed cave studies have been carried out in temperate and cold areas of Europe and North America, and it might be supposed that caves of equatorial regions would have different features. The Mont Hoyo caves are in equatorial forest only one and a half degrees north of the equator, and so provide interesting comparative data.

Mont Hoyo is situated about twenty five miles southwest of Irimu in the Eastern Province of the Congo Republic and reaches a height of 1450 metres. The caves referred to in this paper occur on the western side of the mountain at about 1100 metres. They were first discovered by a European, de San, in 1940. Three years later Ruscart (1951) carried out a scientific and economic survey of the Mont Hoyo area and although his report contains descriptions of several caves he does not put forward any explanation of their formation.

Mont Hoyo appears to be a horst bounded on its western side by a fault scarp. It is composed of Precambrian limestones and shales, usually thinly but occasionally massively bedded. For Precambrian rocks they show remarkable little distortion and only near the escarpment is there any conspicuous folding. Elsewhere the strata are horizontal or gently inclined. The escarpment runs irregularly in a general northeastern to southwestern direction in the vicinity of the Mont Hoyo Hotel where most of the caves occur. Near the hotel a number of rivers run off the escarpment forming a series of cascades where they flow over the outcropping bands of shale. Several of the caves are associated with rivers, but as will be explained later they were not produced directly by river activity. Figure 1 shows the position of most of the caves.

General Description of the Caves

Matupi — This is perhaps the best known of the Mont Hoyo caves and contains one very large chamber opening almost immediately onto the escarpment wall. Two series of passageways lead from it which, though not extensive, are complex due to many interconnections all at approximately the same level. From the plan of the cave (fig. 2), surveyed by compass and pacing, it is evident that joint control has played an important part in its formation and joint lines are clearly visible in the roofs of the passage (fig. 3). The rock is gently flexed within the cave but at the entrance, which is on the escarpment there is considerable folding (fig. 4). The cave floor is covered by extensive deposits of bat guano both in the narrow passages and the main chambers; some passages also show various amounts of clay fill. The cave is noted for its formations (speleothems), most of which occur in the main chamber, but these are now dead and the whole cave is waterless.

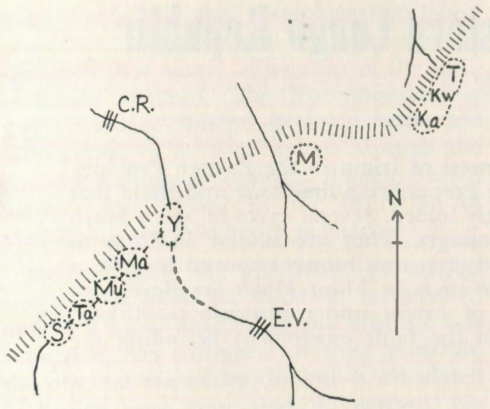

Figure 1
Sketch map showing the relative positions of the escarpment caves of Mont Hoyo.
T = Tsebahu. Kw = Kwama Kwama, K = Kabama, M = Matupi, Y = Yolohafiri, Ma = Mateta, Mu = Mutata, Ta = Talatala, S = Saga Saga, C.R. = Chutes Ruscart, E.V. = Escalier de Venus.

Tsebahu, Kwama Kwama and Kabamba — These names refer to three entrances all leading to the same cave system. They occur at approximately the same height and from each a rift passage descends to the main cave which extends for a distance of over a thousand metres between the two extreme entrances, Kabamba and Tsebahu. The main chamber of Kabamba has a high vaulted roof and is over thirty yards wide. From it the cave continues northwards passing a number of shafts which seem to occur at joint crossings. Small side passages also show a similar joint control and in detail follow closely the rectangular pattern of Matupi. In the lower part of the cave are high rifts which also follow joints (fig. 5). Water is absent from most of the cave though it is present at its lowest part and could be followed for a short distance to a point where it disappeared through a sand bank. Guano is plentiful but fill of other types is less common or obscured. Possibly on account of the amount of fresh guano present there is an abundance of insect life on the cave floor, chiefly cockroaches and beetles.

Yolohafiri and the Hotel caves — This group occurs at approximately the same level but in a lower part of the mountain than the other caves. Yolohafiri and one of the Hotel caves are occupied by active streams which could be followed from their entrance to the exits on the escarpment. They are, on the whole, large meandering caverns with several small branching passages leading off at various levels and in various directions. The main chambers are very irregular in outline due to the many collapses of flaggy limestone in which these caves are situated.

Andemoni and Lipanga — These two caves are situated near the summit of the mountain at a distance of about eight kilometres from the escarpment. Their entrances lie on the lower slopes of valleys and there seems little reason to doubt that they were formed by rivers taking underground routes. Water is present in both caves but only in Lipanga is there an active stream which, during its course inside the cave, falls in three small cascades. Lipanga contains a few formations (speleothems) including a stalactite grill and several rimstone pools.

An unnamed cave close to Lipanga also contains a small stream and differs from the others in being formed almost entirely along a single bedding plane.

Factors Involved in the Formation of the Caves

Joints and Veins — Jointing in the limestone is very common and, as can be seen from the Matupi cave plan, has exerted considerable effect on the layout of its passages. Well marked joint lines can usually be observed in the roofs of the passages for distances of over a hundred feet in some instances. Even though they were not studied in detail it is evident that the joint system is fairly complicated with well formed joints running in a variety of directions. Most of the joint planes are vertical but occasionally they are inclined at an angle and give rise to sloping rifts as in the Kabamba cave.

No calcite veins were seen but there are some thin partings of shale along joint lines,

probably injected during earth movements. At the entrance to Tsebahu contortions in the rock show that the shale was an incompetent material during folding. Minor cleavage structures may also be important lines of weakness, as in Matupi, where etchings on the roof and walls can be seen in the naturally illuminated parts of the cave. They follow minor structures and at one time moss and lichen grew along them, although the surface is now free of vegetation.

Bedding Planes — Since the limestones are for the most part thinly bedded, bedding planes have little effect on the form of the caves and few beds are sufficiently massive to control cave development. The only exception is the cave close to Lipanga near the top of the mountain where there is considerable horizontal development along one limestone bed.

Base Level — All the caves are closely associated with river systems but insufficient data are known about the hydrology of the area to assess the effect of base level control. Yolohafiri and the caves near the mountain summit have streams flowing through them which have been constantly developing lower routes, leaving a series of abandoned higher level caves. Excavation by the river in Yolohafiri is retarded by a thick shale band which gives rise to waterfalls a little way downstream. Erosion in the caves near the mountain summit is probably only limited by the rate at which the surface streams, receiving the cave drainage, entrench their own valleys.

Matupi, Kwama Kwama and their associated caves are now mainly dry and it is presumed that water passages exist at a lower level, as yet undiscovered. Some of the caves near the hotel are being eroded by active streams and most of their chambers and passageways are abandoned higher level caves.

Cave Infilling — Most of the caves contain deposits of bat guano and this conceals practically all other infilling material. Gravel, shingle and sand are found along some of the stream courses but this is recent and not significant in cave formation. Red clay occurs in several places, often in remote positions and may be present in sufficient quantities practically to seal off certain passages. It seems likely that the caves were at one time completely filled with this clay which has since been largely removed. It is not simply a deposit of the present streams.

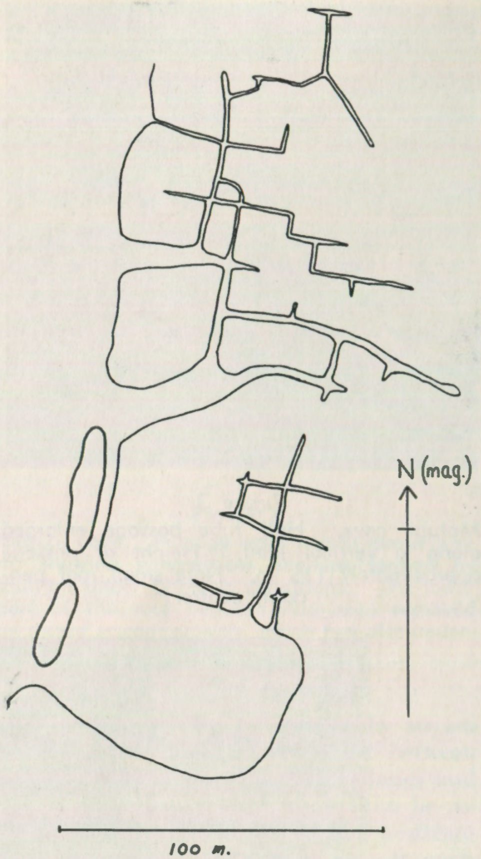

Figure 2
Plan of Matupi cave, Mont Hoyo.

Dripstone abounds in the main chamber of Matupi but is uncommon elsewhere. Most of them are dry but occasionally 'live' examples are found. Terraces of rimstone pools are present along some stream passages, a notable example occurring just beyond the stalactite grill in Lipanga.

Figure 3
Matupi cave. Half tube passage enlarged along a vertical joint. Height of passage approximately 1½ m. Note small roof pendant, centre.

Figure 4
Entrance to Matupi cave. Contortion in limestone close to rift fault.

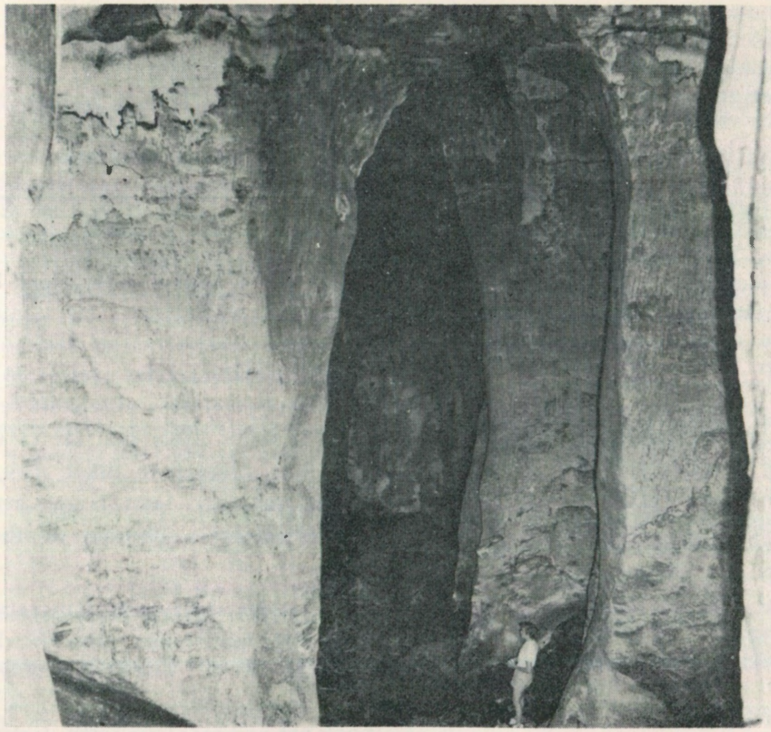

Figure 5
Tsebahu cave. High rift with side passages. Note rounded roof and absence of aven formation indicating phreatic origin.

Figure 6
Kabamba cave. Typical half tube passage in almost horizontally bedded limestone. Note undercutting of wall by recent vadose activity.

Figure 7
Matupi cave. Passage wall, height about 1½ m. showing symmetrical scallops formed by solution under phreatic conditions. The lower part of the roof pendant has been removed by recent stream erosion, which has also undercut the wall.

Vadose and Phreatic Features

Several of the caves are at present occupied by streams but there is much evidence to indicate that they were not formed by the erosional action of flowing water, that is to say under vadose conditions. They owe their origin to the solution of limestone by very slowly moving water at a time when the entire rock was saturated during a phreatic phase.

The principal features which point to a phreatic origin are essentially similar to those described for cave systems found in other parts of the world. Half tube forms (fig. 6), rounded roofs and scalloping of the walls could only have been created if the water completely filled the passageways. The network as seen in the plan of Matupi is not in the least like the pattern of stream tributaries. In addition there is no gradual growth of the passages along their stream course. Erosional features are absent from the walls and ceilings and are only rarely found along some of the present stream course. Roof pendants and symmetrical wall scalloping (fig. 7) are the product of solution and would not be formed by stream abrasion. False floors which exist between an upper and lower series of galleries and rock spans across passages must also be associated with solution effects for a stream abrading downwards would cut through them. Vadose streams which are now flowing in several of the caves are primarily excavating fill from pre-formed passages and only rarely, where the water falls in cascades for example, are they actively eroding solid rock. Even in those places where cave passages are growing upwards by aven formation at the top of the rifts there is no reason to suppose that the original passages were not formed phreatically.

Origin and Geological History of the Caves

It has been demonstrated in the previous section that the caves were initiated during a phreatic phase, but phreatic conditions would not be possible on a fault escarpment, where drainage would be free and rapid. Phreatic conditions could only exist when the limestone was below the water table, and it must be supposed that the caves were formed deep down before displacement by the fault exposed them in their present position.

Although the age of faulting is not known it might reasonably be supposed to date back to the disturbances during the Rift Valley formation which places it roughly in early Pleistocene times. The Mont Hoyo scarp is, however, considerably eroded, rather more so than the Rift Valley scarps, so this might be regarded as a minimum age of the caves. It has been shown that the caves are without doubt joint controlled which means they must post-date the formation of the joints. From this it follows the joints are older than the faulting and not necessarily associated with it directly.

If this reasoning is correct, then the caves must be older than Mont Hoyo itself. It may seem remarkable that caves, even if filled with clay, could be uplifted to such an extent without being destroyed, but the most likely explanation is that the rocks seem to have undergone very little disturbance except at the fault where a zone several yards thick is intensely contorted.

At several places the caves follow, to some extent, the lines of the folds caused by earth movement near the fault so that it might appear that the caves are, in fact, formed along post-uplift lines. This fold control however, is very slight and represents only a very small part of the cave development occurring quite late in their history. It may be significant that the largest chambers are all situated close to the fault scarp, representing extra cave development in the post-uplift times along or near the crushed and folded zone of weakness.

After uplift there would probably be a period of rapid excavation and occupation of the phreatic chambers by vadose streams. Further modification due to roof collapse, formation of stalactites and other dripstone deposits, growth of avens and erosional activity of streams would follow. Eventually the first formed caves would be abandoned as the streams cut their way down to lower levels at the present time.

Davis' (1930) theory of cave formation holds that caves are formed under phreatic conditions below the water table and after earth movement may become open to the surface and occupied by surface and vadose streams. Bretz (1942) follows Davis but also thinks there is a phase of cave infilling by red clay which is largely removed when the surface streams eventually excavate in the preformed caves. The caves of Mont Hoyo provide evidence in full support of these theories.

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SHORTER CONTRIBUTIONS

Bear Cave, A Tufa Cave in Glacial Drift Near Buchanan, Michigan

by Erhard M. Winkler and Alphonse C. Van Besien

Abstract—Two tufa deposits are north of Buchanan, Michigan, along St. Joseph River. The tufa at Bear Cave was precipitated on the Lake Nipissing level of Lake Michigan around interlocking logs of drift wood. A similar tufa deposit occurs at the base of Mocassin Bluff where a few alluvial fans from the bluff empty onto the Algoman terrace of the St. Joseph River. Tufa is still being deposited here. The lime for the extensive deposits was supplied from a bed of what appears to be inter-glacial silt (Sangamon loess?) of variable thickness hitherto undescribed in the geological literature of this area. Solution was secondary in the formation of Bear Cave. The passages are primarily original voids in the tufa or openings left from the decay and removal of organic material in logs buried in the tufa.

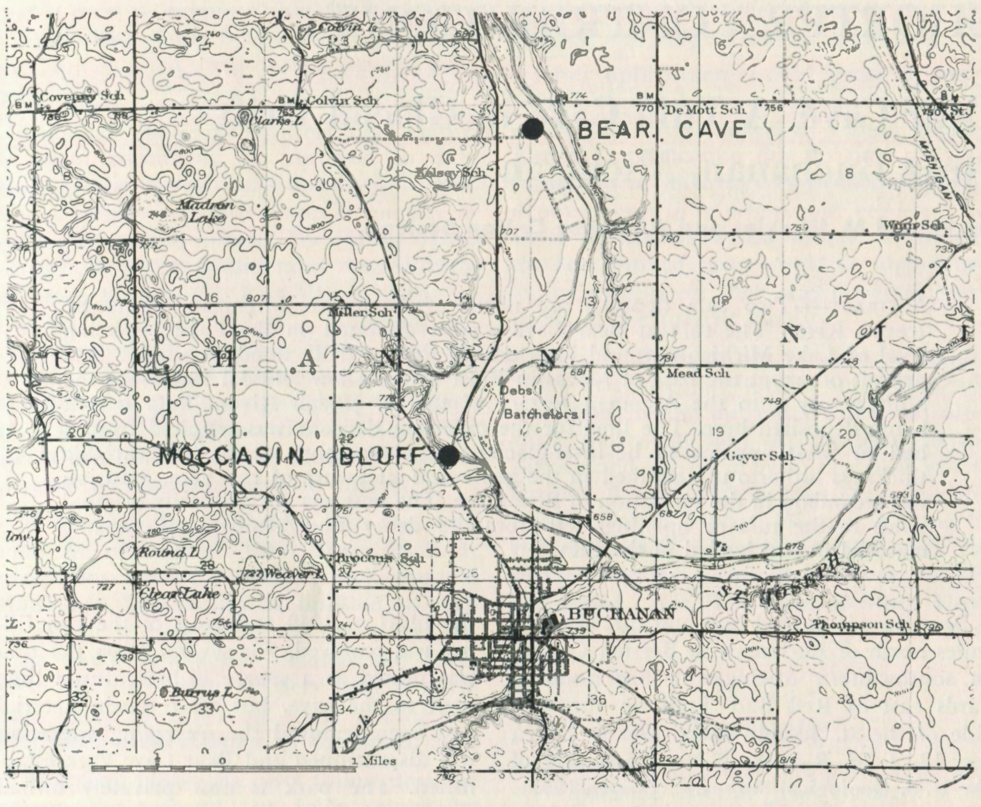
Bear Cave is located four and three-fourths miles north of Buchanan in Berrien County in southwestern Michigan a few hundred yards east of Red Bud Trail on the west side of the St. Joseph River (SW¼ NW¼ sec. 12, T-7-S, R-18-W; Niles quadrangle of the U. S. Geological Survey.) (Davies, 1955). The glacial map of Lower Michigan (1955) shows high lateral moraines of the former Lake Michigan ice lobe west and east of the river terraces which reflect higher stages of ancient Lake Michigan. Lake beds along the lower terraces are especially well developed between Buchanan and Berrien Springs, Michigan.

Bear Cave is in tufa deposits which are at least 18 feet thick. A 40-foot, steep, winding stairway connects the surface with the natural cavern. The winding cave is about 150 feet long, four to six feet wide, and ten to fifteen feet high. Indian tribes occupied it for relatively short periods of time. The cave has been a tourist attraction since a second opening was cut at the bottom of the bluff in 1940 and most of the clay filling was removed. For several years the tufa was mined as a source of agricultural lime next to the cave, but after much material had been removed the quarrying operation was discontinued and Bear Cave Park established. The park is now privately owned and operated.

The tufa at Bear Cave is light brown or buff and has a high porosity. A few well-defined imprints of logs, tree limbs, and leaves can be seen principally on the ceiling and walls of the cave. Most of the leaves represent *Alnus* (alder). Round concentric calcareous algal deposits ranging from two to five inches in diameter are occasional enclosures in the tufa.

At the base the tufa deposit has a sharp limonite band underlain by a band of dark gray to yellowish sandy clay two to three inches thick. Sand and coarse gravel form the base of the sequence. The clay contains a fauna of freshwater gastropods, the clam *Pisidium,* and a rich ostracod fauna with a strong representation of *Candona,* which was described by Winkler (1960). The presence of deeper water ostracodes and the absence of peat probably imply a large permanent lake.

ACKNOWLEDGEMENTS—We should like to express our thanks to the former owners of Bear Cave, Mr. and Mrs. Frank Allen, for their extensive cooperation during the field work, as well as to Mr. F. Wells Terwilliger for criticizing the manuscript.



Map of the vicinity of Bear Cave and Moccasin Bluff. From U. S. Geological Survey Topographic Map, Niles Michigan Quadrangle.

The tufa thins to the west edge of the terrace and is overlain by a few feet of light gray marl. Both the tufa and the marl underlie a river terrace. The slope of the next higher terrace level exposes firmly cemented gravel. West of the tufa terrace the cut for a small road leading up to the top of the bluff shows a sequence of sands and gravels with a thin wedge of light brown silt six inches thick near the 700-foot contour, about 50 feet above the tufa. Grain-size distribution, massive appearance, and the high carbonate content suggest that the silt is derived from loess located nearby. About 50 feet below the silt a small pond is fringed with white marl. The unusually large quantity of lime present in this vicinity is striking. The calcareous silt, although thin, is a likely source for the tufa of Bear Cave.

Another tufa deposit of similar lithology is known along the base of Mocassin Bluff about one mile north of Buchanan and about two and three-fourths miles south of Bear Cave (NW¼ NE¼ SW¼ sec. 23, T-7-S, R-18-W; Niles quadrangle). Here a terrace along the St. Joseph River at a level lower than at Bear Cave occurs. The terrace is at the foot of the steep Mocassin Bluff from which several alluvial fans spread onto the large flat terrace surface. The age of this terrace was tentatively identified as Algoman, about 15 feet above the present stream level. Three higher terraces may be seen in this area. The tufa exposed below the 700-foot level of Mocassin Bluff closely resembles that at Bear Cave. The steep Mocassin Bluff is in part composed of silt with massive appearance which resembles

loess very closely. The silt overlooks the lower terrace displaying its characteristic vertical exposure at this place. It is light tan to buff silt with as much as 45% calcium carbonate and it is overlain by gravel with a sharp but wavy contact. No till is exposed above the loess-like silt. Three layers of concretionary zones occur within the silt, but only two are well developed. One is located about one foot below the top and is about two to three inches thick. The second is a continuous bed a few inches thick located very close to the base of the silt which is about 12 feet thick.

The contact of the silt with the underlying till is believed to coincide with a zone of wet ground studded with scouring rush near the base of the embankment, indicating the water table. Strong flow of water from springs have dissected the alluvial fans which cover the low terrace. Tufa is apparently still being deposited above water level at this place.

There are only two actual exposures of silt along Mocassin Bluff, both about 12 feet thick, one along Mocassin Bluff Road and the other about 100 yards west of it. The thickness diminishes to a few inches near Bear Cave although a distinct bluff can be followed from here to the north, parallel to the course of the river, beyond the site of Bear Cave. A disconformable deposition of the silt on top of a formerly dissected till surface may explain the large difference in thickness within this short distance. The well-developed silt at Mocassin Bluff is similar to that of the Bear Cave site. At both sites the silt has an elevation of about 700 foot whereas the tufa is much lower on a river terrace that is younger than the silt.

Up to this date no loess-like silt nor loess had been reported from any place in southwestern Michigan or northwestern Indiana. The authors tentatively assign the silt to the early Wisconsin and middle Wisconsin which would correspond to the Peoria Loess further west.

Insoluble acid residues were prepared from samples taken at one foot intervals through the upper ten feet at the Mocassin Bluff site nearly down to the bottom contact. The residues were obtained from twenty gram samples:

Depth from top of formation	*% of $CaCO_3$*
surface	43.60%
1 foot	44.05%
2 feet	40.00%
3	45.40%
4	45.50%
5	40.65%
6	47.10%
7	42.75%
8	34.85%
9	33.90%

The profile shows fluctuation down to 6 feet of the acid soluble calcium carbonate, then a clear decrease towards the bottom of the profile. Because mostly gravel overlies the loess no groundwater zone is believed to have formed above the silt. Thus the groundwater may have percolated and leached the lowest part of the silt deposit just above the bottom to concentrate some of the lime in the lower concretionary horizons. The thin concretion zone one foot below the top of the profile may be derived from surface waters which were saturated with humic acid leaching either some lime from the uppermost layer of the silt or from the gravel above redepositing the lime just below.

Geological History of Bear Cave

The St. Joseph River may be considered a characteristic stream in uncemented glacial drift which quickly adjusted its course towards a parabolic long profile; it begins southwest of Jackson, Michigan and ends in Lake Michigan at Benton Harbor, Michigan with an approximate length of 175 miles. Each level of ancient Lake Michigan caused a quick adjustment back to a parabolic long stream profile which is well displayed in the different terrace levels along the stream course. Lake Algonquin was probably the first stage which may have affected the history of this area. It is about 8,000 years old according to Hoff (1958) and was about 25 feet above the present lake level. This stage was followed by the

Lake Chippewa low level which was about 350 feet below the present day level. Zumberge and Potzger (1956) assumed the duration of this latter stage about 4,000 years, Hough (1958) much less. This low lake stage may have caused drastic entrenchment of St. Joseph River into the underlying Mississippian and Devonian shales from Lake Michigan about 22 miles back into the area of Buchanan provided that the discharge of the river at that time was similar or greater compared to the present. The subsequent rising of the Lake Michigan level to between 605 to 620 feet during Nipissing time about 4,000 years ago must have caused a lake along the entrenched channel. The glacial map of the Southern Peninsula of Michigan (1955) shows lake beds along the St. Joseph River from Benton Harbor upstream to about Niles, Michigan. The thin lake beds exposed at the base of the tufa in the cave indicate a quiet water environment for the clay and probably for the tufa. No plant remains could be found within the clay. The tufa deposition was probably restricted to the Lake Nipissing stage. Leaves and driftwood apparently were plentiful close to the shore and they must have been covered with green algae which promoted the lime precipitation from springs. Logs of driftwood were introduced during spring flood activity; these were interlocked with each other to form the framework for the narrow channel of Bear Cave. The organic part has since decomposed entirely forming small cavernous openings, and leaving external molds of tree trunks, branches, and alder leaves. The cave channel probably was kept open during the time of tufa deposition and was spared from being filled with silt and clay. Fill that did accumulate from soil washed in to the cave was later excavated by man when the cave served as a shelter. Solution within the tufa is probably only of secondary importance. Some dripstones, however, can be seen in the cave as well as some dripstone crusts both inside the cave and outside along the bluff to the river. Davies (1955) only mentions northwest south-east trending cracks in Pleistocene marl which form the original passages of the cave.

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Cross Piracy Drainage Development in the Newsome Sinks Area of Alabama

by W. W. Varnedoe, Jr.

Abstract——In the Newsome Sinks area of Morgan County of northern Alabama a young, surface stream cut across and down into an older, underground drainage system that formerly resurged at Skidmore Cave. As a result each captured the drainage of the other. The present alignment of the caves in the area, the Newsome Sinks water resurgence at Hughes Spring and the resurgence at Skidmore Cave can best be explained by this history. Newsome Sinks is today a large landlocked valley containing no surface streams and numerous caves.

Newsome Sinks in Morgan County Alabama has been studied extensively by the Huntsville Grotto of the National Speleological Society. Newsome Sinks is a landlocked, north-south valley about 4 miles long and 1 mile wide (fig. 1). It drains a much larger area, estimated at 14 square miles. All surface runoff disappears underground and there is no surface stream flowing along the axis of the valley. There are numerous caves and sinks and over 50,000 feet of cave passages have been mapped within the valley.

Newsome Sinks and other nearby valleys are cut into Brindley Mountain which is a plateau underlain by limestone of Missisippian age capped by sandstone of the Pennsylvanian. It is a remnant of the Cumberland Plateau, cut off by the Tennessee River. The flat bedded limestones dip very gently to the south-southwest. The plateau upland in this area is about 1100 feet above sea level and the valley floors are about 500 feet lower.

In Newsome Sinks few of the caves actually are connected but they were all formerly part of a common system (fig. 1). The axis of the caves generally line up with the axis of the valley.

A peculiarity, however, exists in the pattern of the caves. It has been assumed that Hughes Spring at the southern end of the valley was the resurgence of the underground streams in the Sinks. However, after detailed mapping was completed it was apparent that Hughes Spring does not lie on the "Newsome Axis". Further, by extending the axis line on across the east-west valley at the southern end of Newsome Sinks, it intersected Skidmore Cave lying further to the south (fig. 2). Since the flow from Hughes Spring did not seem adequate for the drainage of Newsome Sinks and Skidmore discharged a large flow of water, it was speculated that Hughes drained only Curry Cove and Skidmore drained Newsome Sinks by flow under the cross valley.

Flourescein dye placed in Wolf Cave in the Sinks appeared at Hughes Spring two weeks later, however, establishing Hughes Spring as the Newsome Sinks resurgence. In addition, study of topographic maps shows the water from Lynn Falls in Cross Valley disappears in a rock filled sink and can only be accounted for by reappearance at Skidmore Cave.

The pattern of Hughes Cave has the same orientation of other Newsome Sinks caves (fig. 1), but Hughes cave opens into Cross Valley and lies outside Newsome Sinks. Outside the entrance to Hughes Cave there is a fossil stalagmite on the hill slope which shows the cave once extended further into what is now Cross Valley. A search of the opposite bank of this valley also disclosed several sinks aligned with the Newsome Axis, but they could not be entered because of breakdown.

Newsome Sinks, apart from its lack of surface streams, is a normal valley sharing the slope angles of all the mountain sides in the area. Cross valley, on the other hand, is narrow, has steep sides which are in places

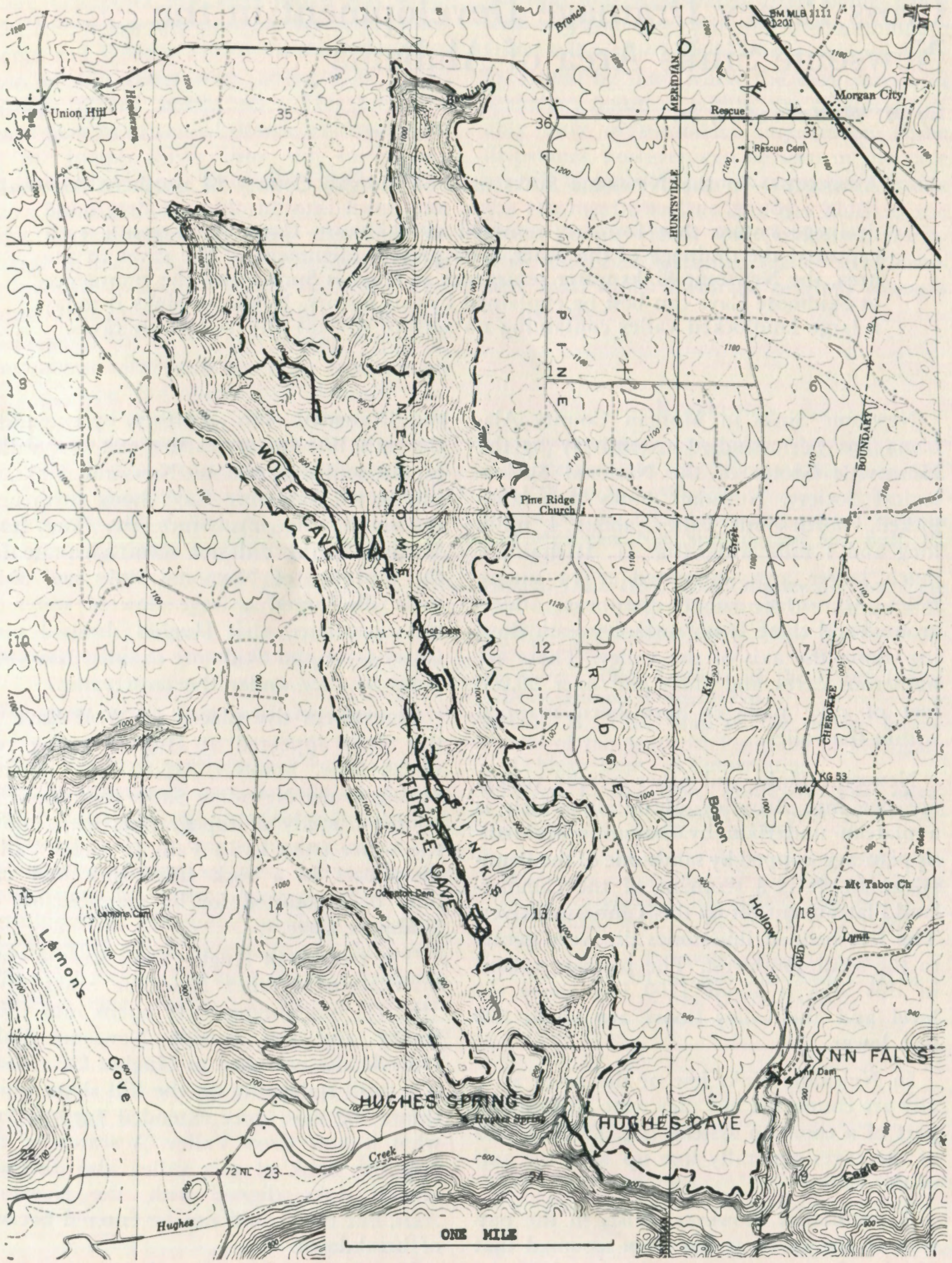

Figure 1
Newsome Sinks and vicinity. Modified from TVA topographic map. Cave patterns shown by solid lines.

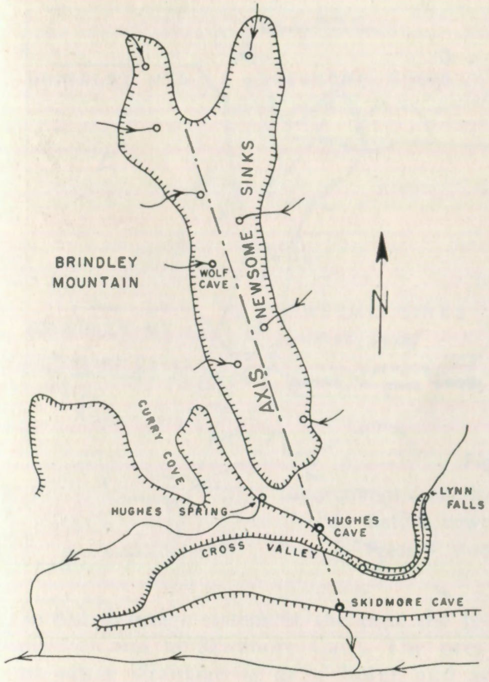

Figure 2
Newsome axis and its extension to Skidmore Cave.

Figure 3
Lynn Falls at the head of Cross Valley.

sheer cliffs, and is headed by a waterfall, Lynn Falls (fig. 3). From this it appears that Cross Valley is geologically younger than the neighboring valleys and Newsome Sinks.

The following sequence of events explains the present topography, cave pattern and drainage system.

The original drainage was a surface one running along the Newsome Axis to the south where it joined another stream heading west at a location a little to the south of the present Skidmore Cave (fig. 4A). By the time the axis was established and the sandstone cap breached, underground drainage developed along this axis. This new drainage was instrumental in the development of the later stages of most of the present caves, including Hughes Cave (fig. 4B). The resurgence was at Skidmore Cave. Cross Valley did not exist, except perhaps as a small surface stream on the plateau upland.

Fairly recently headward erosion in Cross Valley was rapid near Newsome Sinks. The subterranean drainage of the Newsome Axis was intersected, and captured by Cross Valley (fig. 4C). Further down-cutting of Cross Valley has shifted the Newsome resurgence from Hughes Cave to Hughes Spring.

As the Cross Valley grew headwards a sink developed in its bed that diverted the Cross Valley water into the downstream underground Newsome Axis conduit which delivered this flow to Skidmore Cave. This is the pattern that exists today (fig. 4D). The gradients are compatible because Hughes Spring is lower than the old Newsome Axis drainage and the sink in Cross Valley feeding Skidmore is higher (fig. 5).

The caves of Newsome Sinks contain little flowing water. Some, of course, contain small, local streams, but none contain large active streams receiving drainage from major surface sources. The caves represent

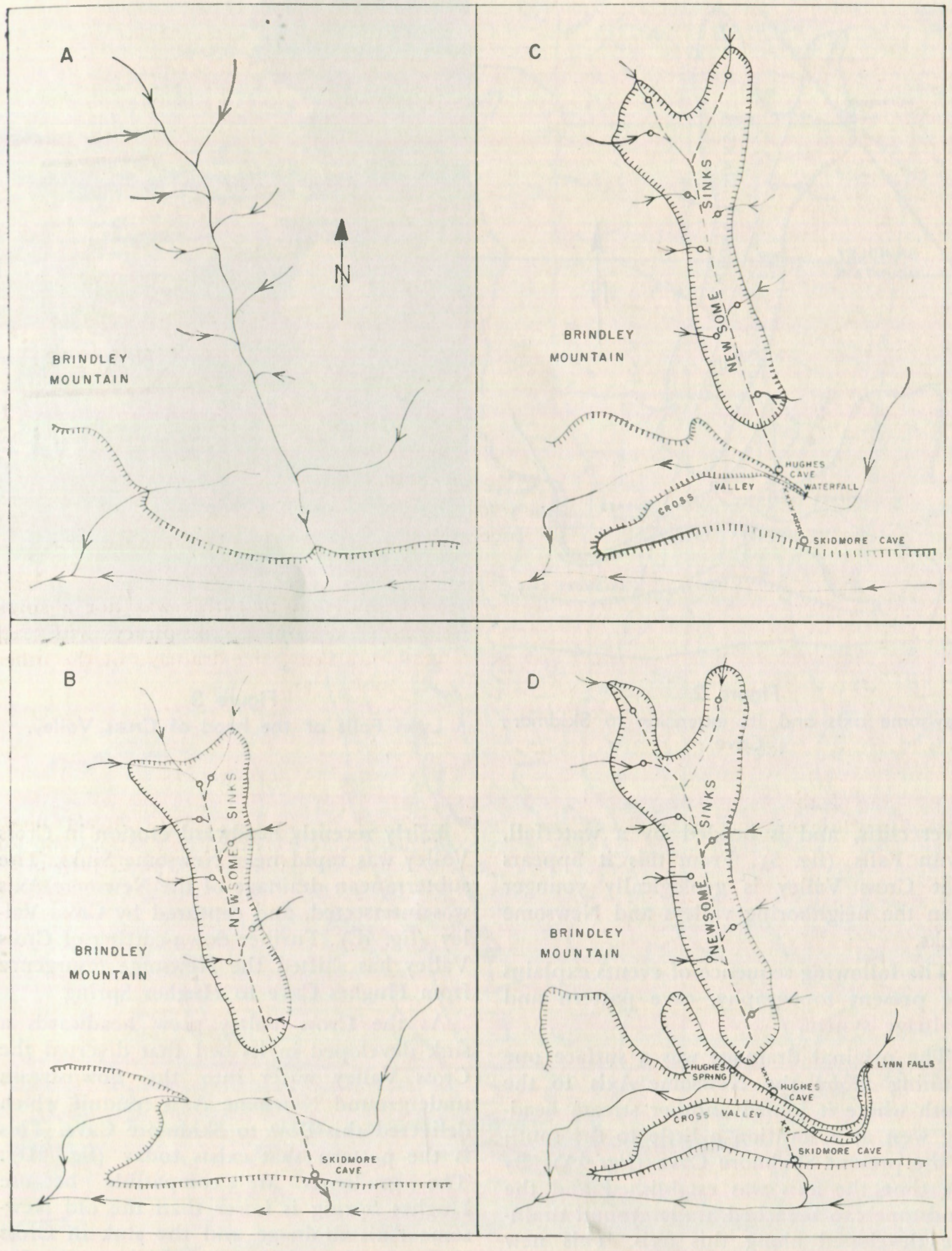

Figure 4—Evolution of Newsome Sink. A. Surface drainage to the south. B. Development of young sinks with diversion of drainage underground. Small surface stream crosses this drainage. C. Cross Valley cuts down into underground drainage and captures it. D. Further headward erosion in Cross Valley diverts its drainage into older underground system (present condition.) Solid lines are surface streams; dashed lines are underground drainage channels. Caves and abandoned underground channels shown by xxxx.

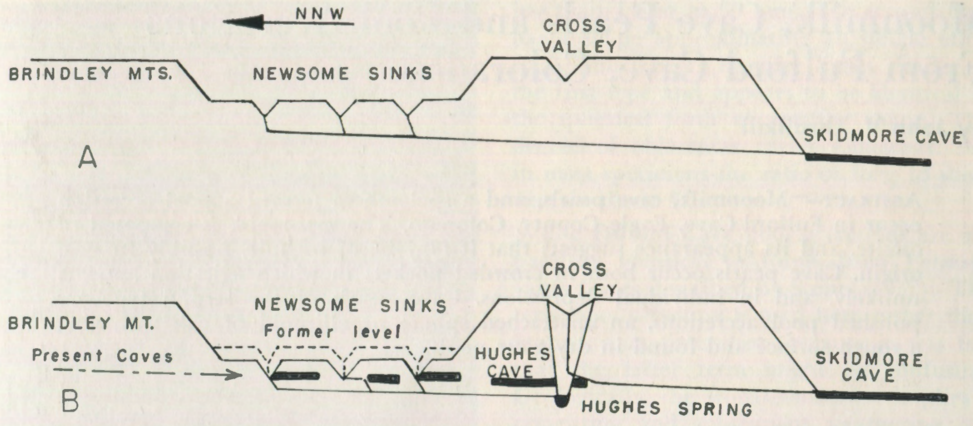

Figure 5

Diagrammatic section along Newsome Axis

A. Before downcutting of Cross Valley

B. Present stage

the old drainage system at the time the resurgence was at Skidmore Cave. The present active drainage is at a lower and as yet inaccessible level in the Sinks.

This theory is in agreement with that proposed by Woodward (1961). The two catastropic events relative to this are the breaching of the sandstone cap to establish the first subterranean drainage gradient and the Cross Valley capture and subsequent lowering of the Newsome Sinks water table.

It is unique in that this was not a single stream piracy but a cross-piracy, with each stream capturing the drainage of the other, both underground.

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Moonmilk, Cave Pearls and Pool Accretions From Fulford Cave, Colorado

by John V. Thrailkill

ABSTRACT—Moonmilk, cave pearls, and a speleothem termed a pool accretion occur in Fulford Cave, Eagle County, Colorado. The moonmilk is composed of calcite, and its appearance suggests that it may be of organic, possibly fungus, origin. Cave pearls occur both in crowded pockets in which agitation appears unlikely, and in individual depressions. Cave pearls of the second type are polished pool accretions, an unattached sphere or ellipsoid of calcite having a rough surface and found in dry gour pools.

During an investigation of the origin and development of Fulford Cave, Eagle County, Colorado, a study was made of the speleothems. Of these, moonmilk, cave pearls, and a form here referred to as *pool accretions* are of special interest. Although all of these have been discussed in the literature, photographs of moonmilk and information on the relationship between cave pearls and pool accretions are generally lacking. The reader is referred to a previously published paper (Thrailkill, 1960) for a description and map of Fulford Cave.

Moonmilk occurs in Fulford Cave as a damp, soft, creamy white encrustation both on the limestone walls and on stalactites and draperies (fig. 1) in several areas in the cave, most notably in the Moonmilk Corridor. A photomicrograph of an impregnated thin-section is shown in figure 2. The dissimilarity in appearance between this figure and figure 2 in the paper by Davies and Moore (1957,p. 26) may be due partly to the latter photomicrograph being of loose material (Moore, personal communication). X-ray diffraction analysis of the Fulford Cave moonmilk by Mr. Albert Dearth of the University of Colorado showed it to be calcite.

The reader is referred to Davies and Moore (1957) and Moore (1960) for descriptions of other occurrences of moonmilk, as well as discussions of the origin of the picturesque name. The appearance of the Fulford Cave moonmilk in thin section (fig. 2) may lend some support to their conclusion that moonmilk is of organic origin. A similar origin has been proposed for lublinite, a fibrous variety of calcite occurring as a coating on surface exposures of marl, travertine, and other rocks in eastern Europe. This material, which has also been called *rock-milk,* was found to contain the hyphomycetes of fungi with the fibers of the hyphomycetes determining the structure of calcite (Ulrich, 1938).

Cave pearls are present in several places in the Moonmilk Corridor and elsewhere in the cave. They were found in two rather different situations. The first and most common occurrence is in small flowstone-lined and water-filled pockets on the passage floor, usually near a wall. One of these pockets contains as many as twenty irregularly shaped but highly polished cave pearls of all sizes up to about 20 mm in diameter, often four or five layers deep. The nucleus of these pearls is usually large and consists of limestone or chert fragments or of grains of minerals exotic to the cave (figs. 3 and 4). In most of these pockets the cave pearls are so crowded it is difficult to believe they have been agitated, let alone rotated.

The second and less common occurrence of cave pearls in Fulford Cave is in small depressions on the floor, usually in gour basins. The inner surface of each of these small depressions is highly polished wet flowstone, and the depression conforms to the size and shape of the contained cave pearl. Because few pearls of this type were found, only one was collected and sectioned. Figure 5 shows the depression from which the pearl was collected, as well as several

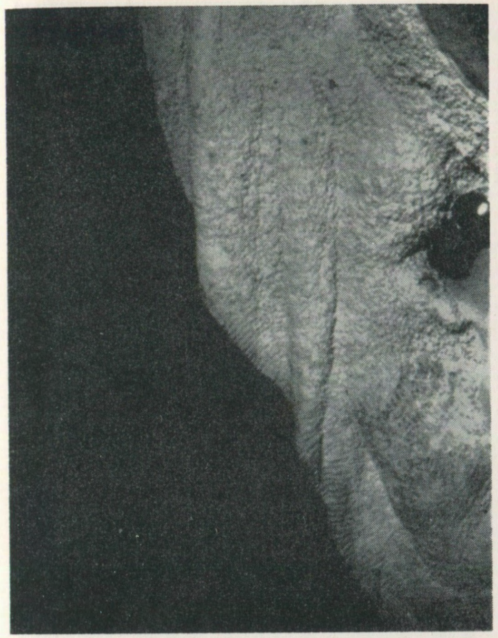

Figure 1
Drapery coated with moonmilk

undisturbed pearls. The collected cave pearl is essentially a polished pool accretion, although the outer 0.3 mm shows an indistinct layering (fig. 6).

A third speleothem of interest in Fulford Cave is what is here called a *pool accretion.* This is an unattached spherical or ellipsoidal body of calcite, specimens of which were found in some abundance in the bottoms of the dry gour pools in the Moonmilk Corridor (fig. 5). These forms are distinguished from cave pearls by their rough outer surface and lack of nucleus (although a small nucleus may have been missed in the two accretions sectioned). Their internal structure consists of relatively large but poorly defined calcite crystals scattered throughout a matrix of cloudy calcite. There is a fairly well defined concentric structure caused by an alignment of pore spaces, and a vague radial structure due to a rough arrangement of the larger calcite crystals (figs. 6 and 7).

Two varieties of pool accretions were observed. The first and most common type is roughly spherical and ranges in size from less than 1 mm to 30 mm. The second type has a shape approximating an oblate ellipsoid. This variety has size limits similar to the first type and appears to be identical to the spherical form except for shape. The degree of oblateness varies somewhat, but in most specimens the ratio of long to short axes is about 3:1.

The Fulford Cave pool accretions are apparently identical to forms from Pennsylvania Caves described by Stone (1932). The term *pool accretion* is used here rather than Stone's *cave concretion* because it was felt that the latter term might be confusing. Geologically, the word *concretion* implies a cementing and subsequent weathering out of a portion of a solid mass of rock, and the French word *concrétion* is almost synonymous with our word *speleothem.*

The very different morphology and environment of cave pearls and pool accretions points to a markedly different origin, and there seems to be little reason to "lump" the two forms. The cave pearls found in small depressions in gour basins are presumably all pool accretions which have been kept moist and possibly agitated by a drip of water, thus acquiring a polish. It is doubtful, however, that agitation is a general requirement for the formation of cave pearls.

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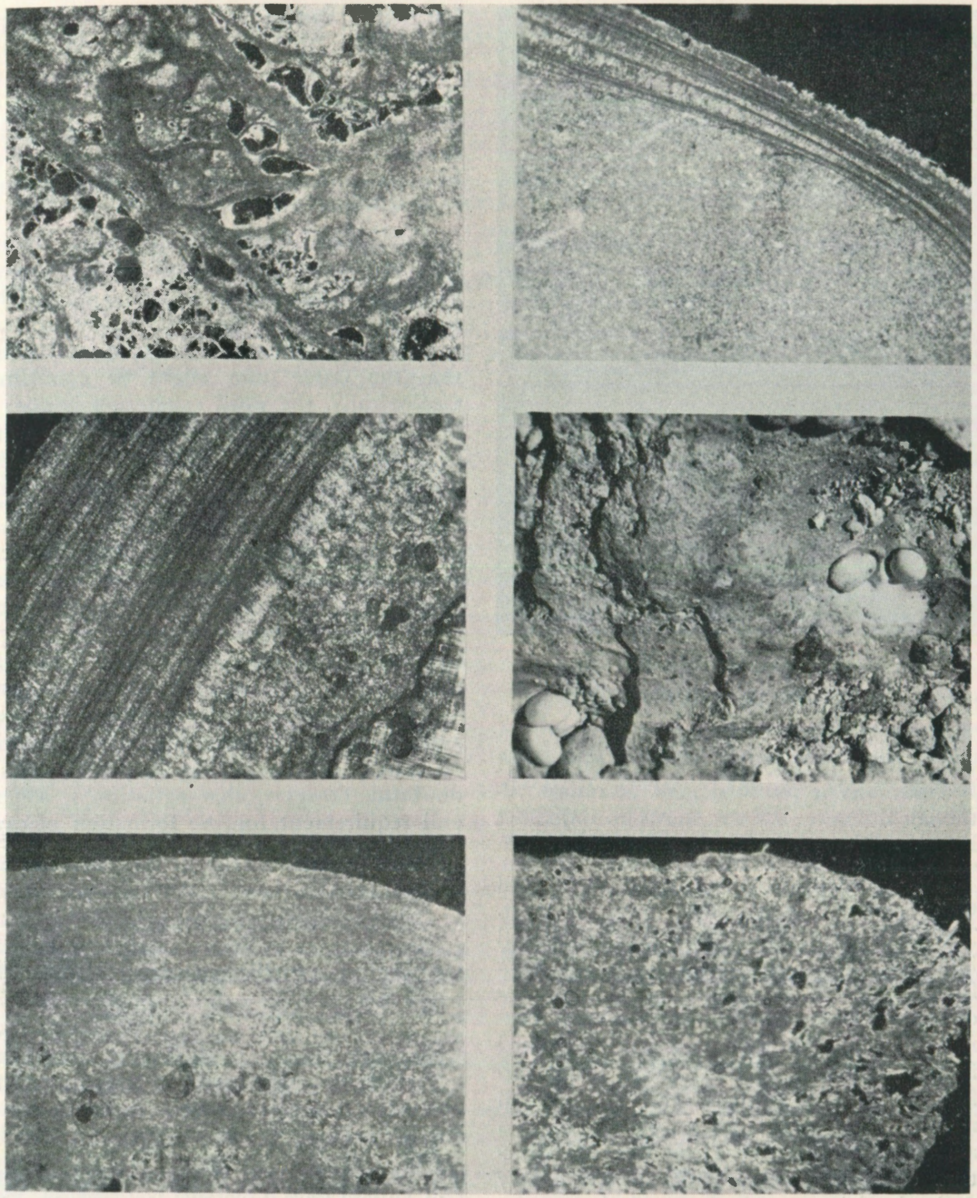

Figure 2—Photomicrograph of moonmilk. Polarized light. X15 (upper left); Figure 3—Photomicrograph of cave pearl from pocket in Moonmilk Corridor. Nucleus is lithographic limestone. Polarized light. X20 (upper right); Figure 4—Photomicrograph of cave pearl from pocket in Stalagmite Room. Polarized light X33 (center left); Figure 5—Cave pearls in small depressions on floor of Moonmilk Corridor. One pearl was removed from group on right for study. Pearls average 15mm in diameter. Rough pebbles in lower right are pool accretions (center right); Figure 6—Photomicrograph of cave pearl from depression shown in figure 5. Center is pool accretion. Polarized light X16 (lower left); Figure 7—Photomicrograph of ellipsoidal pool accretion. Polarized light. X12 (lower right).

BULLETIN
of
THE NATIONAL SPELEOLOGICAL SOCIETY

VOLUME 25 **JULY 1963**

CONTENTS

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