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NATIONAL SPELEOLOGICAL SOCIETY

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Contents

OCTOBER 1965

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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 November 15, 1965.

Papers in Press for Forthcoming Issues

The Origin of Limestone Caverns: A Model From the Central Mendip Hills, England

By Derek C. Ford

ABSTRACT

Theories of limestone cavern genesis are divided into three conflicting groups: vadose, water table, and deep phreatic. Type examples for all three can be found, at short distances apart and occupying similar positions in the regional geologic structure, in caverns of the central Mendip Hills in southwest England. Representative caves are described:

1. St. Cuthbert's Swallet is a drained phreatic system evidencing well integrated cave development to a depth not less than 280 feet beneath the contemporary water table.

2. Swildon's Hole is a water table system with four major cave levels. Accordance of the elevations of principal passages with defined water tables becomes progressively closer through the lower (younger), levels.

3. G. B. Cave is a largely vadose system, descending through the same range of absolute elevation as the phreatic and water table caves. At its lowest points, passages assume a water table form and are abruptly reduced in dimensions.

4. The Cheddar Caves and Wookey Hole discharge the flow of the above-mentioned systems. They are composed of both deep phreatic and water table cave elements.

The evidence is reconciled in a single model: in the vertical plane, cave passages make deep loops below, and returning to, a linear water table which they have created. The amplitude of this phreatic loop penetration becomes reduced with growth in the quantity of minute groundwater conduits which may be utilized to guide principal passages. St. Cuthbert's Swallet and Wookey Hole are parts of single phreatic loops. Swildon's Hole and the Cheddar Caves contain several loops, with adjustments to negative shifts of an allogenic base level. G. B. Cave is the youngest in the sample and occupies an uppermost zone of the rock which was drained and air-filled before a stablized water table existed locally.

The three largest systems are of substantially different ages. The time at which development commenced appears to have been determined by an interplay of three geomorphic factors: volume of available groundwater, hydraulic gradient in the rock and the efficiency of overground discharge of run-off.

INTRODUCTION

At the present time, theory of limestone cavern genesis is in some confusion. Each of the three karst groundwater zones defined by Cvijić (1918, p. 482), has been nominated as the locus of maximum cave formation.

Vadose theories (Dwerryhouse, 1907, Matson, 1909), hold that most cave expansion takes place above a water table (vadose zone), because rates of groundwater flow will be highest there. Processes of mechanical cor-

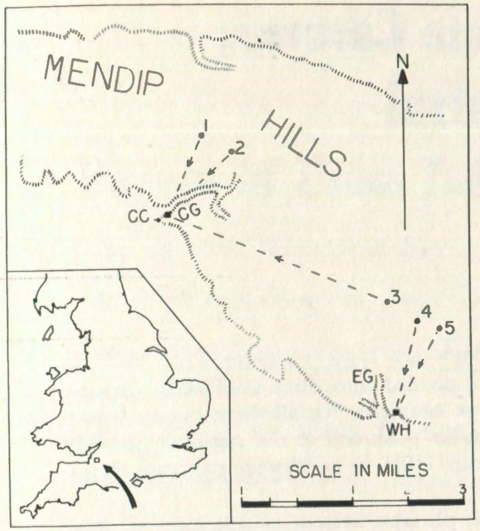

Figure 1.
The central Mendip region showing the location of the principal caves. 1 — G. B. Cave: 2 — Longwood Swallet: 3 — Swildon's Hole: 4 — Eastwater Cavern: 5 — St. Cuthbert's Swallet. CG — Cheddar Gorge: CC — Cheddar Caves: EG — Ebbor Gorge: WH — Wookey Hole. Dashed lines indicate generalized lines of groundwater flow from sink to rising.

rasion can operate effectively, in addition to carbonate solution.

Water table theories (principally those of Swinnerton, 1932, and Rhoades and Sinacori, 1941), argue that the plane of the water table itself will attract the greatest confluence of groundwaters, with a consequent development of the greatest caves along it. Establishment of the water table precedes most of the cavern expansion. Because the elevation of the water table fluctuates with variations in the volume of groundwater flow there may be substantial cave development within a narrow vertical zone above and below its mean position.

Phreatic theories (Davis, 1930, and Bretz, 1942), place cave initiation* and the bulk of expansion at random depth below the water table (deep phreatic zone). From field evidence it is deduced that groundwater solution cannot compete quantitatively with channelled erosion of all kinds at the surface in a region of moderate to high relief. Groundwater erosion thus assumes significance only in a region of low relief, (or Old Age), where the water table will be at, or very close to, the surface of the land.

White and Longyear (1962, p. 167), make a sweeping criticism of the debate:

"the multitudinous theories. . .are neither correct nor incorrect in the general case, *they are irrelevant**, and each may be developed as a special consequence of a specific flow pattern. Each theory is hopelessly entangled with the concept of the karst water table. . . . . .(This does not directly control the velocity of groundwater flow and thus does not control the zone of maximum limestone removal, (the cave)."

Each of the three principal theoretical positions, however, has been supported by recent work in specific cavernous areas*. Caves of the central Mendip Hills in Somerset, England, present an unusually complex situation. There are five large "swallet" or influent caves, (numbered 1-5 in figure 1), which draw water underground at the contact of limestone and non-carbonate rocks and discharge it at two major springs (CC and WH). The five influents occupy nearly identical positions in the regional lithology, structure and relief. At first inspection, the pair in the west appear to be type examples of the vadose theory. No. 3 is a water table cave with four major levels of development. To the east, nos. 4 and 5 are three-dimensional honeycombs that suggest cavern expansion at random depth below any contemporary water table.

*Warwick (1962, p. 72) and other writers distinguish between the conditions of "cave initiation" (or "origin"), and "cave development". The point is discussed on page 125.

*White and Longyear's emphasis.

***E.g.* , for the vadose case, Simpson. 1935, in Yorkshire, England, Daniel, 1961, in Alabama. For the water table case, Sweeting, 1950, in Yorkshire, Davies, 1960, and White, 1960, in the Appalachians; Ek, 1962, in Belgium. For the phreatic case, Bretz (1956) in the Ozarks.

This paper offers a resolution of the apparent tangle of theories. It is based upon a detailed study of three of the influent caves (vadose, water table and phreatic types respectively), and the effluent caverns to which they feed. The morphology and relevant geologic structure of all accessible cave passages was mapped in the field at a scale of 20 feet to the inch. Surface karst and denudation chronological features of the cavernous area were also mapped.

ACKNOWLEDGEMENTS

The writer is indebted to the Professor of Geography, Oxford University, and the Directorate of Scientific and Industrial Research, Great Britain, and to the President of McMaster University, for grants in aid of research. Dr. M. M. Sweeting, Oxford University, gave much critical advice during the preparation of the thesis from which this paper is drawn. The field work upon which it is based could not have been carried out without the aid of many Mendip cave explorers.

THE CENTRAL MENDIP REGION

In the broadest terms, the Mendip Hills are an isolated tongue of high ground thrusting to the eastern shore of the Bristol Channel and bounded by low-lying moors. Figure 2 summarizes the geology and relief of the central hills. They are periclinal structures of Upper Paleozoic age (Welch, 1929, p. 46). Resistant quartzitic sandstones at the cores of the periclines are gathering grounds for surface run-off which crosses a narrow band of shales and sinks immediately upon reaching the flanking limestone (the influent cave sites). The limestone is of Mississippian (Lower Carboniferous) age and 2,000 – 2,800 feet thick.

The limestones are quite pure. $CaCO_3$ comprises 82:64 – 99.80% by volume, and MgO, 0.00 – 10.94% (Chapman, 1905, p. 500). Texture ranges from the calcirudite to porcellanous types. Primary porosity is not significant. The rock is well bedded, bedding being thin toward the base of the formation, where there is much shale and some discon-

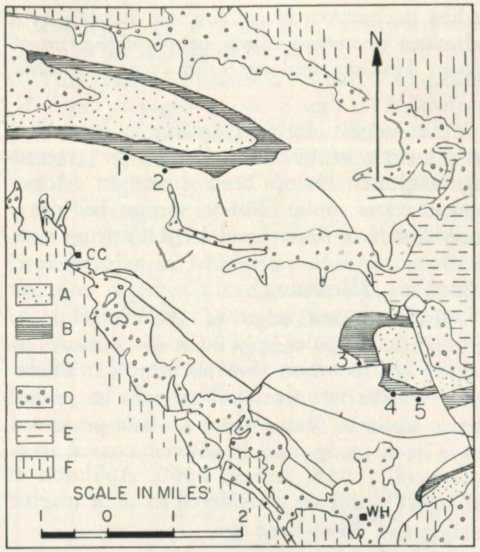

Figure 2.
Geology of the central Mendip region. Numbers as in Fig. 1. A — Devonian sandstone; B — Carboniferous shale; C — Carboniferous limestone; D — Permo-Triassic dolomitic conglomerate; E — later, impermeable, Mesozoic strata on the plateau; F — Mesozoic and younger impermeable strata bounding the Mendip Hills.

tinuous chert interlamination. In the area of the sample caves, beds dip southward at 15° – 45°. Jointing is dense and there has been some major faulting. In Permo-Triassic times, the Upper Paleozoic structure suffered arid denudation and was partly buried by a dolomitic conglomerate of local derivation. This is preserved as valley fill in parts of the present massif. The hills were then wholly buried by strata of later Mesozoic age. Rejuvenation of the early structure and the start of exhumation is attributed to the Tertiary, probably Miocene (Jones, 1930, p. 74).

Today the oldest erosion feature is a later Tertiary surface bevelling the dipping limestones to form a plateau at 800 – 850 feet above m.s.l. It is dissected by dry valleys converging upon the dry limestone gorges of Cheddar and Ebbor (fig. 1). The valley floors are dotted with karst depressions. There is much evidence to show that the valleys were re-activated periodically in the later Pleistocene, when permafrost layers formed in periglacial climates prevented much or all of the run-off from sinking underground (e.g. Reynolds, 1927, p. 191).

The largest regional spring is located at the mouth of Cheddar Gorge, the groundwater clearly having been drawn to a topographic low point. Ebbor Gorge has no equivalent, but the second largest spring rises 950 yards to the southeast at a lower level (Wookey Hole rising).

The southern edge of the central Mendip Hills drops steeply from the plateau elevation to 70 – 200 feet above m.s.l. There are remains of erosional terraces at several levels upon it. Those which are best preserved have been recognized in adjacent coastal areas (Driscoll, 1958, Arber, 1960, ApSimon et al., 1961), and are considered to be of marine origin and Pleistocene age.

The modern climate is Temperate West Coast Marine type (Thornthwaite, 1948, type $B_4B'_1ra'$). Mean annual temperature at 800 – 850 feet is c.48°F. with a mild winter. Mean rainfall is 45.0 inches per year and evenly distributed throughout the seasons. Water available for runoff is 21.0 inches. The budget of the principal springs indicates that almost all of the runoff must pass into the limestone mass.

The Sample Caves

1. *A phreatic influent cave: St. Cuthbert's Swallet*

The cave is entered at 783 feet above m.s.l. and 200 feet south of the shale contact. The lowest accessible point is a siphon at 382 feet. Approximately 5,000 feet of cave passage are shown in figure 3, a simplified map of the system.

The modern surface catchment area is 0.35 square miles, yielding a mean flow of 0.45 c.f.s. to the cave. In the past, catchment may have been as large as 3.5 square miles. It has been reduced by the development of independent sinkholes. The water rises at Wookey Hole, 1.6 miles to the south, at 200 feet above m.s.l.

The cave is composed of four distince geomorphological areas. *Area A* (figs. 3 and 4), is a system of rough, recent vadose shafts dropping very steeply to the shale contact. They descend into and through *Area B*, a central complex of collapsed phreatic chambers of much greater age. Phreatic elements are traced up to 660 feet. They may have extended much higher but have been lost in collapse. The central complex was a honeycomb of bedding plane, joint and fault fissures perched upon the shale basement. All preserved surfaces show large but shallow solutional pocketing (Bretz, 1942, p. 690). There are no evidences of erosion or deposition at a standing water surface. The complex developed below a water table that must be set at 660 feet or higher.

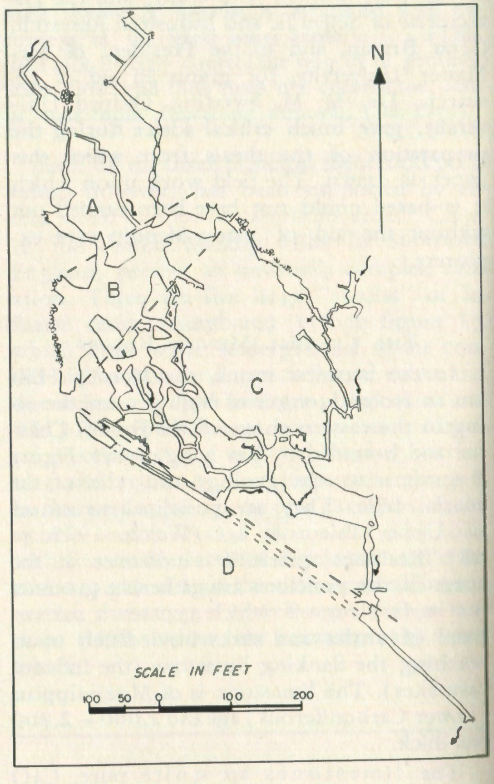

Figure 3.
Plan of part of St. Cuthbert's Swallet. Letters are keyed to the text. Curly arrows indicate the location and direction of flow of principal modern streams (adapted from a C. R. G. Grade 5 survey by B. E. Ellis, D. A. Coase, D. C. Ford and others).

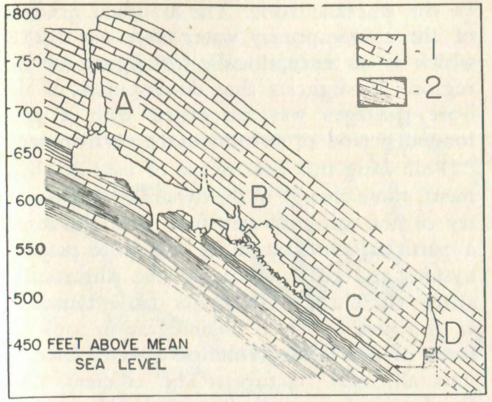

Figure 4.
Diagrammatic long section drawn through the western half of St. Cuthbert's Swallet. Letters are keyed to the text. 1. limestone, 2. passage beds and shale. Abscissa scale is the same as the ordinate scale.

It was drained through *Area C*, the "Warren." This is a system of tubes of much smaller volume than the great voids of *Area B*. It descends from 540 to 420 feet above m.s.l. and is largely confined to two bedding planes eight feet apart, at the shale contact. Differential movement occurred along both planes, making them exceptionally favorable for groundwater penetration.

The Warren discharged water into *Area D*, the "Rift." This is the largest feature in the cave, a tall narrow passage following a major fault. Water passed to the southeast along it. Roof elevation is as high as 500 feet above m.s.l., the highest places occurring directly above points of discharge from the Warren. This suggests that aggressive groundwater drove upwards with some force before moving to the southeast. The solid floor is wholly masked by a later fill, which also chokes the central section of the Rift to the roof. The higher walls show shallow pocketing, as in *Area B*. The modern free surface stream passes out of the southeast end of the Rift into a short, underfit passage and the terminal siphon.

The Warren passages are the most instructive part of the phreatic system. Figure 5 shows a reconstruction of their form as it was at the close of the first distinguishable phase in the history of the cave.

The system consists of a tiered structure of four passages (a, b, c and d) following a sinuous course close to the strike but descending gently to the southeast. Passage *a* has the highest general elevation and passage *d* the lowest, but all head in the central complex at c. 540 feet and were contemporary in development. They are linked by short sections oriented close to true dip. The orientation of the near-strike passages is sometimes determined by local joints, but for most of their length there is no structural explanation of the confined sinuous course fol-

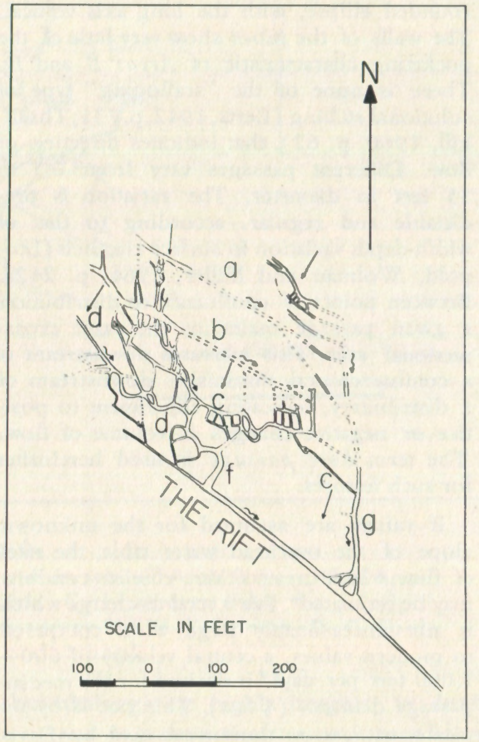

Figure 5.
Plan of the Warren bore passages, St. Cuthbert's Swallet, at the close of the first phreatic phase. Letters are keyed to the text. Straight arrows indicate direction of true dip. Curly arrows indicate direction of contemporary flow. Dashed lines show hypothetical extensions of known passages.

lowed across a steeply dipping bedding plane. These passages fed into shorter, joint-determined sections, *f* and *g*, which linked them to the Rift. *F* and *g* show the same erosional form as the bedding plane passages.

The form of these passages is remarkable, being a nearly perfectly rounded tube, sometimes centered on the guiding fracture plane, sometimes developed directly above it. Variations from the circular occur in only one situation: where a small tube is becoming confluent with a larger one, the roof of the small tube is raised and the floor lowered to attain an accordant junction. This draws the circular form out into that of a rounded ellipse, with the long axis vertical. The walls of the tubes show very little of the pocketing characteristic of *Areas B* and *D*. There is none of the "scalloping" type of solutional etching (Bertz, 1942, p. 731; Thrailkill, 1960, p. 62), that indicates direction of flow. Different passages vary from 0.5 to 15 feet in diameter. The variation is predictable and regular, according to that of width-depth variation in surface channels (Leopold, Wolman and Miller, 1964, p. 242). Between points of confluence or distribution a given passage maintains the same cross-sectional area. This increases downstream of a confluence and diminishes downstream of a distributary, indicating adjustment to positive or negative changes in volume of flow. The term *bore passage* is used hereinafter for such features.

If values are assumed for the unknown slope of the overhead water table, the rates of flow which created these efficient conduits may be estimated*. For a total discharge which is not unreasonably large when compared to modern values, a central velocity of 600 – 1,000 feet per day** is derived for the median passage diameter (4 feet). This rate is by no means as low as those envisaged by Davis (1930, p. 554) or Woodward (1961, p. 43) in the phreatic zone. The assumed gradient of the contemporary water table was 1:250, which is an exceptionally low figure for the region. It suggests that development of the bore passages was associated with a prolonged period of stability in the environment.

Following this first phase of cave development, some change in the dynamics or chemistry of the groundwater flow brought about a partial disintegration of the bore passage system and enlargement of the phreatic fissures elsewhere. The water table remained at or above 660 feet. Disintegration took the form of preferential solution directed laterally into all local fractures. The efficient, confined pipe form was destroyed. Dimensions of the disintegrated passages are very irregular, with much deep pocketing.

The phreatic disintegration phase ended when the water table fell from 660 + feet to, or a little below, its modern level of 380 feet above m.s.l.* The fall was evidently rapid, there are no traces of stillstand at intermediate levels. The vertical amplitude of the fall (at least 280 feet) is much greater than that of any in the other caves of the region. Approximately 66% of the volume of the modern cave can be attributed to the phreatic erosion.

2 *A water table influent cave: Swildon's Hole*

Swildon's Hole heads at 790 feet above m.s.l. in a valley floor at the shale-limestone contact, and 700 yards northwest of St. Cuthbert's Swallet. The explored system terminates in a siphon at 320 feet above m.s.l. Approximately 13,000 feet of cave passage is shown on figure 6.

Surface catchment is 0.48 square miles, yielding a mean flow of 0.72 c.f.s. Vigorous flooding may occur once or twice a year, when flow into the cave exceeds 10 c.f.s. for short periods of time. The water is thought to flow to the Cheddar spring, 4.125 miles distant, but it has not been possible to check this by dye testing.

The cave is largely developed in the basal 200 feet of the limestone. Angle of dip is 17°-21° in a westerly direction. Approximately one half of the passage length is guided by bedding planes. The southern side of the cave is in a major fault zone ("F", fig. 6).

*The Chezy-Manning formulae derived for standard drainage engineering were used. The writer is indebted to Dr. K. L. Murphy of the Department of Civil Engineering, McMaster University, Hamilton, Canada, for advice on their selection.

**0.007 – 0.012 feet per second.

* The modern water table is probably raised a little by vadose stream aggradation in the Rift.

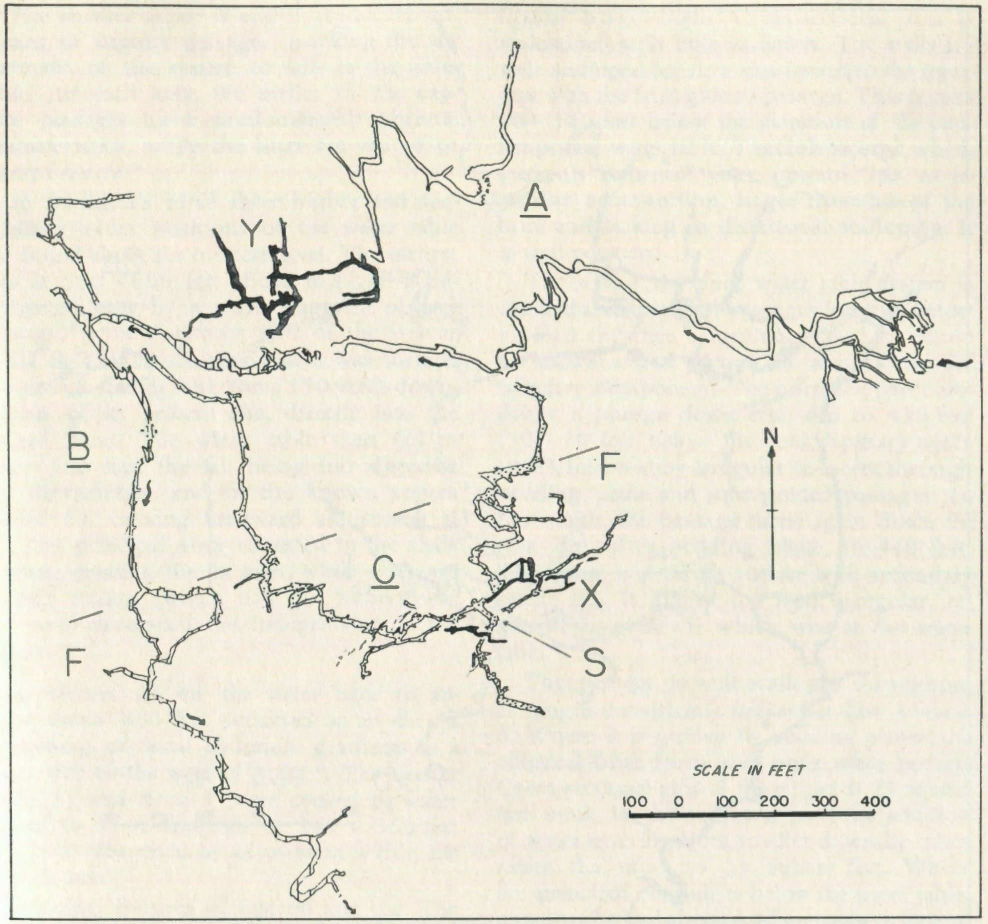

Figure 6.
Plan of Swildon's Hole. Passages of Area A in white: Area B — vertical lines: Area C — dotted: Areas X and Y in black. Curly arrows show the principal modern stream (from a C. R. G. Grade 6 survey by W. I. Stanton, D. C. Ford, C. H. Kenney and others).

Swildon's Hole has three principal morphological areas (fig. 6). *Area A* comprises two independent inlet passages that descend to 400 feet. They are predominantly vadose in character, with up to 40 feet of gravitational entrenchment beneath limited original phreatic passages. Mean gradients are 1:4 – 1:5 (sub-parallel to the underlying shales), but thalwegs are highly irregular in detail.

Area B is the active stream passage below the confluence of the vadose channels. It has the form of a "Master Cave" (Warwick, 1962, p. 52). Low arches, bore passages and bell-shaped solutional forms are mingled with gravitational trenches and walls undercut by lateral solution at floor level. The thalweg is composed of long, gentle sections of pebble and cobble shoals, interspersed with shorter, steeper trenches and *sumps* (sections that are wholly water-filled). The sumps are

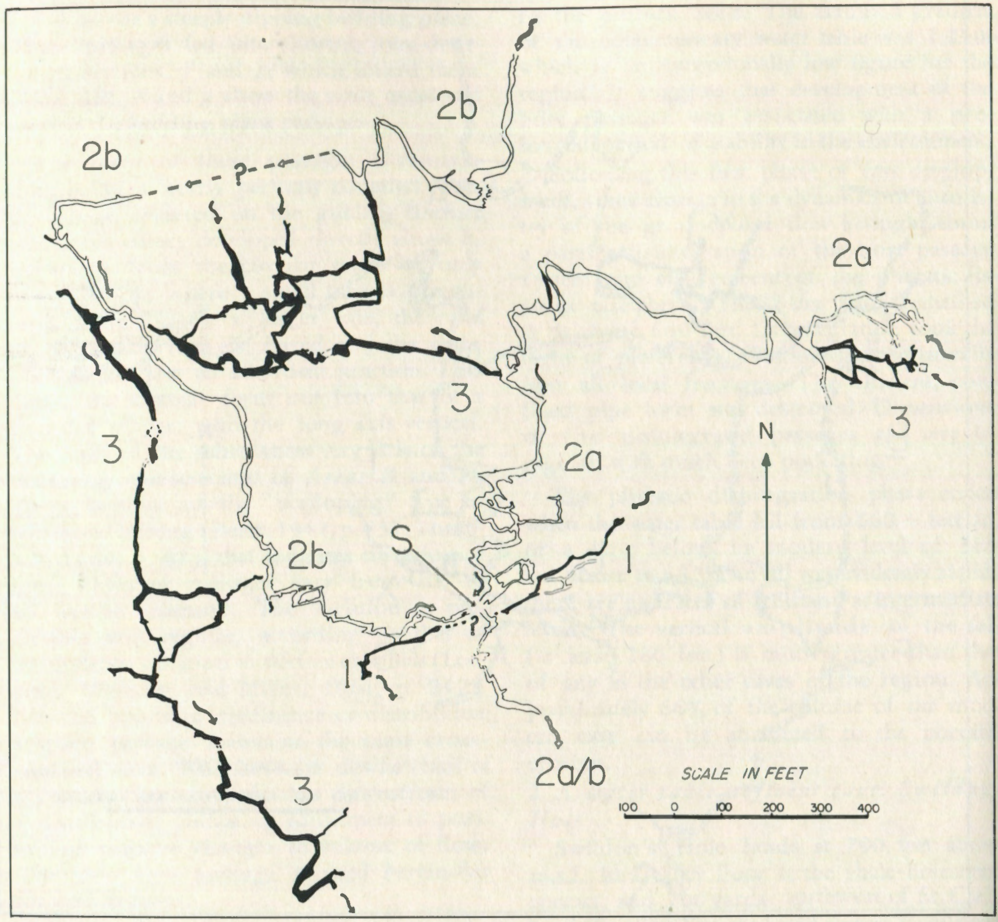

Figure 7.

The general sequence of development in Swildon's Hole. The first and third passage systems are in black, the second system in white. Curly arrows show the contemporary streams.

up to 40 feet in length and 15 feet deep. The overall gradient is 1:37. In the Master Cave area the groundwater is moving away from the shale contact. There is no alternative structural determinant of the gradient and no doubt that the stream here defines the local elevation of the water table. In flood this may rise 20 feet for a few hours.

Area C is the "abandoned cave." It comprises three passages becoming confluent at point "S", 480 feet above m.s.l. The overall gradient (descending from north to south), is also approximately 1:37, but the passages climb and descend frequently within a range of 420 – 500 feet above m.s.l. so that this general value is never realized along the constituent parts. Bore passage forms, where bedding planes guide the conduits, and well-pocketed bell shapes on joints or faults leave little doubt that erosional expansion occurred under conditions of complete or nearly complete waterfill. Floors are aggraded with a fine sand/silt/clay mixture, mantled with later collapse and flowstone.

The smaller areas, *X* and *Y*, are local complexes of capture passages, marking the adjustment of the system to falls in the water table. In each area, the earlier of the capture passages have predominantly phreatic characteristics, while the latter are vadose or *paraphreatic.**

In Swildon's Hole three higher and successively older positions of the water table are found above the modern level. The earliest was at 550 – 600 feet above m.s.l. It is represented only by a small fragment of cave passage in the southeast part of the system ("1", fig. 7). At the time that it was formed the surface stream sank some 150 yards downstream of its present site, directly into the faulted zone. The water table then fell to 480 – 500 feet, the fall being introduced at the downstream end of the known system (point *S*), causing headward adjustment at *X*. The principal sink retreated to the shale contact, creating the 2a inlet, while a second surface stream (which had not hitherto fed the cave) developed the independent 2b tributary.

A second fall of the water table to approximately 400 feet occurred as an abrupt steepening of local hydraulic gradients at a place well to the west of point *S*. The master cave (3), and *Area Y* were created by water drawn to it. A final fall to 340 – 360 feet has been absorbed by adjustment within the master cave.

Principal features of interest are: a. The definition of the water tables in the vertical plane and b. The re-organization of the plan form of the cave which occurred at the successive water table levels.

a. *definition in the vertical plane.* The 550 – 600 feet water table is known only by a loop of cave passage below it. This has two components: a lengthy passage following a single bedding plane down true dip, and, from its lowest point, a shorter passage which climbs directly up a major fault dipping at 75°. The bedding plane passage has an elliptical bore form. Cross-sectional area is maintained with little variation. The walls are well scalloped for flow downward to the intercept with the fault-guided passage. This occurs 90 – 140 feet below the elevation of the contemporary water table. The fault passage, which evidently returned water upward, has an irregular cross-section, larger than that of the bore and lacking in directional scalloping. It is well pocketed.

The 480 – 500 foot water table system is again characterized by passages looping below its level and then returning toward it. Figure 8a shows a true section of part of the 2b tributary component. The northern part comprises a plunge down true dip to 430 feet (50 – 70 feet below the contemporary water table), followed by irregular re-ascent through bedding plane and joint-guided passages. In the south, the passage turns again down the true dip of a bedding plane, to 440 feet. The plane is a thrust surface with secondary calcite fill. It guides the final, irregular, reascent to point *S*, which was at the water table.

The passage is well scalloped throughout its length consistently indicating flow toward *S*. Where it is guided by bedding planes, the elliptical bore form and none other occurs. Cross-sectional area of the ellipse is 25 square feet until, in the southern part, the addition of water entering from smaller downdip tubes raises the value to 33 square feet. Where the principal conduit is below the water table, occasional blind chimneys have been driven up cross-joints toward it. The highest chimneys close in unpenetrated rock between 480 – 500 feet, although their cross-sectional areas at the base (and hence, the volume of water that could be driven up them under uniformly distributed pressure), are very variable.

The 2a tributary, although shorter, is an almost perfect image of the 2b form. Cross-sectional area of its bore passage segments is 20 square feet.

The absence of larger particles in the floor deposits indicates that contemporary flood water was only able to lift fine sand and smaller material up the long ascending gradients of the 480 – 500 feet system. Taking the velocity required to entrain fine sand*,

* "paraphreatic" – formed under conditions of frequent alternation between wholly water-filled and drained, gravitational flow. See Tratman, 1957.

* From Hjulstrom, 1939, quoted in A. E. Scheidegger, 1961, p. 135.

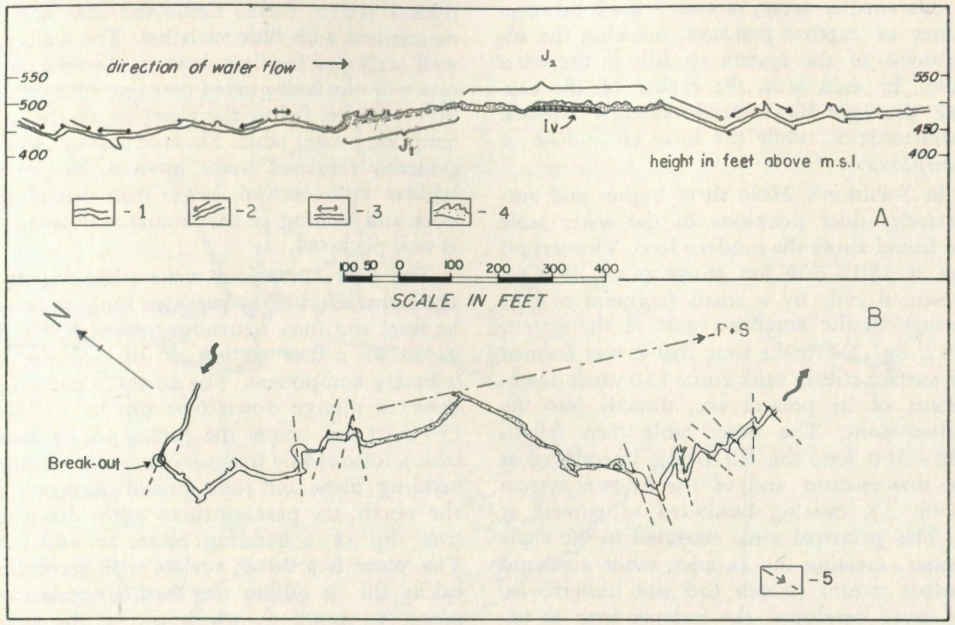

Figure 8.

A — Long section through a part of the 2b tributary system, Swildon's Hole. 1 — passage developed in bedding plane: 2 — bedding plane passage aligned down true dip: 3 — bedding plane passage aligned on true strike: 4 — passage developed on joints. Iv — isolated vadose trench.

B — plan of the passage shown in fig. 8A. 5 — location and direction of descent of independent phreatic tubes. The dash-dot line indicates the shortest course between the point of break-out and Point S.

a flow of 8 – 12 c.f.s. through point *S* (confluence of tributaries 2a and 2b), is obtained. This closely matches modern flood values. As the fine sand was probably a flood deposit, a climatic environment comparable to that of the present is suggested for the erosional expansion of the 480 – 500 foot cave. Independent evidence shows that flood amplitude was approximately 20 feet, as in the active cave today.

Joints played a more important part than bedding planes in the composition of the master cave. The characteristic looping course beneath a contemporary water table occurs, however, in places. Figure 9 shows a passage which plunged downdip to 40 feet beneath the 410 foot surface, then turned upward on a short joint to join a long, flat elliptical bore in another joint.

Adjustment of the illustrated section to the 340 foot water table has taken the form of an erosional grading that cannot be matched in the earlier systems. The elliptical bore has been entrenched to a maximum depth of 20 feet. Those parts of the down dip tube and short reascending joint below this new floor level have been infilled with coarse (pebble/-cobble) deposits and a short, direct bypass driven through the rock. No fracture of any kind guides the position of the bypass.

An important feature of this gradation is the presence of coarse fill. This could reach the head of the master cave at the 410 foot water table phase and later because the higher inlets descended to it without lengthy, sub-water table loops to act as filters. The *known* parts of the 480 – 500 foot system were protected by such headward filters, which remain choked today. Thus, when a short loop beneath the water table in the master cave became filled with coarse detritus, the next flood was able to force a more efficient bypass through unbroken rock before it had accumulated enough pressure to flush its former route. With a finer grade of fill, looping passages in the abandoned cave could always be cleared before the pressures required longer bypasses. This is why a rather flat, steadily descending profile is associated with a water table gradient of 1:37 in the master cave, while the fossil cave profile is very irregular despite its identical overall gradient.

b. *Plan form re-organization at water tables.* Two quite distinct modes of re-organization of the active hydrological system are illustrated by Swildon's Hole. First, the fall of the water table to the 480 – 500 foot level permitted the surface stream to sink at the shale/limestone contact instead of its earlier position 150 yards downstream, and independent tributary (2b) was drawn into the system*. It is of interest to show how the waters from the two new sinks found their way through the rock to the large cave already existing at point *S*.

From the sinks, the new streams opened bedding plane bore passages, oriented down true dip. Particular joints were used by the water to climb from a lower plane to a higher one, where the descent was recommenced. Passage dimensions were small: 0.5 – 1.0 square feet. True dip orientation guided the water away from point *S*, (SSW of the sinks), and carried it to lower elevations.

Confluence was attained by the waters breaking out laterally through the south walls of the downdip bores. Breakout occurred at those places closest, or nearly closest, to point *S* on the plan. Upstream of the breakout the downdip bores were ultimately expanded to 25 square feet in area; downstream they retain the 0.5 – 1.0 square foot size. The course selected between the break-out and point *S* was determined by two types of features: some large joints, favorably positioned and oriented, and other downdip tubes. The latter were independent of both the older, fault-located,

* The waters of tributaries 2a and 2b have now been combined by a surface stream capture on the shale outcrop.

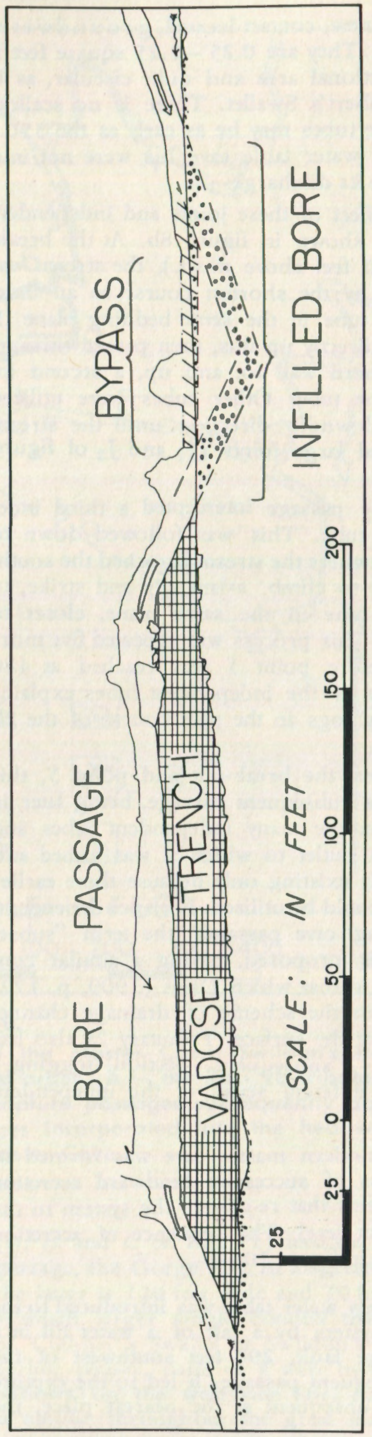

Figure 9.

Long section of part of the Master Cave, Swildon's Hole, illustrating the graduation of irregular phreatic passages to a lower water table. White arrows indicate past phreatic flow and the new, contact-located, groundwater streams. They are 0.25 – 1.25 square feet in cross-sectional area and quite circular, as in St. Cuthbert's Swallet. There is no scalloping. The tubes may be as early as the 550 – 600 feet water table cave but were not integrated to its discharge.

The effect of these joints and independent tubes is shown in figure 8b. At the break-out (430 feet above m.s.l.), the stream was directed by the shortest course to an independent tube in the same bedding plane It turned directly up this, then passed through its southern wall to, and up, a second independent tube. Other tubes were utilized in the downdip direction until the stream intersected large joints (J_1 and J_2 of figure 8b).

The J_2 passage intercepted a third independent tube. This was followed down to 440 feet, where the stream breached the southeast wall to climb, aslant dip and strike, to another tube in the same plane, closer to point *S*. This process was repeated five more times before point *S* was reached at 480 feet. Use of the independent tubes explains the many jogs in the plan course of the 2b tributary.

Between the break-out and point *S*, this is thus a subsequent passage, being later in time than the many independent tubes and the main outlet to which it was aimed and apparently existing only because these earlier courses could be utilized. For such dependent, integrating cave passages, the term "subsequent" is proposed, having a similar connotation to that which Davis (1909, p. 172) gives it in the scheme of drainage channel patterns at the surface. Tributary 2a also has a lengthy "subsequent" section, jogging a cross a different series of the independent tubes.

The modern master cave was formed by a process of successive headward accretion of segments that re-graded the system to the 410 – foot level. The sequence of accretion is shown in figure 10.

The new water table was introduced to the known system by a fall of a water fill in a very large fault, 290 feet southwest of the 2b subsequent passage. It led to the capture of the subsequent at the nearest place, the base of the loop down to 440 feet (Capture 1, figure 10). This steepened the hydraulic gradient higher up the 2b tributary, causing a second capture development which utilized the lower part of capture passage 1. The higher part was abandoned while its cross-sectional area was only four square feet. The lower part now constitutes a segment of the active master cave with a dimension of 40 – 50 square feet for expansion during the 410 foot phase.

Further headward captures occurred. Retrogressive adjustment was terminated at the downstream ends of the steep vadose inlet passages. Regrading of the latter took the form of an entrenchment, 10 – 15 feet in depth, of the floors of vadose channels that had been cut when the streams fed the earlier 480 – 500 foot system.*

In Swildon's Hole, approximately 40% of the volume of the void area can be attributed to erosion below the contemporary water tables, 40% to erosion above them, and the remainder to erosion along them.

3. *A vadose influent cave: G. B. Cave*

G. B. Cave is four miles northwest of Swildon's Hole. It is entered at 830 feet above m.s.l. at the shale contact. The lowest point is a choke of fill in the mainstream course, at 400 feet. Approximately 4,500 feet of passages are known (figure 11).

The modern surface catchment is 0.31 square mile, giving a mean flow of 0.43 c.f.s. at the sinkhole. The catchment was probably double its present size in the past. It has been reduced by recent capture in the shale outcrop. The G. B. groundwater is discharged at the Cheddar spring, 83 feet above m.s.l., and 1.53 miles to the southwest.

The cave is developed in the base of the Mississippian limestone, which is locally much disturbed by differential movement along the shale contact. Major faults and joints have been the important guides of passage development. Filling with secondary calcite is not

* See Ford, 1965, for a summary of the vadose evidence. A second entrenchment marks adjustment to the latest (340 foot) water table.

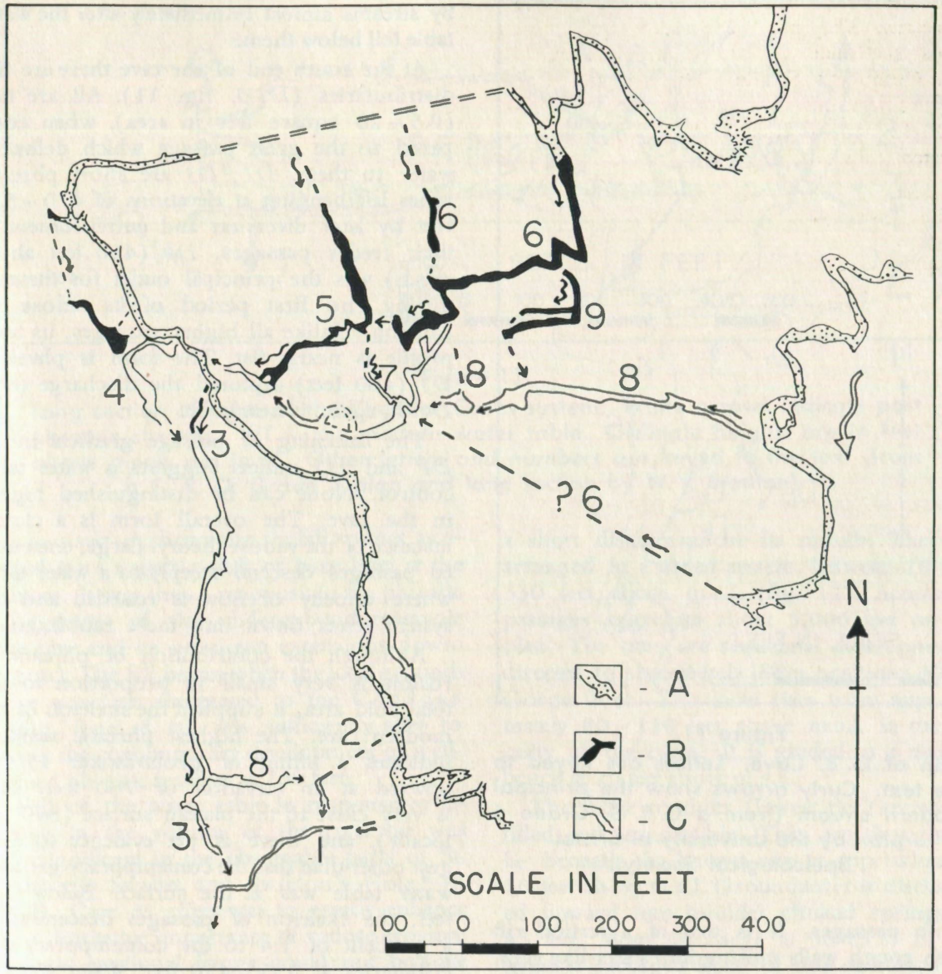

Figure 10.

The sequence of captures which created the Master Cave, Swildon's Hole. Numbers give the captures in chronological order. A — the 2a - 2b tributaries which were captured: B — segments of capture passages subsequently abandoned: C — segments of capture passages incorporated into the headward growth of the Master Cave.

common in the fracture planes. These are unusually open.

The form and pattern of the cave is comparatively simple. Ninety-five percent of its volume can be attributed to erosion and breakdown under vadose conditions. Most passages descend at steep but steady gradients to the south. Three independent tributaries (*A*, *B* and *C* of fig. 11) feed the principal passage, the Gorge (*G*). At its greatest width, the latter is 120 feet wide and 70 feet high – a much larger single conduit than any in St. Cuthbert's Swallet or Swildon's Hole. In earlier phases, parts of it were by-passed by oxbows on the west side (*01*, *02*). Form is similar throughout the great majority of

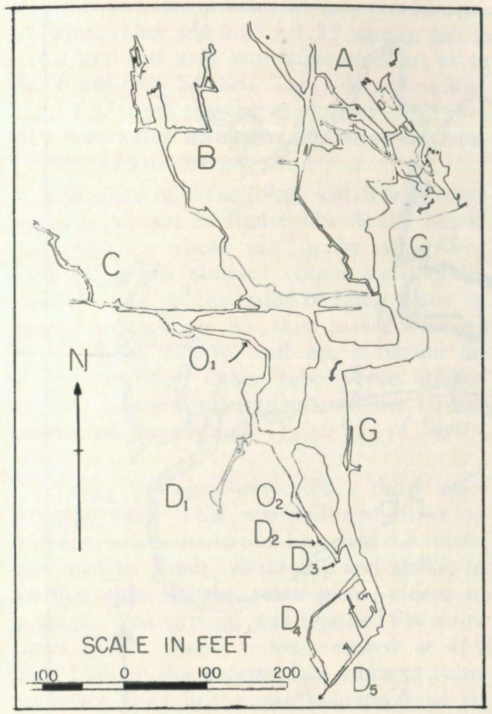

Figure 11.
Plan of G. B. Cave. Letters are keyed to the text. Curly arrows show the principal modern stream (from a C. R. G. Grade 6 plan by the University of Bristol Speleological Society).

these passages. It is that of a vertical rift with rough walls closing into a slot-like roof. The topmost two or three feet of the slot may have a distinctive smooth surface, attesting to erosion under conditions of complete water fill. But there is no pocketing. The rough walls indicate 10 – 40 feet of gravitational entrenchment down the guiding fracture. At the floor level, there has been an important measure of lateral corrasion. This has triggered a widespread regressing collapse of the walls.

Tributary *C* and the earliest passage of the *A* group are exceptional. They have a rather irregular but definite bore form, with some small pocketing and wall scalloping. There is no collapse. These phreatic passages are much smaller than the vadose rifts. They have been preserved because they were abandoned by streams almost immediately after the water table fell below them.

At the south end of the cave there are five distributaries (*D1-5*, fig. 11). All are tiny (0.5 – 20 square feet in area), when compared to the great passage which delivered water to them. *D1/D3* are short phreatic tubes left hanging at elevations of 470 – 535 feet by later diversion and entrenchment in their feeder passages. *D4* (450 feet above m.s.l.) was the principal outlet for the cave during the first period of its vadose expansion. Unlike all higher passages, its long profile is nearly flat. The form is phreatic. *D5* (400 feet) captured the discharge from *D4*. It, again, is nearly flat.

The flattening of passage gradient in the *D4* and *D5* outlets suggests a water table control. None can be distinguished higher in the cave. The overall form is a classic instance of the vadose theory. Large, entrenched passages descend steeply to a water table where velocity of flow is retarded and the system shuts down into mere rabbit holes.

Although the contribution of phreatic erosion is very small in proportion to the total void area, it supplied the skeleton of the modern cave. The highest phreatic residual indicates a lifting of groundwater 15 feet upward at an elevation of 760 feet. This is very close to the plateau surface (840 feet locally), and there is no evidence to suggest other than that the contemporary groundwater table was at the surface. Below 760 feet, the skeleton of passages descended at a gradient of 1:4 to the contemporary distributaries at 535 – 450 feet. Water was not dispersed into higher outlets. This observation implies either that (a.) the local water table possessed a gradient of 1:4 or (b.) that phreatic passages were expanded to a depth of at least 310 feet below the minimal water table level (760 feet above m.s.l.). The first alternative cannot be supported. The local rock has a denser and more open mesh of fractures than that at Swildon's Hole, where no stable water table gradient steeper than 1:37 was found. There is no supplementary evidence, such as the accordance of terminal elevation of phreatic chimneys.

The water table then fell to stabilize at a level of 450 feet (*D4*). Correlation of the phase history with that derived from the Cheddar cave system shows that the fall was not produced by a negative shift of base level at the effluent. It was thus a response to the phreatic enlargement of the underground reservoir (the cave and its unknown extensions downstream). The fall began when the local groundwater was first integrated to the Cheddar spring. Rate of fall was sufficiently slow to permit the contemporary development of well-formed phreatic features at 760 feet.

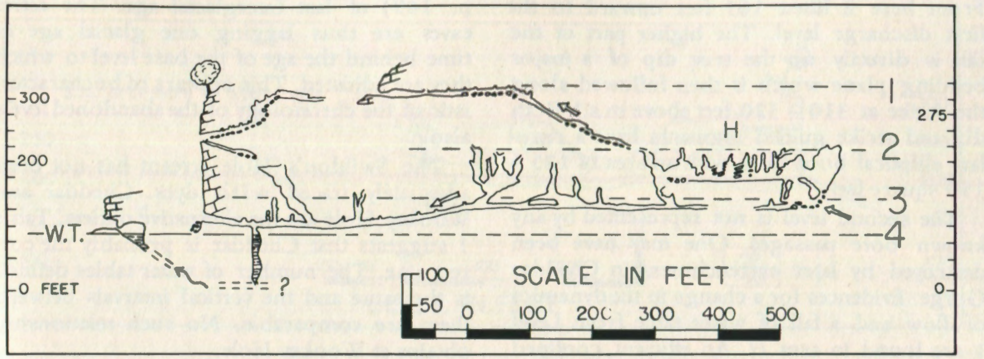

Figure 12.
Long section of parts of the Cheddar Caves system. White arrows indicate past phreatic streams. WT — the modern water table. Ordinate heights are in feet above mean sea level. Other letters and numbers are keyed to the text (from a C. R. G. Grade 6 plan and long section by W. I. Stanton).

Fall of the water table in response to increase in the volume of the reservoir was not significant in the observable parts of St. Cuthbert's Swallet and Swildon's Hole.* In G. B. Cave its effect was to greatly enhance the quantitative importance of vadose erosion. Phreatic erosional forms could not become large because the duration of the phreatic phase was comparatively short. Vadose forms are particularly large because the local joints and faults were more readily penetrated here than in the other sample caves.**

4. *The outlet caves: a. The Cheddar Caves*

The caves are located in the very steep south wall of the famous Cheddar Gorge, a short distance above its mouth. They are arranged in a tiered system, between 10 and 330 feet above m.s.l. (fig. 12). Accessible passages aggregate about 5,000 feet on the plan. The tiers are sequential developments, directed to successively lower positions of the Gorge floor. This now rises from approximately 80 – 110 feet above m.s.l. in the locality of the caves. It is graded to a marine bench at 70 feet above m.s.l..

The active conduits (lowest tier) are water-filled and inaccessible. They are thought to lie beneath the known cave at approximately 10 feet above m.s.l. Groundwater is discharged upward into boulder choked springs at 83 feet. Mean discharge is 36 c.f.s. In the greatest recorded flood of this century, flow was 166 c.f.s. and the water table rose 25 feet to fill the lowest parts of the abandoned caves. In low water conditions the water table is flat.

Volumes of flow are thus much higher at Cheddar than in the known feeder caves. A proportionate increase in the size of the erosional forms may be expected. This is not found in general *e.g.* cave rooms formed at points of collapse are not as large as in G. B. Cave), but does occur in the case of bore passages.

The caves are predominantly phreatic in form. The earliest water was drawn down tension joints into comparatively open bedding planes at the head of the caves (*H*, fig. 12).

* In Swildon's Hole the low water elevation of the table at point *S* probably fell from c. 495 feet to the quoted figure (480 feet) in response to enlargement of the reservoir.

**The genesis of G. B. Cave is discussed in detail by Ford, 1964a.

From here it lifted 165 feet upward to the first discharge level. The higher part of the lift is directly up the true dip of a major bedding plane which is then followed along the strike at 310 – 320 feet above m.s.l. Both dip and strike guided segments have a regular, elliptical bore form with an area of 125 – 150 square feet.

The second level is not represented by any known bore passages. One may have been destroyed by later entrenchment in Cheddar Gorge. Evidences for a change in the dynamics of flow and a fall of water table from Level 1 are found in area *H*. An efficient, confined bore passage there, which fed the original phreatic lift, was disintegrated into a honeycomb of widely anastomosed bedding plane voids and deeply pocketed chimneys. The chimneys are the product of solutional water attacking from below. Despite large variations in the elevation of their bases, their cross-sectional dimensions, and the magnitude of the fractures that they follow, all close within a range of 200 – 225 feet above m.s.l. This is taken as the position of a second abandoned water table.

A finely formed bore passage defines a third abandoned level. The passage has an elliptical cross-section, with an area of 260 – 290 square feet. Its elevation varies between 90 – 120 feet at the roof. It is consistently scalloped for flow from area *H*. There is little pocketing within the bore passage but associated chimneys, drilled up every cross-joint in the roof, are deeply pocketed.

TABLE 1
MARINE BENCHES AND WATER TABLE ELEVATIONS
(Heights in feet above m. s. l.)

Marine benches	Cheddar Caves	Swildon's Hole
310 - 340	310 - 330	550 - 500
220 - 240	200 - 225	480 - 500
120 - 130	135	410
70	83	340 - 360

Table I shows the relationship between defined water tables at the Cheddar effluent and the marine benches along the south flank of the Mendip Hills. There can be little doubt that these determined water tables in the caves, through the intervening agency of periodic entrenchment of the Gorge floor by periglacial surface streams. The 70 – foot bench may be the "Main Monastirian" (Zeuner, 1959, p. 149) of last interglacial age. The active caves are thus lagging one glacial age in time behind the age of the base level to which they are adjusted. This appears to be characteristic of the chronology of the abandoned levels also.

The Swildon's Hole stream has not been adequately traced with dyes. Cheddar and Wookey Hole are the alternative outlets. Table 1 suggests that Cheddar is probably the correct one. The number of water tables defined is the same and the vertical intervals between them are comparable. No such relationship obtains at Wookey Hole.

b. *Wookey Hole*

This final cave is again distinct. It is an active effluent, entered through a small gorge created by regressive collapse at the cave mouth. The underground stream can ordinarily be followed for 600 feet upstream of the resurgence (fig. 13). The cave ceiling is flooded at this point. Divers have followed the channel for a further 800 feet. The range of elevation is 150 – 290 feet above m.s.l. The water table is at 200 feet and is nearly flat under mean conditions. Mean discharge is 22 c.f.s.

The accessible cave is developed in dolomitic conglomerate (Permo-Triassic), filling a valley in the Mississippian limestone. The conglomerate is composed of limestone fragments in size from pebbles to small boulders, with a red calcareous silt matrix that composes 50 – 80% of the volume. It is regularly bedded, and fractured by a large vertical fault that parallels the trend of the cave. As at Cheddar, the principal discharge conduits tend to follow bedding planes, developing large blind avens by erosion upward where the main fault or lesser fractures are touched. Erosional forms in the dolomitic conglomerate are the same as those created in a similar geomorphic environment in the Mississippian limestone (*i.e.,* at Cheddar). Smooth solutional facets cut across boulder and matrix without notable differentiation.

Although it now constitutes a water table cave, the significant erosional expansion of Wookey Hole occurred under phreatic conditions. The cave was first opened when groundwater leaked round an impermeable dam, which had impounded it at a much higher level. Early flow was thus driven by

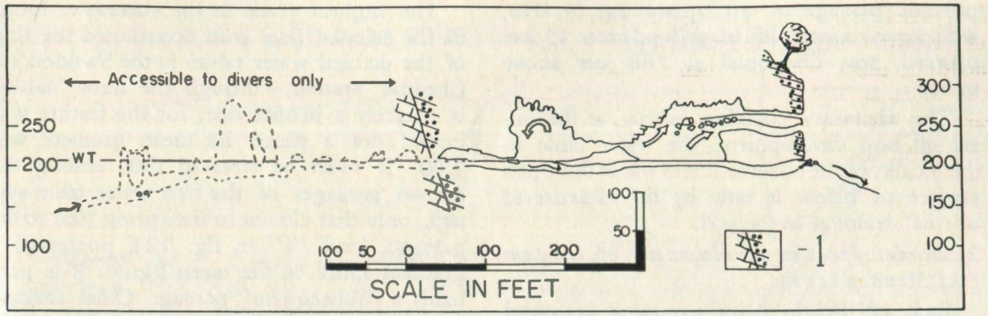

Figure 13.
Long section of Wookey Hole. 1 — approximate position of the limestone-dolomitic conglomerate contact. White arrows show past phreatic streams. WT — the modern water table. Ordinate heights are in feet above mean sea level (based on plans by J. W. Duck and the Cave Diving Group of Great Britain).

an exceptionally steep pressure gradient. The cave has a regular form, climbing upward to the spring from its furthest known point. On reaching the well-fractured conglomerate, the principal conduit was divided into a series of distributaries, ranging in elevation from 190 – 260 feet. Distributaries at 250 – 260 feet became the dominant conduits and define a first water table. Flood surge in the early constricted cave was able to drive blind chimneys upward to 290 feet (*i.e.* the known vertical amplitude of phreatic ascent is 140 feet). As at Cheddar, the avens are deeply pocketed while the bedding plane passages which fed water to them are only shallowly indented, it at all, but well scalloped. A negative shift of the external base level later permitted distributaries at 200 – 210 feet above m.s.l. to take all of the flow and develop the modern water table.

DISCUSSION

1. *Conditions at cave origin and early development.*

As footnoted on page 110, various workers have distinguished conditions of "cave initiation" or "origin" and "cave development".* The principle is that, at the moment that "cave development" commences, the groundwaters have a widespread mesh of minutely open, randomly distributed, solutional courses to choose from. The mesh constitutes the first phase of the cave history. Many variables of the erosional environment may be quite different from those which develop the later cave. Some authorities hold that the initial courses will be wholly water filled (*i.e.* all limestone caves have a phreatic origin), and others consider that a water table may already be established at depth in the rock.**

* Warwick, 1962, p. 72; Davies, 1960, p. 16; Woodward, 1961, p. 42.

**Davies, 1960, p. 17; Woodward, 1961, p. 53.

White and Longyear (1962, p. 165) have suggested a value which supplies a useful quantitative measure of the change from initial to developed conditions. When a diameter or width of the order 5.0 mm. is obtained in a groundwater conduit, turbulent flow may commence. The rate of solution is greatly increased as a result. 5.0 mm. may therefore be taken as the threshold value for cave "development."

In the central Mendip examples, there is no doubt that the conditions of cave origin *and* early development were wholly phreatic. All early conduits evidence expansion beneath a water table to dimensions many times greater than the threshold value. The extreme illustration is G. B. Cave – the most "vadose" in the sample. The straight-line gradient from the earliest stream sink to the contemporary spring was as steep as 1:12. Despite this and the exceptionally open fracture situation, a phreatic passage of 20 square feet in area, with an associated lift of groundwater 15 feet upward, was developed at 760 feet above m.s.l.

The necessary corollary is that, at the onset of cave development, the water table in the locality of the observed caves was at the open surface or below it only by the measure of normal drainage in the soil.

2. *Model for the development of the central Mendip caves.*

Each of the three influent caves appeared to exemplify one of the three conflicting general theories of cavern development (page 109). But the evidence described above indicates that all of the accessible caverns are parts of a single model of development, which is complex because there have been successive negative shifts of the allogenic base level.

Because it has the most extensive passageways at both inlet and outlet ends, the Swildon's Hole – to – Cheddar cave system best illustrates the model. What is known of the earliest development is a descending bore passage and fault ascent at the swallet and a similar deep loop at Cheddar. The nature of the later 2a and 2b tributaries and early master cave at Swildon's show that it was by many such loops across the dipping strata that a principal groundwater stream made its way to the spring. Co-existent were the small tubes utilized as "targets" during the development of the subsequent passages. They conveyed independent streams far below the local elevation of the first principal stream.

Because they are known at and below this elevation, it is supposed that similar tubes existed above it. In G. B. Cave, the *D1 – D3* distributaries are instances. With enlargement by solution, air entered the head of the system, depriving these highest passages of the hydrostatic pressure needed to drive the phreatic lifts in their courses. The lowest conduits would normally have had the longest flow paths, and hence, the greatest aggregate friction*. Those conduits which were centrally placed in the vertical range of the developed tubes and/or most directly aligned upon the effluent in plan view, were thus preferentially selected and expanded to comprise the accessible caves.

* As in Swinnerton's model, 1932, p. 675.

The highest place in the successive loops of the selected flow path constituted the first of the defined water tables in the Swildon's – Cheddar system, – though the term "table" is scarcely a proper one, for the feature was linear, not a plane. Its mean gradient was 1:80. It must be stressed that, among the known passages of the first water table system, only that closest to the spring (the strike passage, level "1" in fig. 12), possessed a gradient close to the mean figure. It is probably a "subsequent" passage. Other passage segments are not subsequents, and their gradients are always much steeper, determined by the local dip of the bedding or being near-vertical ascents of joints and faults.

At St. Cuthbert's Swallet, major faults guided the turbulent groundwater flow to particularly penetrable bedding planes at the shale contact. These discharged it to the Rift, whence, it now appears most likely, the water made a re-ascent (phreatic lift) through some undiscovered chimney. The known phreatic cave, comprising more than one mile of meshed passages on the plan, thus constitutes a single loop to 280 feet or more beneath the contemporary water table.

In G. B. Cave, the critical point is that, in all save its lowest parts, the observer is in that zone through which the air-water contact fell (with expansion of the reservoir), *before* a particular central line of tubes was selected to define the water table. The zone is 380 feet in depth: development of G. B. Cave did not commence until base level at the spring had fallen as low as 120 feet above m.s.l. St. Cuthbert's Swallet and Swildon's Hole are earlier systems, developed to higher allogenic base levels, so that the equivalent zone is much shallower.

Figure 14 illustrates characteristics of the central Mendip model for the case of two successive water tables. Only the principal flow paths are drawn. Features are:–

i. restriction of the model to two simple groundwater zones (vadose and phreatic), at any given time. As Corbel has noted for other regions of Temperate Oceanic climate, Cvijic's concept of a central zone of alternating free air and water fill is not applicable in any broad picture because cave systems are adapted to clear flood water with only a small rise in the head.

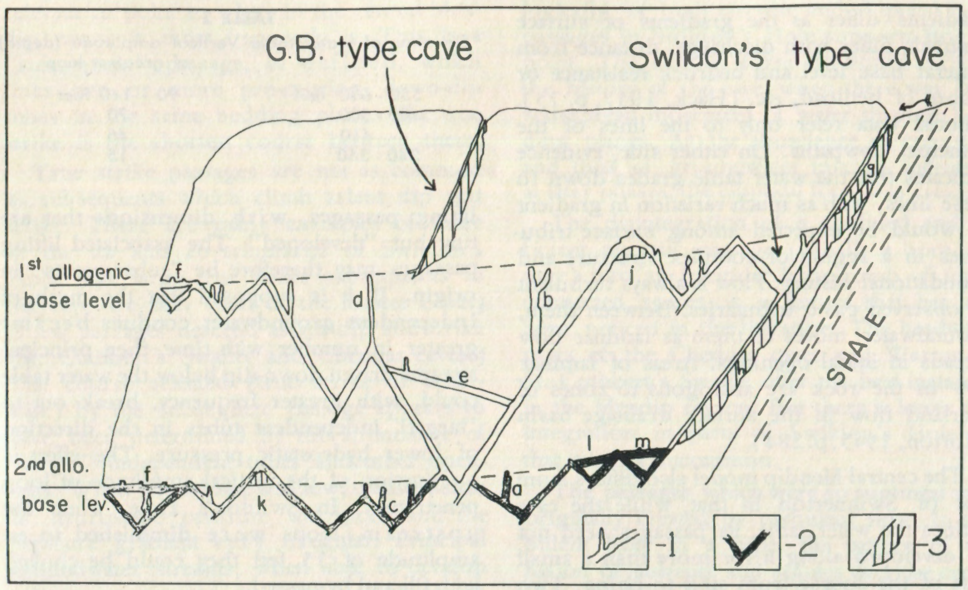

Figure 14.

Diagrammatic long section to illustrate the proposed model, drawn for the case of two successive water tables. Only principal flowpaths are shown. 1 — phreatic passages of the first water table: 2 — phreatic passages of the second water table: 3 — vadose passages. Letters are keyed to the text. The vertical scale is exaggerated.

ii. the predominance of sub-water table loops in the flow path.

iii. Figure 14, a, b, c, and d, blind chimneys driven upward by solution, to close at the water table.

iv. Figure 14, d: it is not necessary for all loops to ascend to the water table. For long distances over the principal flow path, an air-water contact may only occur at the tops of blind chimneys, which are backwaters.

v. Figure 14, e and f: "subsequent" passages, short-circuiting upward or downward loops at places where the frequency of independent tubes permits their development.

vi. Figure 14, g and h: vadose trenches descending from the surface sink to the first water-filled place in the flow path. Trench h, graded to the lower water table, is regressing through trench g with a nick-point.

vii. Figure 14, j and k: isolated vadose trenches cut through crests of upward loops which reach to the air-water contact. The trench of figure 8A, Swildon's Hole, is an example.

viii. Figure 14, l and m: bypass passages (page 121), short-circuiting detritus-choked sub-water table loops in the upstream parts of the flow path.

The model is coincident with views of Swinnerton (1932) and White and Longyear (1962). Within considerable limitations it may be said that the large caves are accordant to particular water tables. But the latter do not precede them in the cave zone. Each cave (or principal flow path) determines its own water table gradient. There is great variation within this small region of rather uniform limestone characteristics. Mean water table gradient between Swildon's Hole and Cheddar is 1:80, St. Cuthbert's – Wookey Hole is 1:40 and G. B. Cave – Cheddar, 1:19. Such variation is not compatible with any concept of a regional water table at depth in the rock in the cave initiation stage.

Variation in gradient reflects: (i.) differing mean volumes of flow into the different caves, (ii.) the distance and difference in height between surface sink and rising, (iii.) structural factors. In other words, the water table gradients differ as the gradients of surface channels differ with discharge, distance from general base level and bedrock resistance or calibre of bedload, etc. (Hack, 1957, p. 75). Quoted data refer only to the lines of the principal flowpaths. On either side, evidence indicates that the water table grades down to these lines, with as much variation in gradient as would be expected among surface tributaries in a region of complex structure and denudational history. Flow is always turbulent in observed cave tributaries. Between them, groundwater moves to them as laminar flow threads in initial openings. Areas of laminar flow in the rock are analogous to zones of overland flow in the surface drainage basin (Horton, 1945, p. 284).

The central Mendip model also differs from that of Swinnerton in that, while the cave defines the water table, its passages need not be developed along it for more than a small part of the course from sink to rising. Two factors are involved here: first is the abundance of early independent conduits from which the principal flow path is selected. In the sample caves abundance was not very great. An appropriate example is the sequence of headward captures which composed the master cave in Swildon's Hole. The addition of each new capture steepened the local hydrostatic head. Capture 6 (fig. 10) has a phreatic lift of 35 feet vertically upward in its central parts. The lifting waters were directed into capture 5, which lay only 80 feet away from the base of the lift and a little *below* its elevation. But integration over this short route, eliminating the lift, did not occur for a long while. If there had been more independent couduits the principal flow paths (with many "subsequent" segments in their composition), would have been altogether flatter.

The second factor follows from this point, illustrates it, and is dependent on time. Table 2 quotes the greatest vertical amplitude of phreatic loops below the successive water tables of Swildon's Hole. Loop penetration became shallower with time.

Each loop has two components: a descending down-dip tube in the bedding, and a lifting (returning) component which usually follows a joint or fault. Many of the dip tubes carry on downward beyond the bases of the

TABLE 2

Elevation of water table	Vertical amplitude (depth) of greatest loop
550 - 600 feet	90 - 140 feet
480 - 500	70
410	45
340 - 360	15

lifting passages, with dimensions that are tiny but "developed". The associated lifting passages may therefore be "subsequents" in origin. If it is supposed that the mesh of independent groundwater conduits became greater in number with time, then principal streams driven down-dip below the water table could, with greater frequency, break out to "target" independent tubes in the direction of lower hydrostatic pressure. The effect is a reduction of the vertical amplitude of loop penetration. In Swildon's Hole, when the upstream loops were diminished to an amplitude of 15 feet they could be choked with fill and by-passed.

Hence, in the central Mendip caves, the Swinnerton water table passage is only created after a lengthy previous history of "developed" expansion with large sub-water table caves. The best development (most uniform gradient) is restricted to upstream parts which receive coarse stream fill from feeder caves in the vadose zone.

It is apparent that many features of the central Mendip model result from the fact that, in the overall course from sink to rising, the karst groundwater must follow a line close to the bearing of true dip in steeply dipping, well-bedded rocks. In four of the five caves almost all of the descending phreatic streams followed bedding planes and may be considered cases of artesian development (Glennie, 1954), the flow being trapped by an impermeable limestone bed above. The expanded bedding planes usually have a prominent shale parting, secondary calcite or chert fill or evidence of differential slippage, which has greatly increased their permeability.

3. *The strike passage and its variants.*

The phreatic cave passage guided by a bedding plane and oriented along its true strike has received particular attention in recent literature (White, 1960). In verbal discussion it has sometimes been considered an infallible index of the water table. It is

relevant to point to what, in the central Mendip caves, it most frequently is. This is a "subsequent" passage, as defined, which links two or more pre-existing, down-dip tubes in the same bedding plane. The true strike is the shortest course between them.

True strike passages are not as common as subsequents which climb aslant dip and strike. There are many examples of these in the 2a and 2b tributaries of Swildon's Hole. The large true strike subsequents in the cave are later (part of the Master Cave.) A few passages descend aslant dip and strike (St. Cuthbert's Swallet and Cheddar caves.) The course selected (true strike or otherwise) by the subsequent passage appears to have been determined by the abundance of "target" independent tubes and other openings. When abundance was low, transmission of hydrostatic pressure was poor and the pressure gradient very irregular. Principal groundwater streams, when able to develop subsequent courses, tended to climb across dip and strike, aimed directly at points of markedly lower pressure. With increase in the abundance of target tubes, the pressure gradient became much more regular and shallow. The true strike, or "shortest distance", course was then selected by the subsequents.

4. *Bore passages and disintegration in the phreatic zone.*

In the central Mendip caves it has been found that those passages defined as bore passages vary regularly and closely in their dimensions, apparently in accordance with the variation (in place) of volume of phreatic groundwater flow.

The bore passages have either a circular or an elliptical cross-section. White and Longyear (1962, page 162) have suggested that the ellipse is the stable shape where turbulent flow occurs in a bedding plane in the phreatic zone. On the contrary, Bögli (1964, page 89) argues that the ellipse is the product of excessive solution occurring at the point where water moving everywhere through the bedding plane (in minute channels) mixes with that moving through major conduits. The "mixing corrosion" thus effects a distortion of the form of the major conduits, which would be circular if they were entirely produced by water fed in at the upstream ends. The distribution, in place and chronologically, of both circular and elliptical bore passages in Swildon's Hole supports Bögli's point. The circular form developed early in the history of the cave, when there was little widespread movement of water through bedding planes. As the abundance of conduits, and thus of groundwater mixing, increased at later phases, the elliptical form was created.*

The disintegration of a confined and efficient phreatic passage (usually a bore passage) into an irregular honeycomb of interconnected spaces is a feature that has not been noticed in the literature. The headward parts of the Cheddar caves and Warren of St. Cuthbert's Swallet offer the best instances in the Mendip region. But there is lesser disintegration in parts of Swildon's Hole, so that it is not uncommon.

The passages which were so ruptured were originally created by turbulent flow. In turbulent flow in a pipe, the pattern of diffusion of solvents and solutes over the cross-section will be dictated by the central area of most rapid flow. The shape of this tends to the circular. When a turbulent flow in a pipe is allowed to decelerate, it changes to a non-linear laminar flow at a much higher velocity than that at which the opposite change occurs during acceleration (King, Wisler and Woodburn, 1956). In laminar flow, the pattern of diffusion is determined by the shape of the extant boundary walls.

It is suggested that reversion to a non-linear laminar flow, perhaps abetted by the increasing significance of mixing corrosion as the noted phreatic caves aged, was the mechanism of disintegration. Unless the bore form was perfected under the turbulent regime, leaving no opening along the guiding structural line, non-linear laminar flow would expand the opening and destroy the efficient passage form.

In St. Cuthbert's Swallet, deceleration to a non-linear laminar flow may have been

* In many cases (but not in the central Mendips) it seems likely that widespread movement of groundwater in bedding planes was achieved in the initial phase. Thus the earliest developed bore passages would have the elliptical form. Note also that Thrailkill (1960, page 62) describes circular forms that were created quite late in the phreatic history of a Colorado cave.

produced by the cave expansion. If this lowered the water table at the head of the system faster than erosion lowered it at the outlet, loss of headfall would reduce the velocity of flow. The exceptionally low water table gradient over the bore passages (page 127) supports the idea of lowering at the head.

At Cheddar, the critical reduction of flow was probably a product of climatic change. Evidence of such change during the early phases is well preserved in these caves. It may have been a factor at St. Cuthbert's Swallet.

5. *The comparative ages of the swallet caves.*

In the central Mendip Hills the base level advantage has always been to the west (toward the general base level of the Bristol Channel.) At the surface, this has enabled the Cheddar Gorge to capture the headwaters of its easterly neighbor, Ebbor. But underground, the earliest distinguishable phase of development in St. Cuthbert's Swallet is substantially older than the equivalent in Swildon's Hole which, in turn, is older than G. B. Cave, *i.e.* from east to west the swallet caves become younger. Several factors may have contributed to this anomaly.

St. Cuthbert's Swallet has had the largest catchment basin at the surface while G. B. Cave has the smallest modern catchment. It is tempting to propose simply that the greater the volume of water available at the head of the underground system, the sooner will the transition from cave origin to cave development take place. However, during its earliest phases of development, Swildon's Hole had a catchment smaller than that of the modern G. B. The latter has had its catchment greatly reduced by surface stream capture in the shales.

In the central Mendip region, no swallet cave appears to have developed until the straight-line gradient between the sink and the outlet became steeper than 1:100. In these terms, it is easy to understand why St. Cuthbert's Swallet is older than nearby Swildon's Hole. Its outlet (Ebbor – Wookey Hole) is much closer. G. B. Cave is yet closer to the Cheddar outlet, with its noted base level advantage.

A third factor is that of the thalweg gradients of the valleys in which the cave sinkholes are located. These valleys are now dry below the sinks but they are well preserved. While the St. Cuthbert's Swallet stream exploits the Ebbor headfall underground, its *surface* valley is part of the Cheddar headwater. Thalweg gradient is c. 20 feet per mile and can never have been much greater, because there is insufficient relief available below the plateau erosion surface. The same thalweg gradient occurs over Swildon's Hole. But, because of the proximity of the deeply entrenched lower section of Cheddar Gorge, thalweg gradient at the G. B. sink is 120 feet per mile. Eighty percent of the dolines in the region are aligned along dry valley floors where, in instances of approximately equal discharge, frequency of occurence is inversely proportional to the thalweg gradient (Ford, 1964b). Both the swallet caves and the dolines are the products of solvent water attacking downward into the rock.

Integrating these factors: St. Cuthbert's Swallet is older than Swildon's Hole because it had a larger catchment and a great headfall advantage underground: Swildon's Hole is older than G. B. Cave because its overground drainage was less efficient (velocity of flow was lower), permitting more water to leak into the rock within a given time.

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Semidiurnal Movement Along a Bedrock Joint in Wool Hollow Cave, California

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ABSTRACT

A 22-hour record of strain has been obtained from three mutually perpendicular transducing cells across a joint in the wall of Wool Hollow Cave, Calaveras County, California, in the hope of determining the causes of strain. The rock cover at the instrument station is 25 meters thick, and the air temperature at the instruments ranged from 11.25-11.42° C during the observations. The joint, which strikes N. 50° E. and dips 27° NW, cuts recent travertine as well as the bedrock. The slippage showed two maxima and two minima roughly coincident in time with the theoretical earth tides. At high tide the hanging wall of the joint moved southeastward up the dip 0.4 micron with respect to the footwall, and just before high tide, 0.6 micron northeastward along the strike of the joint. A seismic disturbance with an amplitude of 0.01 micron from the Moluccan earthquake of March 21, 1965, was superimposed on the slower tidal fluctuations during a five-hour period, and it appears to have caused abrupt deflections as large as 0.03 micron, indicated by steps on the strain record.

INTRODUCTION

Joints are rock fractures along which, in contrast with faults, no visible displacement has occurred. Nearly all rocks are jointed, and in bedded rocks the joints are generally almost perpendicular to the bedding. The spacing of the joints is dependent on the rock type, and in limestone the typical spacing averages about one or two meters. When a jointed rock mass is deformed, most of the strain may occur by movement along the fractures rather than within the rock between them. Several kinds of natural phenomena can cause temporary displacements along joints. The present study was designed to measure the magnitude and direction of the displacement caused by one of these, namely earth tides.

Earth tides, like those that affect the ocean, result from differences in the gravitational attraction of the moon, and to a lesser extent the sun, at various places on the earth. These differences result from differences in distance between the attracting body and the various parts of the earth. The tides are semidiurnal, that is, there are two complete tidal cycles a day. On the moon side of the earth, the moon exerts a greater than average upward pull and therefore causes a tidal bulge. On the other side of the earth, the moon exerts a *less* than average *downward* pull and therefore causes a second bulge to form. As the earth rotates under the moon, these bulges produce two daily high earth tides separated by low tides. The tides have horizontal components of displacement as well as vertical ones.

Two chief considerations led to the selection of a cave for this study. First, the relatively constant temperature underground minimizes the deformation due to temperature change, at the surface this would be far larger than the deformation we were attempting to measure. Second, a cave, unlike a mine, has relatively stable walls. In a mine, adjustment of the wallrock to the presence of the openings continues for many decades after the mine has been opened.

A major part of the success of this study was due to the expert assistance of David R. McClurg, who also supplied much of the essential equipment. William R. Normark and Gerald N. Davis helped install and service the instruments, and Ellis Hedlund mapped the cave. The cave is on the property of the Calaveras Cement Company, and we are grateful to the officers of the company for permitting the observations to be made. The manuscript was read by Frank C. Calkins and Richard Blank of the U. S. Geological Survey, and by Bruce A. Bolt of the University of California, Berkeley.

Setting

The study was conducted in Wool Hollow Cave (fig. 1), which lies at 38.1° N., 120.4°W., about 5 kilometers northeast of the town of Vallecito, Calaveras County, California. It is in limestone of the Calaveras Formation of Permian age (Clark, 1964), which there strikes N. 80° E. and dips 80° S. Several parallel joints are exposed in the walls of the cave, and the one selected for the investigation strikes N. 50° E. and dips 27° NW (fig. 2). This joint cuts both walls of a rather narrow room located 35 meters from the entrance to the cave and 25 meters below the surface of the ground, and the instruments were placed on the north wall, which bears the joint's downdip trace (fig. 3). A small amount of efflorescent material occurs along the joint in places on this wall, but on the opposite wall a thick welt of travertine has formed along the updip trace of the joint. The joint cuts the travertine as well as the bedrock, but neither is visibly displaced.

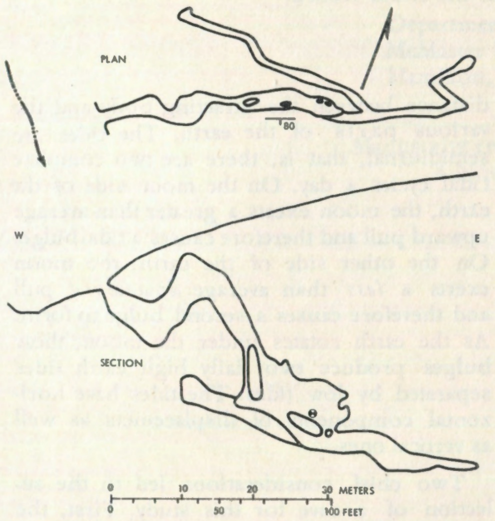

Figure 1.
Plan and section of Wool Hollow Cave, Calaveras County, California. The circular symbol on the section marks the instrument station.

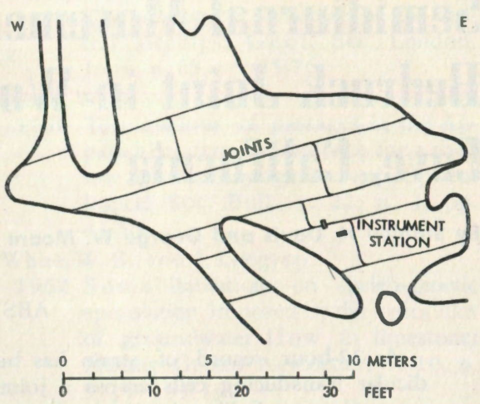

Figure 2.
Instrument station and joints within Wool Hollow Cave.

Temperature

The temperature was measured by four Yellow Springs model 401A thermistors. The scale on the readout instrument (fig. 4) has graduations equal to 0.1°C, and temperature readings were estimated to the nearest 0.01°C, but the precision was probably no better than 0.02°C. Temperature was also measured with a mercury thermometer placed in a corner of the cave about five meters from the other instruments. This thermometer could be read with a precision of about $\pm$0.05°C.

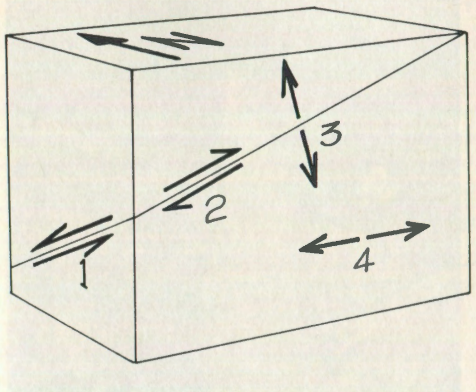

Figure 3.
The joint studied in Wool Hollow Cave and a diagrammatic representation of the movements measured by transducing cells 1-4.

The temperature fluctuation was similar at all the thermistors. High readings corresponded in part to times of greatest human activity in the cave. Although a sheet of house insulation protected the instruments against direct transfer of heat by radiation, the temperature rose about 0.05°C for each person that remained in the instrument room for more than half an hour. The uppermost thermistor was about 30 cm above the main instrument station and the lowest about 150 cm below it. A distinct temperature stratification was observable at all times, the temperature at the upper thermistor always was from 0.3 to 0.4°C above that at the lower one. The total range in temperature was from 11.31-11.59°C in the uppermost thermistor to 10.97-11.41°C in the lowest, and readings from the other thermistors were intermediate between these values. The mercury thermometer registered temperatures of 12.00-12.10°C during the period of the observations.

All the instruments showed a diurnal temperature fluctuation almost exactly in phase with the outside temperature. As the rock cover is too thick for the outside temperature variations to have been transmitted through it, the variations were probably due to convective air movements from the surface. We could feel light breezes in the constricted passages toward the entrance from the instrument room.

Figure 4.
Temperature-readout meter above, and strain-readout meter below. The instrument station is under the sheet of insulation at top center. All photographs by Ellis Hedlund.

Strain

The strain measurements were accomplished by anchoring L-shaped metal brackets on either side of the joint and determining the movements between the brackets by means of three mutually perpendicular transducing cells (fig. 5). The brackets were anchored by means of screws which served to expand lead fillers placed within 1- x 3-cm holes drilled into the rock. An additional transducer, attached to a pair of brackets about 20 cm from those spanning the joint, measured expansion and contraction of unfractured rock.

The counting brackets on the cells (fig. 5) were attached by common C clamps (fig.

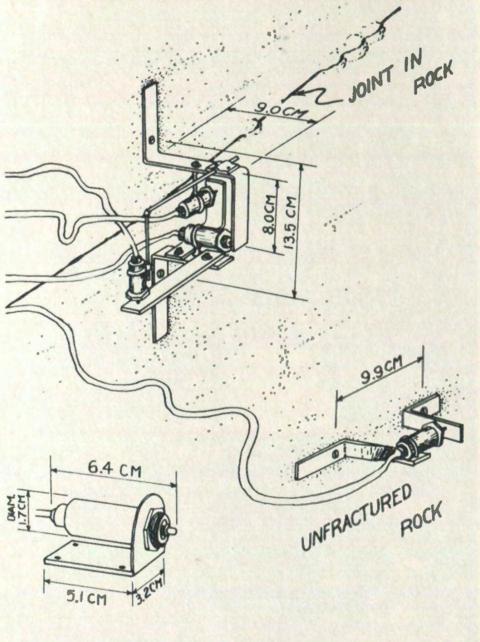

Figure 5.
The transducing cells and brackets in relation to the joint on the cave wall. The cell holders are diagramatic. A cell and its mounting blade are shown in the lower-left inset.

6). The moving tips of the cells were placed in compression before the clamps were tightened. With this arrangement, the cells could measure a total range of movement of about 25 microns (0.025 mm).

The transducers used were Statham strain gages in which the diameter and electrical resistance of a stretched wire changes when its length changes. The strain indicated by three of the cells was measured by a Baldwin model SR-4 strain-readout meter, which uses the wheatstone-bridge principle. The readings from the cells were recorded about every 15 minutes. Laboratory calibration indicated that 100 scale divisions of the meter were equivalent to a displacement of the cell tip of 0.12 micron. The instrument could be read to ± 3 scale divisions, giving a precision of roughly ± 0.003 micron.

Output voltage from the fourth cell was recorded directly on a Varian model G-40

Figure 6.
The instrument station in Wool Hollow Cave.

Figure 7.
Recorder which traced continuous record of east-west strain on the joint.

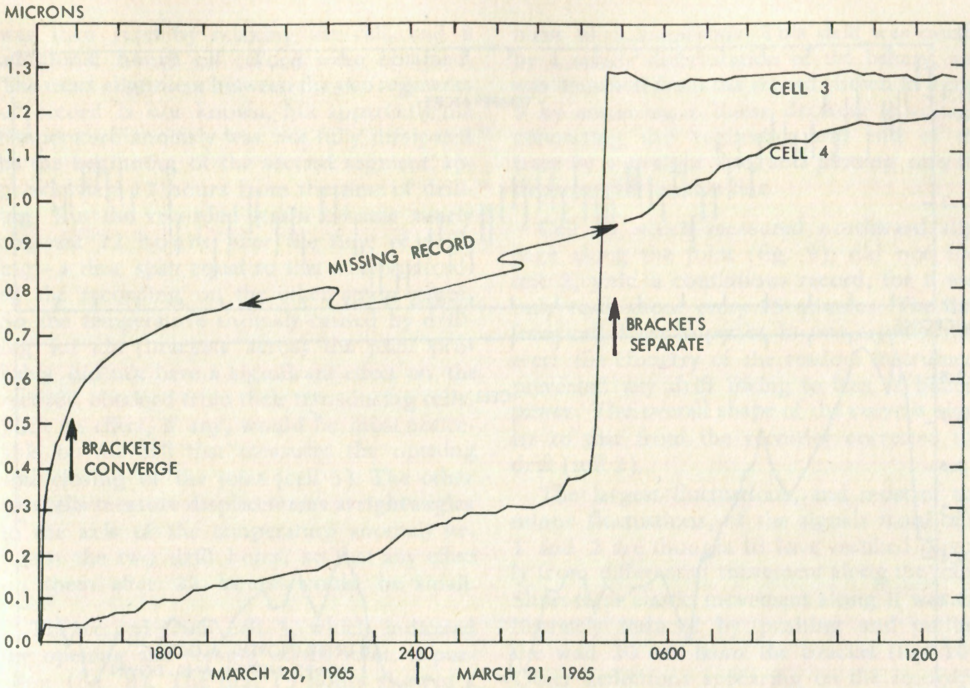

Figure 8.
Transducing cell 4 records the contraction of unjointed limestone caused by dissipation of the heat introduced by drilling holes for the brackets. The holes were drilled 3 hours before the start of the record. The relation between the segment following the missing record and that before it is approximate. Transducing cell 3 records the separation of the brackets spanning the joint.

recorder (fig. 7). Excitation voltage for the 350-ohm cell was supplied by a 6-volt dry-cell battery; power to run the recorder, by a portable generator outside the cave. Slight changes in the speed of the generator caused small changes in the chart speed, but the circuitry of the recorder is such that the voltage from the transducing cell is independent of this power source, and fluctuations of the generator had no effect on the value of the recorded strain. Laboratory calibration indicated that a change of 1.0 millivolt in the output from the cell was produced by a displacement of 1.1 micron at the tip of the cell. Chart sensitivity was 1.0 millivolt per 25 cm of chart width. Chart fluctuations of 0.5 mm can be detected, so a precision of about 0.002 micron was possible with the recorder.

The transducing cell on the brackets that spanned unfractured rock (cell 4) specifically tested the effect of the temperature anomaly created by the setting of the brackets themselves. Whereas 22 hours elapsed from the time when the brackets across the joint were placed until the recording began, the brackets holding cell 4 were placed only three hours before the recording. All the holes were drilled with a power drill holding a carbide-tipped rock bit, and those for cell 4 were 9.9 cm apart. Unloaded, the drill turned at 400 revolutions per minute, and the temperature in the holes probably reached about 100°C.

The first part of the curve from cell 4 is clearly exponential (fig. 8) and indicated that the limestone between the drill holes is cooling and shrinking, thereby compressing the gage. After 5 hours of record, the cell capacity was exceeded, and no readings were obtained for the next 9 hours. The instrument

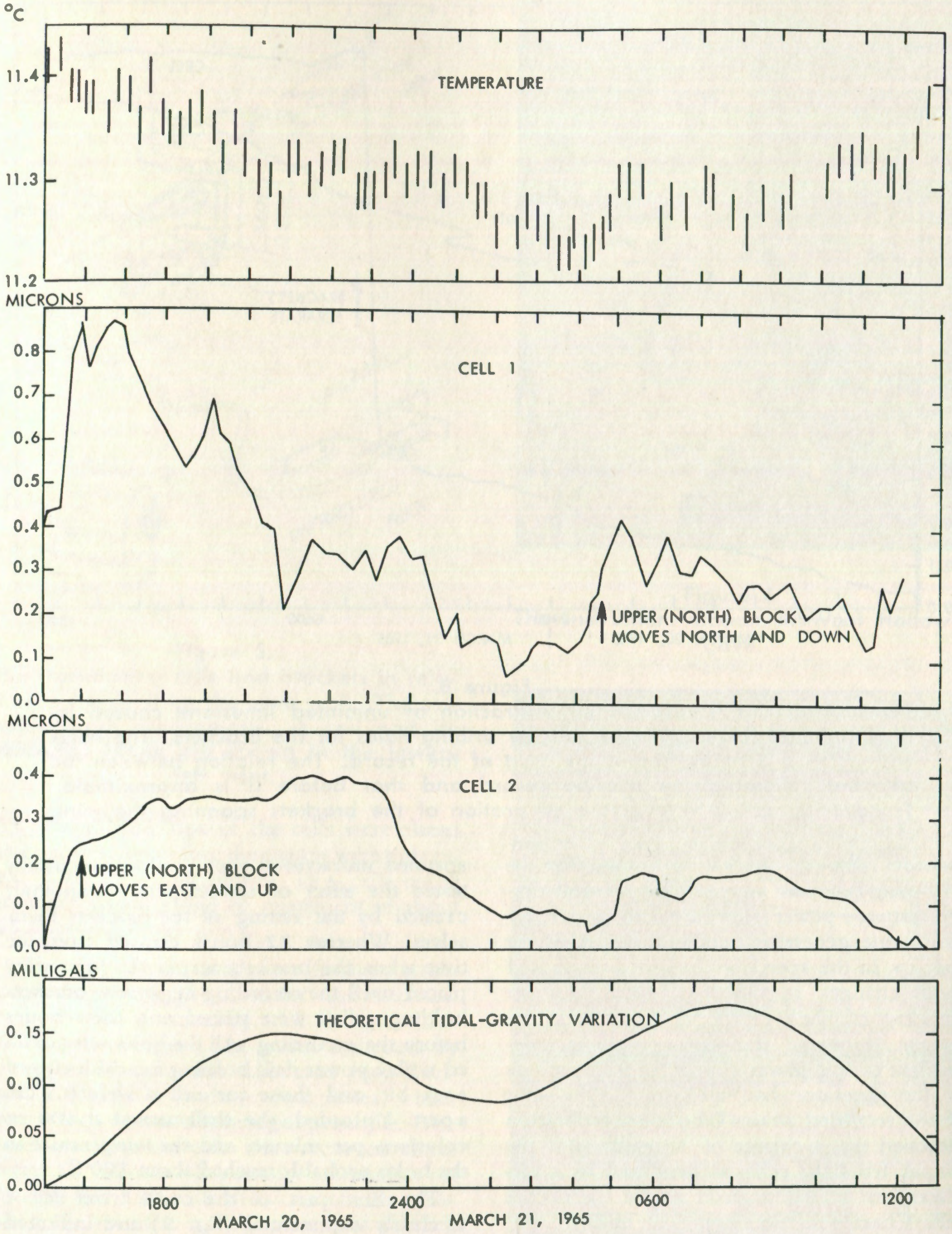

Figure 9.
Temperature at the main instrument station, northward and eastward slippage of the hanging wall of the joint, and the theoretical tidal-gravity variation at Wool Hollow Cave from 1500 March 20 until 1300 March 21, 1965.

was then reset by relaxing the cell, and 8 additional hours of record were obtained. The exact alignment between the two segments of record is not known, but apparently the temperature anomaly was not fully dissipated by the beginning of the second segment, approximately 17 hours from the time of drilling. But the recorded strain became nearly constant 22 hours after the time of drilling – a time span equal to that which preceded the recording on the other strain gages. So the temperature anomaly caused by drilling for the brackets across the joint probably did not have a significant effect on the records obtained from their transducing cells. Such an effect, if any, would be most noticeable on the cell that measuses the opening and closing of the joint (cell 3). The other two cells measure displacements at right angles to the axis of the temperature anomaly between the two drill holes, so that any effect on them after 22 hours would be small.

The record from cell 3, which indicated the opening and closing of the joint, is puzzling (fig. 8). The first 13 hours showed a slow extension of the strain gage, after which a large displacement took place. It was not abrupt but lasted over a period of about 15 minutes, and was followed by a relaxation for the next hour, after which the record remained quite steady. We can only speculate on the meaning of these results. One possibility is that a temperature anomaly was first created in the limestone by the drilling, that we then screwed the bracket tightly against the limestone, that the slow dissipation of the thermal anomaly over the succeeding hours caused a steady buildup of stress between the bracket and the limestone, and that, finally, the bracket slipped a little at its contact with the limestone.

The recorder was run continuously at a rate of about 1.0 cm/min for slightly more than 1300 minutes from 1500 March 20 to 1300 March 21, 1965. Thus 13 meters of rather detailed chart record is available for a study of the movement of transducing cell 2, which indicates eastward slippage along the joint. A condensation of this record is given in figure 9, and selected tracings from the chart are reproduced in figures 10 and 11. A long-term drift of the voltage took place, which resulted in a total apparent displacement of 2.2 microns. This drift was caused by a steady deterioration of the battery, and was removed from the record shown in figure 9 by assuming a linear decrease in voltage, connecting the beginning and end of the trace by a straight line, and plotting only the departures from that line.

Cell 1, which measured northward slippage along the joint (fig. 9), did not, like cell 2, yield a continuous record, for it was only read about every 15 minutes. The data from cell 1 are superior in one aspect, however: the circuitry of the readout instrument prevented any drift owing to loss of battery power. The overall shape of the curve is similar to that from the recorder corrected for drift (cell 2).

The largest fluctuations, and most of the minor fluctuations, of the signals from cells 1 and 2 are thought to have resulted directly from differential movement along the joint. Short-term elastic movement along it was deliberately induced by pushing and pulling the wall 20 cm from the bracket (fig. 10). Other deflections appearing on the recorder, however, were unrelated to our movements within the cave. Walking past the instruments had no effect so long as the thermal insulation was in place. The rather regular exponential curve obtained from cell 4 (fig. 8) indicates that almost all the movement was concentrated along the joint and that little occurred within the solid rock.

During the 22 hours of record, both cells 1 and 2 show two maxima and two minima that are roughly coincident in time with the theoretical earth tides plotted on figure 9 from Service Hydrographique (1964). At high tide the hanging wall of the joint moved southeastward up ths dip 0.4 micron with respect to the footwall, and just before high tide, 0.6 micron northeastward along the strike of the joint.

The geometry of the mountings for cells 1 and 2 indicates that the net deformation caused by expansion of the brackets owing to changes in the cave-air temperature should be small. The long-period diurnal temperature fluctuation, both within the cave (fig. 9) and at the surface, seems to bear little relation to the subsurface strain.

A lack of wind at Wool Hollow during the observations was evidence of stable at-

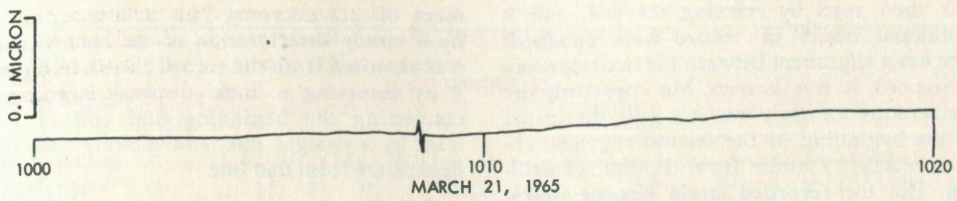

Figure 10.
A tracing of a segment of the recorder chart (cell 2), on which an upward deflection indicates an eastward movement of the upper or northern block. This trace is typical of most of the 13 meters of record, except that the small deflections at 1008 were caused by pushing and pulling the wall about 20 cm from the transducing cells.

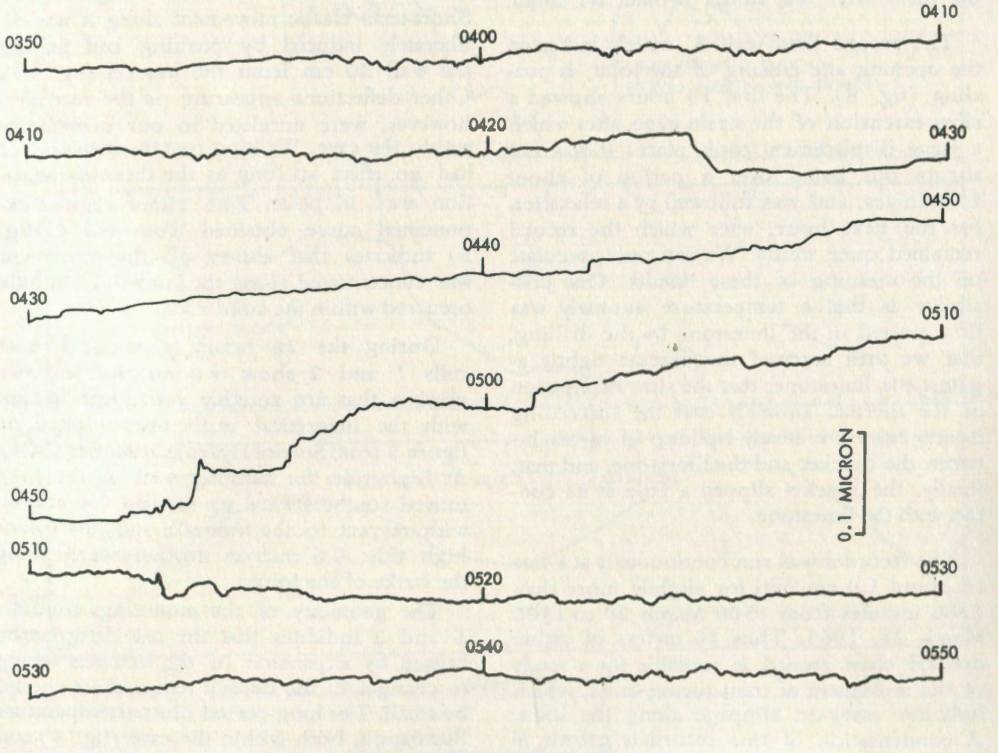

Figure 11.
A tracing of the recorder chart from 0350 to 0550, March 21, 1965, showing oscillations probably caused by a distant earthquake, and showing step-like displacements that may have been caused by the oscillations.

mospheric conditions. The nearest barometric station, at Stockton, California, recorded a total pressure fluctuation of 0.07 inch of mercury during the observation period, and only 0.01 inch between 2000 March 20 and 0600 March 21. This period of nearly constant barometric pressure coincided with a period of intensified movement along the joint. The area of the cave also had not received precipitation for eight days prior to the tests, so neither air nor subsurface-water loading can have caused important stress fluctuations within the rock. And because the cave is 200 kilometers from the Pacific Ocean, loading by oceanic tides probably had little effect.

Most of the continuous record showed rather smooth changes in transducing cell 2. Figure 10 is typical. Considerable activity occurred, however, between 0300 and 0800 March 21, with the greatest intensity between about 0458 and 0700 (fig. 11). This activity as also detected by cell 1, and was seen as needle fluctuations on the other readout instrument, which operated independtly of the recorder.

Two types of disturbance occurred along the joint during this interval–prolonged sinusoidal movements having amplitudes of 0.005-0.01 micron and periods of 10-40 seconds, and rather abrupt step-like displacements as large as 0.03 micron.

The sinusoidal movements probably represent the seismic signal from an earthquake in the Molucca Sea which occurred at 0308, Wool Hollow Cave time, and had a Richter magnitude of 6.2 (U.S. Coast and Geol. Survey, 1965). The Rayleigh surface wave from this earthquake was received at the Oroville seismographic station, 190 km north of the cave, at 0458 (B. A. Bolt, Univ. Calif. Berkeley, personal commun., 1965). The total vertical ground displacement recorded on the Oroville long-period seismograph was 26 microns at a period of 40 seconds.

The cause of the abrupt displacements is uncertain. Their correlation in our records with the interval having the largest seismic vibrations suggests a causal relation. We tentatively suggest that the seismic vibrations reduced the static friction along the joint, thereby permitting the abrupt displacements to take place.

Conclusion

The most important regular source of stress causing movement along the joint studied in Wool Hollow Cave is thought to be the tidal force. Since the overall trend of movement recorded from cells 1 and 2 corresponds well with the theoretical tidal changes as calculated from the earth's gravity field, the evidence here presented suggests that in mio-latitudes as much as about 1 micron (0.001 mm) of semidiurnal strain can be induced along joints by earth tides. But, because a seismic event on the other side of the earth produced more than 0.01 micron of movement, a nearby earthquake would probably cause displacements larger than those recorded from the earth tides. Deep inside a cave, such an earthquake produces a low-pitched rumble that resembles prolonged thunder (Lange, 1961).

These observations in Wool Hollow Cave may be the first that show movements along an individual joint, but several investigators have measured semidiurnal variations of the separation of two points in mines (Ozawa, 1957, Benioff, 1959). In those cases in which the measured points were many meters apart, most of the strain was probably distributed along several joints, and the sense of the measured movement depended on the angular relation between the joints and the measured line.

Many workers have previously theorized that semidiurnal movements related to earth tides probably occur along joints, and some, such as Hodgson (1961), have believed that the joints themselves may have been formed by such tidal flexing.

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Cutters and Pinnacles in Greene County, Missouri

By L. D. Fellows

ABSTRACT

The Burlington Limestone (Mississippian) crops out in southwestern, central and northeastern Missouri and has been extensively leached. Its less soluble constituents have accumulated, forming a layer of residuum of variable thickness. Precipitation infiltrated the residuum became weakly acidic, and moved laterally in the direction of slope of the bedrock surface. Patterns of water movement were etched into bedrock forming dendritic systems of cutters that "drain" into valleys, sinks, joints, and depressions in the bedrock surface that have no topographic expression. Cherty horizons were leached more slowly than non-cherty horizons. Chert beds and nodules became part of the residuum and were slowly draped over pinnacles. Engineering problems caused by cutters and pinnacles hinder urban development, highway construction and quarrying.

INTRODUCTION

Cutters, incisions in bedrock formed by leaching at the bedrock-regolith contact, and pinnacles, upward projections of bedrock between cutters, have developed in the Burlinton Limestone (Mississippian) throughout Greene County, Missouri. They are present in the subsurface, have no topographic expression, and can be observed only in excavations. Engineering problems caused by them are numerous and varied, particularly in urban areas. The purpose of this report is to explain the origin of cutters.

Nodules and thin beds of chert were formed when limestone was replaced by SiO_2. Although some nodules are spheroidal, most are irregular in shape, flattened parallel with bedding planes, and range up to two feet in diameter. Cherty horizons ranging up to six feet in thickness are commonly separated by 10 to 30 feet or more of non-cherty, bioclastic limestone (fig. 2).

The Burlington Formation consists of 95-98 percent $CaCO_3$ and small amounts of chert, clay- and colloidal-sized particles of SiO_2, montmorillonite, and possibly illite (Buckley and Buehler, 1904, Giles, 1935, Robbins and Keller, 1952, and Runnels and Schleicher, 1956). When subjected to chemical weathering processes, calcite dissolves whereas less soluble constituents accumulate as residuum.

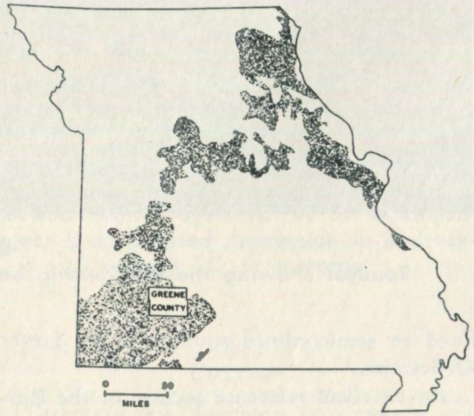

Figure 1.
Index map of Missouri showing outcrop areas of Mississippian strata (stippled).

Although the Burlington Limestone has relatively low intergranular porosity and permeability, water moves through it along major joints and bedding planes. Some shallow wells in the county produce water from bedding planes or joints in the Burlington Formation. However, most groundwater comes from con-

Figure 2.
Chert nodules in the Burlington Formation.

Figure 3.
Roadcut showing the relationship between cutters, pinnacles, and residuum.

fined or semiconfined aquifers of the Lower Ordovician.

An excellent reference section of the Burlington Limestone is exposed in roadcuts along U. S. Highway 65 in southern Greene County, Missouri (Sec. 21, T. 28 N., R. 21 W.). One drives southward from the Lake Springfield bridge through approximately 135 feet of section, the top part of which has been eroded.

Geology

Regional geology. Greene County, Missouri, is on the southwestern flank of the Ozark uplift, deeply eroded since the end of the Paleozoic Era, within outcrop belts of Mississippian strata that dip gently toward the west and southwest (fig. 1). The Burlington Limestone is the most widely exposed unit in the county. Younger Mississippian units crop out to the west. Erosion along the crest of the Ozark uplift exposed rock of Precambrian, Upper Cambrian, and Lower Ordovician age, the outcrop belts of which collectively form the Ozarks (Beveridge, 1951, 1962). Large areas within the Ozarks and adjacent to them are underlain by relatively soluble rock units in which a variety of solution features have developed.

Burlington Limestone. The Burlington Limestone and the overlying Keokuk Limestone, both of Mississippian (Osagean) age, are transitional, have similar lithologic characteristics, and can be differentiated only on the basis of fossils. For practical purposes they are usually lumped together as a single

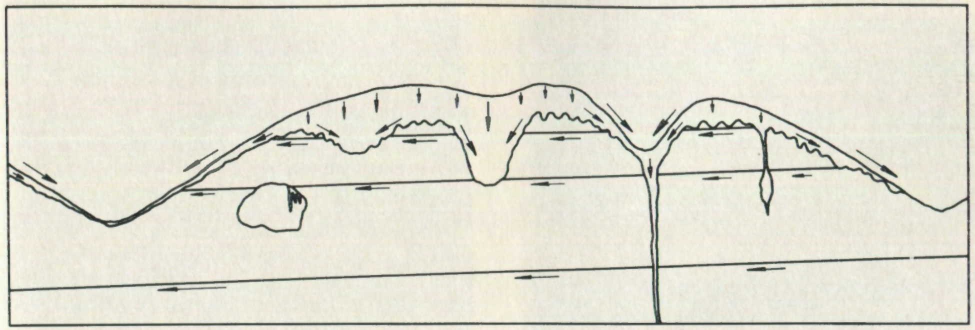

Figure 4.
Movement of water in the subsurface.

unit. The combined thickness of these formations varies from 0-200 feet in Greene County. Both units consist almost entirely of bioclastic limestone composed largely of crinoid columnals and calyx plates. Fossils of other marine invertebrates such as brachiopods, bryozoa, blastoids, and solitary corals are common.

Leaching

Leaching is the continuous removal of minerals from the soil profile, regolith, and/or bedrock by chemical processes such as hydrolysis, ion exchange, carbonation, solution and chelation. For an explanation of these processes refer to Keller (1957).

The extent to which bedrock has been leached depends on its lithologic characteristics as well as climate, pH of subsurface water, and length of time during which chemical weathering occurred. If bedrock is soluble, dense, has well-developed joints and bedding planes, and is overlain by porous and permeable regolith, it might be intensely leached (Thornbury, p. 317-318). Limestone, dolostone and gypsum are more readily leached than other rock types. Water that infiltrates is made weakly acidic by the addition of hydrogen ions provided by ionization of carbonic acid, decomposition of organic matter, roots of growing plants, acid clay, and/or oxidation of sulfide minerals. Temperature determines the rates of chemical reactions, some of which double or triple with an increase of 10°C.

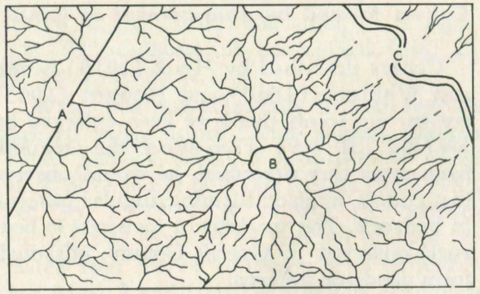

Figure 5.
Hypothetical map showing relationship between cutter systems. A = enlarged joint. B = isolated depression in bedrock surface. C = stream.

Cutters

Terminology. Leaching of outcrops results in the formation of solution pits, facets, and furrows, depending on the slope of the bedrock surface (Smith and Albritton, 1941). Lapiés are furrowed surfaces developed in outcrops having steep or moderately steep slopes (Cvijic, 1924, p. 26). Comparable surfaces form by leaching at the bedrock-regolith contact. Incisions in bedrock that formed beneath regolith were named *cutters* by Hook (1915, p. 65), who described their relationship to phosphate deposits in south-central Tennessee. Pinnacles are upward projections of bedrock between cutters. Cutters and pinnacles are comparable with solution furrows and ridges, respectively.

Figure 6
Cutters formed by enlargement of joints.

Cutters described by Hook (1915), Smith and Whitlatch (1940) and Howard (1963) are merely joints that have been enlarged by leaching. If one removed the regolith from areas they described, he would observe that cutters intersect forming angular patterns in bedrock. The majority of incisions in bedrock observed in Greene County, Missouri, form dendritic patterns.

If other types of furrows in bedrock were present in areas described by Hook, Smith and Whitlatch, and Howard, they were either not considered to be cutters or were overlooked. Howard (p. 46) mentioned secondary furrows leading at steep gradients into the main cutters (enlarged joints) but did not elaborate on them. On the basis of observations made in Greene County, Missouri, the writer feels justified in including incisions in bedrock that form dendritic patterns as cutters.

Origin of cutters. Most leaching of the Burlington and Keokuk Limestone has occurred at the bedrock-regolith contact. Less soluble constituents of the limestone, stained red with ferric oxide, have accumulated as residuum, which has an average thickness at Springfield of 15-20 feet. Residuum, which in other parts of southwestern Missouri attains a thickness of 60-70 feet, is commonly thickest beneath crests of hills (figs. 3 and 4).

Precipitation easily permeates residuum but not bedrock. Most farm ponds in southwestern Missouri do not hold water unless they

Figure 7.
Isolated depression in bedrock surface with dendritic cutter systems draining into it. This depression has no topographic expression.

are specially treated or constructed. Jamison and Thornton (1963, p. 7) concluded that the unusually high permeability of residuum is due to well-developed soil and subsoil structure and stable aggregation. Another possible explanation is that the clay- and colloidal-sized particles that comprise the residuum are nearly equidimensional rather than platy and allow water to move easily.

Water moves laterally through partially leached limestone at the bedrock-regolith contact and through the basal portion of the residuum, which is usually saturated. Open channels are not present. Lateral movement of water, determined by the slope of the bedrock surface, is toward joints, isolated depressions in the bedrock surface that have no topographic expression, and the floors of valleys and sinks, many of which are joint-

Figure 8.
Dendritic pattern of cutters.

Figure 9.
Cutters formed beneath hillside.

Figure 10.
Close-up of dendritic cutter pattern formed beneath hillside.

controlled (fig. 4). Subsurface water does not move uniformly through the residuum. Patterns of movement, many of which are dendritic, are etched into bedrock and are known as cutters. Enlarged joints and depressions in the bedrock surface are major cutters into which dendritic patterns of cutters drain (fig. 5).

Movement of water along a joint results in enlargement of the joint (fig. 6). Lateral movement of water toward a joint results in incision of dendritic cutters into bedrock along both sides of the joint. If residuum is compacted above the joint or is transported by water along the joint faster than it can accumulate, a topographic depression or sink results.

Isolated depressions in the bedrock surface, possibly of structural origin, are enlarged and deepened by leaching (fig. 7). Water that moves laterally toward these depressions forms dendritic patterns of cutters leading into them (fig. 8). In their initial stage of development, these depressions have no topographic expression, have no outlets, and are enlarged slowly. If an outlet later develops, the rate of water movement, leaching and enlargement increases. A sink might eventually form as described in the preceding paragraph.

Cutters that form dendritic patterns can be seen in roadcuts along Interstate Highway 44 north (Sec. 6, T. 29 N., R. 21 W. and Secs. 1, 2, and 3, T. 29 N., R. 22 W.) and west (Sec. 13 and 24, T. 29 N., R. 23 W.) of Springfield, along U. S. Highway 60 southeast of Springfield (Sec. 15, T. 28 N., R. 21 W.) and along U. S. highway 65 southeast of Springfield (Sec. 21, T. 28 N., R. 21 W.).

Water that moves laterally through regolith toward valley floors forms slightly different cutter patterns, excellent exposures of which were observed at only one locality (SE¼ NE¼ sec. 33, T. 27 N., R. 21 W.). The bedrock surface was exposed by removal of approximately four feet of residuum prior to quarrying (figs. 9 and 10). These cutters are closely spaced and are not incised deeply into bedrock.

Role of cherty horizons. As described previously, the Burlington Limestone in Greene County has cherty horizons a few feet thick separated by non-cherty limestone units (fig. 2). Because chert is much less soluble than limestone, cutter development nearly stops when a cherty horizon is reached. In other words, cherty horizons serve as temporary base levels of leaching, as do other relatively less soluble units (Dicken, 1935 p. 723, Hamilton, 1948, p. 44, Howard, 1963, p. 48). By the time a cherty horizon has been destroyed by leaching, most of the overlying limestone has dissolved. Although some isolated pinnacles might be left standing above the cherty horizon, the bedrock surface is essentially devoid of pinnacles (fig. 11B). As leaching and cutter development continue, chert beds or nodules become part of the residuum (fig. 12). The tops of newly formed pinnacles are all at approximately the same level, the level of a former cherty horizon (figs. 3 and 13). Eventually chert nodules or beds get draped over the pin-

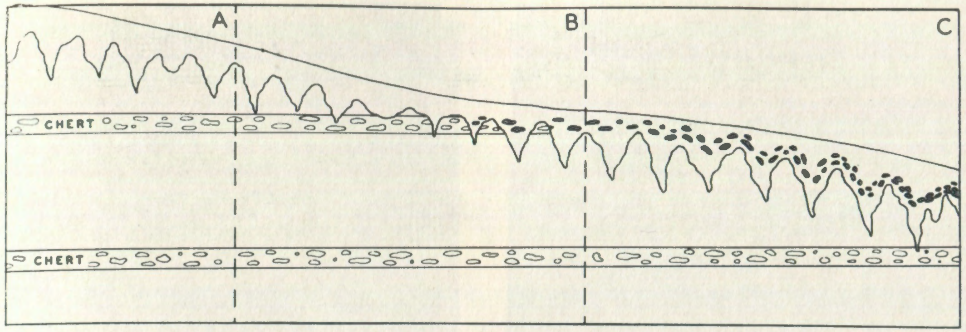

Figure 11.
Stages in cutter development. Chert in residuum is shown in black.

nacles (fig. 11C and 14). Hook (1915, p. 66) described the development of cutters in essentially the same way.

The bedrock surface is continuously being lowered by leaching and the land surface by erosion. The bedrock surface has been lowered fastest and residuum has accumulated. Chert nodules or beds slowly decompose, get broken apart, and are concentrated near the surface, essentially as lag gravel.

In Pulaski County, Missouri, NW¼ SE¼ sec. 18, T. 36 N., R. 10 W.) near the town of Devils Elbow is an unusual example of residual chert beds. (Beveridge and Hayes, 1960, p. 30). Draped over pinnacles of the Gasconade Dolostone is 22 feet of residuum, mostly thin beds of chert, that represents approximately 260 feet of original bedrock section. K.H. Anderson, a geologist on the Missouri Geological Survey staff, concluded that part of the Gasconade, all of the Roubidoux, and the lower part of the Jefferson City formations are represented in the residuum.

Variability of solution features. Because many variables are involved in leaching, solution features differ considerably from one locality to another. Variables include amount of precipitation, ratio of precipitation to evaporation, solubility of bedrock, porosity and permeability of bedrock, degree of development of joints, presence of relatively permeable or less soluble strata in the section, water holding capacity of regolith, and length of time leaching has occurred. This discussion of the origin of cutters and pinnacles is restricted to Greene County, Missouri, although the Burlington Limestone crops out throughout the state. Even within Greene County, one finds a variety of solution features because of differences in the stage of development and/or topography (fig. 11).

Similar solution features and related problems exist in other portions of this state and other states that have relatively soluble rock units at or near the surface. In the areas described by Hook (1915), Dicken (1935), Smith and Whitlatch (1940), and Howard (1963) joints are apparently much better developed and linear cutters (enlarged joints) predominate. Howard was the only writer to mention the dendritic variety. Although enlarged joints are common in parts of Greene County, the majority of cutters observed are of the dendritic variety formed by water moving laterally through residuum.

Figure 12.
Chert nodules surrounded by residual red clay.

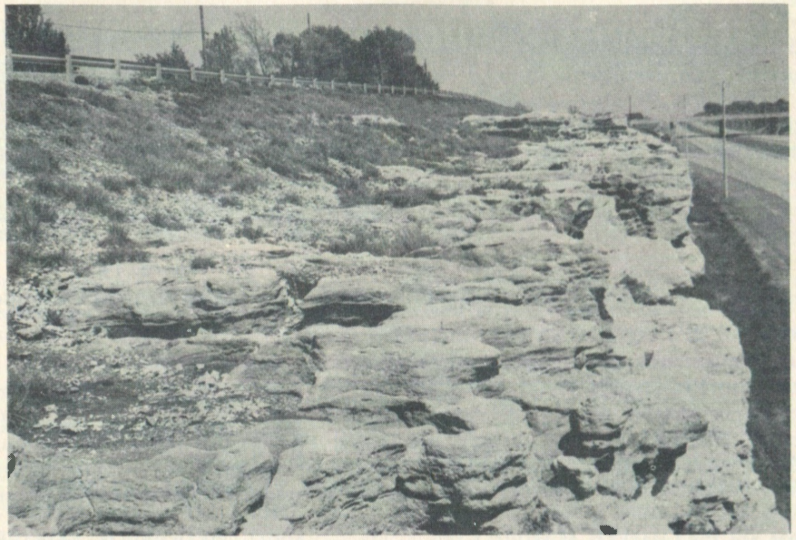

Figure 13.
Pinnacles, the tops of which are at the level of a cherty horizon that has been almost completely leached.

Usage of term lapiés. Smith and Albritton (1941, p. 73) and Thornbury (1954, p. 319) described the general lack of agreement concerning the definition of lapiés. Cvijić (1924, p. 44) stated that although lapies are most extensively formed on bedrock surfaces with no regolith or vegetation, they can also form beneath a regolith cover, However, it is the writer's opinion that the term lapiés should be restricted to those features formed in soluble rock with no regolith cover. Cutters and pinnacles form at the bedrock-regolith contact.

Cutters form because of a slow lateral movement of water, charged with hydrogen ions, through regolith on a gradient related to the slope of the bedrock surface. Lapiés result from a rapid flow of water, with few hydrogen ions added, across surfaces of bare bedrock on gradients related to topography. Cutters are best developed on gently inclined surfaces, whereas lapiés form on moderate to steep slopes. Cutters and pinnacles are exposed at the surface either by accelerated erosion or by artificial removal of regolith. Lapiés form directly at the surface.

Cutters and engineering problems. Because large areas of bedrock in which cutters have developed are rarely exposed, it is difficult to detect any order or pattern. However, examination of cutters exposed in roadcuts, quarries and other excavations revealed that systems of cutters developed in an orderly manner and that cutters developed beneath hillsides are different from those beneath crests of hills. An understanding of this relationship enables one to visualize the configuration of the bedrock surface and to anticipate some of the engineering problems that commonly hinder urban development, highway construction, and quarrying.

PINNACLES

Pinnacles are merely the relatively unleached portions of bedrock that have been left between cutters. They vary considerably in size, shape, and stage of development from one locality to another (fig. 11). All pinnacles are irregularly shaped with indentations that have developed along finely fragmental or finely crystalline units (figs. 2, 2, 6, 7, 8, 10, and 13). Some have been completely detached from the underlying bedrock and are surrounded with residuum. Fossil fragments are etched out in relief on the sides of pinnacles and small scale cross stratification is often visible.

Figure 14.
Thin beds of chert draped over the top of a pinnacle.

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A Hypothesis for the Formation Of Rimstone Dams and Gours

By William W. Varnedoe, Jr.

ABSTRACT

Rimstone dams, common in hot spring deposits and often found in cave streams, are thin, dam-like structures oriented at right angles to stream flow and growing vertically, with their top edges horizontal. Supersaturation in the depositing waters is a prerequisite for rimstone formation. Once supersaturation has been attained, rimstone growth may be triggered by 1) transition from laminar to turbulent flow, 2) transition from subcritical to supercritical flow at Froude Number = 1, or 3) the writer's hypothesis, that an increase in velocity over an obstruction causes a decrease in internal pressure and resultant degassing of the solution, resulting in the buildup of vertical dams.

Although rimstone pools and associated phenomona are common in many United States caves and are well known to American cavers, there is relatively little mention of them in the literature. There is an extensive European literature on the subject which has been reviewed by Warwick (1952). Warwick lists a large number of theories for the origin and development of rimstone, most of them highly speculative, and all of them based on intuitive arguments. It is the purpose of this note to inject a new hypothesis into the discussion. It is proposed that the unique distribution and geometry of rimstone dams can be explained by changes in the flow of the stream from which the dams are being deposited.

Rimstone dams come in a continous spectrum of sizes from gours one-fourth inch deep and one half inch wide such as are often found on sloping flowstone surfaces to huge dams several feet high and spanning wide passages. They are seldom thicker than a few inches and all are oriented at right angles to the local stream flow. They all grow vertically with their top edges horizontal. The travertine which makes up the dam tends to be rough and porous, unlike the dense material of normal flowstone.

Rimstone dams occur in many caves and also occur on the surface as hot spring deposits and as deposits of normal limestone springs in tropical climates. In cave occurrences, rimstone dams are sometimes associated with flowstone deposits. One can find a continuous change from rather widely spaced rimstone dams in the downstream portion of a stream passage to a flowstone choke at the upstream end.

It has been fairly well shown (Moore, 1962, Holland et al., 1964) that travertine deposition takes place in caves by loss of carbon dioxide rather than by evaporation of solution. There exists, then, the possibility that the rate of travertine deposition on the bottom of a flowing stream is limited by the rate of diffusion of carbon dioxide away from the growing travertine-water interface. Any process which enhances the carbon dioxide loss will activate travertine deposition if the water is initially saturated or supersaturated.

The close association of rimstone dams and flowstone suggests that a critical supersaturation is a necessary prerequisite for rimstone deposition. If the water is only saturated or undersaturated no travertine can form. If it is highly supersaturated, then almost any disturbance will trigger travertine deposition and a normal flowstone, such as is observed in the headwaters of some rimstone depositing streams, will be formed. However, if the supersaturation takes on an intermediate value, deposition may not take place until triggered by changes in the hydraulic characteristics of the stream.

There are several hydraulic mechanisms which may serve to trigger deposition and in the absence of further experimental evidence, one cannot distinguish between them. They are listed below:

1. In the case of pre-existing pools, the velocity of flow may be so slight that the flow regime is laminar. Transportation of carbon dioxide to the pool surface in this case can take place only by diffusion (aided perhaps by slow convection). At the point where the pool spills over its lip, a transition to turbulent flow conceivably could take place. The turbulent mixing would then be much more efficient at bringing carbon dioxide to the surface, if it occurs, and thus this transition would provide a triggering mechanism for the deposition of travertine.

2. The transition from subcritical to supercritical flow at Froude Number = 1 takes place in a number of cave streams. In subcritical flow the water is tranquil and has a smooth surface. In supercritical flow considerable frothing takes place which could aid in conducting the carbon dioxide to the surface of the stream and discharging it into the atmosphere. This transition can be observed in many cave streams and may be responsible for the initiation of the rimstone dam at a particular point.

3. The new hypothesis results from rimstone dams behaving hydraulically like rectangular weirs. The large increase in velocity as the water sweeps over the crest of the weir results in a decrease in internal pressure with possible resultant degassing of the solution. This mechanism might be responsible for buildup of the vertical dam since deposition would be at the top where the velocity abruptly increases.

The velocity being greater and the pressure being less at a low point, such an irregularity would, therefore, disappear through preferential deposition. This would provide a stabilizing process to maintain the elevation contour of the dams.

Also, the more supersaturated the solution, the less pressure drop would be needed to trigger the deposition. In most cases, the edge along the very top receives the most deposition and the dam grows straight up. Occasionally, the water might be supersaturated enough to unload for the small drop in pressure just before reaching the peak of the dam. If this happens, the dam would bend upstream as it grows. Dams are sometimes observed to have reached the point that their upper edges have become almost horizontal.

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