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CONTENTS

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Stratigraphy of and Characteristics of Cavern Development in the Carbonate Rocks of New Jersey

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ABSTRACT

Caves of various sizes and shapes occur in the "Kittatinny limestone" of northern New Jersey; however, until the recent subdivision of this thick carbonate section, the formation or formational members in which the caves occurred was not known. Compilation of the cavern data and recognition of the formations or formational members in which the caves occur has resulted in interesting conclusions and provides clues for finding additional caves.

Recent mapping by the authors in the dolomitic rocks of northern New Jersey, detailed core logging in conjunction with an examination of the open rock core trench at the Spruce Run Reservoir site, examination of the Round Valley Reservoir pipeline trench, and preparation of the *New Jersey Cave Bulletin* form the basis of the results obtained in this study.

The relationship between cavern development and the stratigraphy and lithology of the "Kittatinny limestone" is compared with the geologic relationships of caves found in other carbonate formations in New Jersey. These results show that the relatively coarse-grained dolomitic units of the "Kittatinny" in New Jersey are more amenable to solution than are finer-grained rocks; consequently, there is a greater chance of finding caves in those formations containing coarse-grained units. The results of studies of caves in dolomitic rocks are compared to those of caves found in the Devonian-Silurian-Ordovician limestones and in the Precambrian Franklin marble. Studies on limestone solution by Rauch (1970) indicate that caverns develop more readily in fine-grained limestones than in coarse-grained rocks, which is the opposite of what we find in the dolomitic rocks of New Jersey.

INTRODUCTION

In northern New Jersey, there are several large areas of carbonate rocks totaling more than 225 square miles in area (Fig. 1). These rocks range from Precambrian to Devonian in age. The Cambro-Ordovician units consist principally of dolomite and the Silurian-Devonian formations principally of limestone. The Cambro-Ordovician "Kittatinny limestone", actually a dolomite, is the most important from a speleological viewpoint as it contains 70% of the more than 100 known caves in the area. Until recent detailed mapping was carried out, all caves in the dolomitic rocks were assigned to the "Kittatinny" since it was not known where in the 3,000 to 4,000 ft. of the section the caves were developed.

The authors currently are mapping the "Kittatinny formation" in Sussex and parts of Warren counties and are subdividing it into formations and members. As a result of this study, most of the caves now can be assigned to formations and, in some cases, to specific members. The results of this cave comparison study are based upon carbon-

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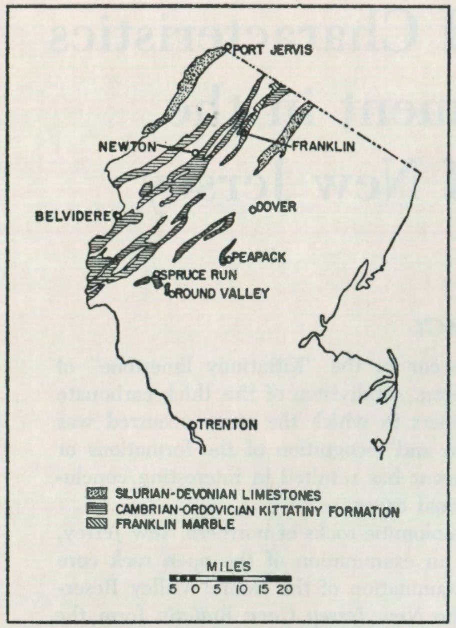

Fig. 1—Map of Northern New Jersey showing Outcrop Areas of Carbonate Rocks.

ate-rock mapping, data gathered during core logging, geologic examination of the core trench excavation at the Spruce Run Reservoir site, and mapping of the Round Valley pipeline excavation.

Stratigraphy

In New Jersey, most of the caves are found in rocks ranging in age from Precambrian to Devonian (Table 1). The Precambrian rocks include granites, gneisses, and marbles. The marbles are collectively known as the Franklin marble. The Franklin marble is composed of several different rock types and, at its type locality in the Franklin-Sterling Hill area, it is divided into three separate bands (Hague, *et al.,* 1956), which are interbedded with gneiss. The Wildcat band is approximately 300 ft thick. The Franklin 1 and 2 bands each are 1500 ft thick. The Franklin is a coarse [1] crystalline marble consisting principally of calcite and dolomite with some very siliceous units. The Franklin-Sterling Hill marble area has been mined for zinc since the early 1800's. Several caves have been found during mining. The formation contains 26 caves. Six caves are more than 100 ft long. The largest is 780 ft long.

The Cambro-Ordovician formations contain 71 caves (Dalton, 1969). The number of caves in each formation is as follows: Leithsville 10; Allentown 19; Rickenbach 8; Epler 21; and undifferentiated "Kittatinny" 13. Sixteen of these caves exceed one hundred feet in length.

The Hardyston formation is the basal Cambrian clastic unit in northern New Jersey. It varies from a feldspathic sandstone to a silty shale to a quartzite and is not known to be cavernous. The Hardyston grades upward into the Leithsville formation, which is the basal formation of the "Kittatinny" sequence.

The name "Kittatinny formation" was given by the early workers to all of the Cambro-Ordovician carbonate rocks in New Jersey older than Black River. It is a 3,000 to 4,000 ft thick sequence of dolomites with minor limestone and shale beds. The name "Kittatinny" is being phased out and the unit is being subdivided into four definite formations: The Leithsville and Allentown of Cambrian age and the Rickenbach and Epler of Ordovician age.

These new formational names were brought into New Jersey from Pennsylvania by Drake (1965) and subsequently were adapted by Markewicz (1967). Later work by Markewicz and by Markewicz and Dalton on the Cambro-Ordovician dolomites indicates that a yet more-detailed subdivision is possible. Field work suggests that the upper part of the Epler may be subdivided in the future to include a fifth formation equivalent to the Ontelaunee formation of Pennsylvania. Detailed examination of the Ontelaunee formation in Pennsylvania by Markewicz suggests that the top of the Epler formation may be correlative with the Ontelaunee. Each formation includes several recognizable lithofacies, making it possible to divide the formations into members. Because of their economic and environ-

[1] See Table 3 for grain-size comparison.

TABLE 1. Late Precambrian and Lower Paleozoic Carbonate Formations of New Jersey

Age	Group	Name	Thickness (ft)	Description
Devonian		Middle Devonian Shales and Sandstones	4300	Sandstones, shales, and conglomerates
		Onondaga Limestone	250	Black, cherty, fine to medium grained limestone
		Esopus Grit	180	Black gritty siltstone
		Oriskany Formation	170	Siliceous limestone to calcareous sandstone
		Port Ewen Shale	150	Calcareous shale
		Becraft Limestone	20	Gray, cherty, fossiliferous limestone
		New Scotland Formation	180	160 ft of calcareous shale 20 ft of cherty limestone
		Stormville Sandstone	10	Calcareous sandstone
		Coeymans Formation	40	Coarse, crystalline, crinoidal limestone with some chert
Silurian		Manlius Limestone	15–45	Blue to black, thin to irregularly bedded limestone
		Rondout Formation	40	Interbedded dolomite and limestone with some shale
		Decker Formation	80	Calcareous shale and fossiliferous limestone
		Bossardville Formation	100	Gray to black, massive to laminated, argillaceous limestone, dolomite and shale
		Poxono Island Formation	200	Calcareous to dolomitic shale to dolomite with limestone beds
		High Falls Formation	1500	Red and green shale and sandstone
		Shawangunk Formation	1500	Quartzites, conglomerates, and shale
Ordovician		Martinsburg Formation	8000	Black shale, slates, sandstones; brown to red shale with sandstones and limestone
		Jacksonburg Foundation	250	200 ft black agrillaceous limestone to calcareous shale 50 ft black limestone with conglomerate zones
	Beekmantown / "Kittatinny"	(Ontelaunee Formation?)	350	Dark gray to black, very cherty, fetid dolomite and fine to medium grained dolomite
	Beekmantown / "Kittatinny"	Epler Formation	700	Gray, massive, laminated dolomite with shale and local limestone lenses
	Beekmantown / "Kittatinny"	Rickenbach Formation	500	Dark gray to black medium bedded crystalline dolomite
Cambrian	"Kittatinny"	Allentown Formation	1400	Interbedded, light and dark gray dolomites with cryptozoa, oolites, and inter-formational chip conglomerates
	"Kittatinny"	Leithsville Formation	450–600	Dark gray, irregular bedded to massive dolomite with thick argillitic shale near base
		Hardyston Quartzite	0–200	Calcareous sandstone, quartzitic sandstone, quartzite, and conglomerate
Precambrian		Franklin Marble	1500	Light gray to white, very coarse grained graphitic calcareous dolomitic to siliceous marble

mental importance and their potential as sources of ground water, these formations and members are described in more detail below.

The largest areas underlain by carbonate rocks in New Jersey are those previously mapped as the "Kittatinny limestone". The lowermost "Kittatinny" unit, the Leithsville formation of Lower-Middle (?) Cambrian age, consists of three generally recognizable members. These are to date unnamed, pending further work. The lower Leithsville is a medium- to coarse-grained, thick- to massively-bedded [2] though rubbly, undulating, silty to locally sandy dolomite containing scattered white dolomite clots and crystals and frequent discontinuous masses or lenses of pyrite. Sulphide is most abundant in the lowest part of the formation. This unit, though it lies on the Hardyston quartzite or at some places directly on the Precambrian, is generally covered and may contain many small sinkholes. The large caves at Peapack and Leigh Cave near Allerton are developed in this member. Above the lower member, there is a shaly to siliceous dolomite unit approximately 100 to 150 ft thick which we have named the Hamburg member. This member, the most resistant lithologic unit in the Leithsville, occasionally underlies a topographic high. There are no known sinks or caves in the Hamburg member. The upper member, the thickest unit of the Leithsville, is a fine- to medium-grained, medium- to thick-bedded dolomite that grades upward into the Allentown formation. Except for the Hamburg member, the Leithsville generally underlies a subdued to marshy lowland topography containing many large sinks. The authors believe that there are extensive leached areas in the Leithsville, especially in the lower and upper parts, but no caves have been found because of thick overburden and the wet terrain that is typical of the formation.

The Allentown formation (Upper Cambrian) is presently divided only into a lower member (in Pennsylvania, named Limeport by Howell, *et al.* [1950]), and an upper member; however, further subdivision will follow as work progresses. The lower (Limeport) member includes three units: a lower, fine-grained, cyclicly-bedded, oolitic and cryptozoan-bearing dolomite; a medium-dark-gray to black, medium- to coarse-grained, cryptozoan and oolitic dolomite containing silt and sand; and an upper fine-grained unit. The lower unit flanks the first prominent topographic high above the Leithsville lowland; the middle unit generally forms a smooth, covered bench; the upper member forms a series of irregular small ridges and troughs with sinkholes occupying many of the small, shallow depressions. The largest cave system in the state is located in the lower (Limeport) member. The upper Allentown is a thick, uniform sequence of massively-bedded, fine- to medium-grained, dense dolomite with some sand and occasional cryptozoa. This sequence typically underlies a higher terrain which varies from a smoothly rolling landscape to one containing irregular shallow troughs and rugged pinnacles. Locally, there are small sinkholes and small caves in the troughs. The Allentown grades upward through a more variable, frequently sandier and coarser crystalline dolomite into the Rickenbach formation.

The Rickenbach formation (Lower Ordovician) now is divided only into lower and upper members; however, because of the variability of the formation, further subdivisions may be proposed. The lower Rickenbach resembles in part the lower Allentown and underlies a subdued topography similar to that developed on the upper part of the lower (Limeport) member of the Allentown. The lithology of the upper Rickenbach varies along the strike. Generally, it is a medium- to coarse-grained dolomite containing a few fine-grained beds. These rocks typically underlie a swampy, subdued topography. Sinkholes of all sizes are quite common. Most of the caves in the Rickenbach and many of the largest sinkholes occur in the upper half of the formation. At some localities, the upper part of the Rickenbach contains a paleo-solution breccia which may underlie a topographic high. The upper Rickenbach grades almost imperceptibly into the Lower Laminated member of the Epler formation.

[2] Thin beds are up to 1 ft thick, medium beds are up to 2 ft thick, thick beds are up to 6 ft thick, and massive beds are more than 6 ft thick.

The Epler formation (Lower Ordovician) includes five members, in ascending order: the Lower Laminated, Big Red, Upper Laminated, Black Jack, and Harmonyvale. If, as previously stated, field work should indicate that the upper portion of the Epler formation is equivalent to the Ontelaunee formation of Pennsylvania, the Black Jack and Harmonyvale would then become members of the Ontelaunee formation. The Epler generally underlies a high terrain surmounted by narrow to wide linear ridges of bedrock.

The Lower Laminated member consists of massive, dense, fine-grained dolomite containing thin, closely-spaced, siliceous laminae. It can be as much as 400 ft thick. The Lower Laminated member grades upward into the Big Red member. The Big Red is a reliable key horizon in mapping the Epler, although its lithology is laterally variable. Typically 50 to 150 ft thick, it includes three dominant lithologies:

1. Finely- to crypto-crystalline, siliceous dolomite occuring as an inter-formational breccia varying from a few ft to more than 100 ft in thickness. Typically, it occurs in the upper part of the member. Some shaly beds may be present, also.
2. Very fine-grained, siliceous dolomite containing thin, raised, resistant laminae or shards of silt which give the rock a ribbon-like to irregular, patchy appearance. Typically, the rock has a reddish-colored rind when broken.
3. Lenticular, locally massive, blue-gray, amorphous to very fine-grained limestone in beds from 1 in. to 2 ft thick. Thin beds are separated by siliceous ribbons. Locally, and south of Newton especially, the limestone may be somewhat crystalline, patchy, mottled and fossiliferous.

Above the Big Red lies the Upper Laminated member. It is from 150 to 250 ft thick and consists of thick-bedded to massive, fine- to medium-grained, laminated dolomite similar in character to the Lower Laminated member. This unit may contain large masses of breccia formed during Lower Ordovician erosion and karstification. These breccias seem to be the fillings of paleo-caves. They have been observed to cut hundreds of feet of section, from the top of the Epler down into the Rickenbach. The lower part of the member is transitional with the Big Red. The upper part of the member passes by intercalation of beds into the overlying Black Jack member.

The Black Jack member, 200 to 250 ft thick, is a massive, medium- to coarsely-crystalline dolomite containing a large amount of thick, irregularly bedded, rugose chert. This rock typically has a strong fetid odor when struck with a hammer. Silicified fossils have been found by the authors in rugose, cherty dolomite near the top of the member.

Throughout most of northern New Jersey, the eroded surface of the Black Jack is overlain by the Jacksonburg formation; however, there are isolated areas where the Black Jack is not eroded. At these places, it grades upward by intercalation of beds from massive, cherty, sparkly dolomite into fine-grained, thin- to medium-bedded, dense dolomite containing scattered cherty zones and small lenses of thin-bedded limestone. This member, named the "Harmonyvale" from a type exposure near that place, is found at several widely separated locations and is upwards of 150 ft thick.

No hiatus is known to be present between the Black Jack and the Harmonyvale. The Harmonyvale member, therefore, was continuous in pre-Jacksonburg time and was either completely removed or deeply eroded during the hiatus between Beekmantown and Jacksonburg deposition. As mentioned above, additional field work may indicate that the Black Jack, or the Harmonyvale, or both, are the stratigraphic equivalent of the Ontelaunee formation of Pennsylvania. Due to the large amount of chert which it contains, the most rugged rock exposures in the Epler sequence are formed by the Black Jack member. Very few sinkholes are found in the Epler formation. Most of the caves known to occur in the Epler are very small. The Upper Laminated member contains the largest number of caves and sinking streams.

The Middle Ordovician Jacksonburg formation is separated from the Lower Ordovician carbonates by an unconformity which is best exposed at the abandoned Sarepta Quarry near Belvidere. The Jacksonburg can be divided into two distinct units, a lower sequence and an upper sequence, which are similar to the Myerstown (lower) and Hershey (upper) formations of Pennsylvania.

The lower unit, commonly known as the "cement limestone", is a medium- to dark-gray, fine- to coarsely-crystalline limestone that locally is a high-calcium limestone. There are some beds of light- to medium-gray calcarenite. The total thickness of the unit is about 200 to 300 ft. One cave has been found in the limestone and another, at Sarepta Quarry, at the contact between the limestone and the underlying Harmonyvale member of the Epler formation.

The upper unit, known as "cement rock", consists of a dark-gray to black argillaceous limestone containing a pronounced cleavage. Several beds of coarsely crystalline limestone occur within this member. Its total thickness may reach 600 ft. No caves have been found in the cement rock.

A thick sequence of dark-gray to black shales and siltstones, known as the Martinsburg formation, occurs above the Jacksonburg in the northern part of the state. A vari-colored rock sequence overlying the Jacksonburg in the Clinton area is considered by most workers to be pre-Martinsburg in age. However, Markewicz (1967) has suggested on the basis of his recent mapping of the High Bridge Quadrangle that these vari-colored shales, siltstones, sandstones, and limestones represent a near-shore facies of the Martinsburg Formation. The Jutland member of the Martinsburg (new name for the lower part of the Martinsburg in the High Bridge area) contains several limestone beds totalling more than 100 ft in thickness. These limestones vary from fine-grained, ribbony to irregular, platy beds with local quartzose calcarenites. There are no known caves in these limestones in New Jersey, although several are known to exist in the "Jutland" in eastern Pennsylvania.

The Lower and Middle Silurian are represented by a thick section of quartzites, sandstones, and shales known as the Shawangunk conglomerate and the High Falls formation. The Poxono Island formation, which overlies the High Falls, varies from a calcareous shale to a dolomite with some laminated limestone beds. The Poxono Island is several hundred feet thick and grades into the overlying Bossardville limestone.

The Bossardville, the basal unit of the Upper Silurian, varies from a gray-to-black, massive, laminated, argillaceous limestone, to a dolomitic limestone. The upper part of the formation grades into a limy shale. The Bossardville has a maximum thickness of about 100 ft. It is overlain by the Decker formation, which varies from a limestone near Port Jervis, New York, to a calcareous sandstone at Walpack Center. The Decker also occurs locally in the Green Pond Mountain area north of Dover, New Jersey. The Decker has a maximum thickness of 80 ft. In the northwestern part of Sussex County, the Decker is overlain by the Rondout formation, a unit which varies vertically from sandstone to limestone to dolomite. Its total thickness is only about 40 ft. The Rondout is succeeded by the Manlius limestone (Epstein *et al.*, 1967), a blue-to-black, thin-bedded, flaggy limestone (locally massive). There are no known caves in the Poxono Island, Bossardville, Decker, Rondout, and Manlius formations.

The Lower Devonian is represented by the Helderberg group. In northwestern New Jersey, the Helderberg includes five, typically thin, formations. The Coeymans limestone, with a maximum thickness of 40 ft, is a coarsely crystalline limestone containing crinoid stems, brachiopods and some chert. The Stormville sandstone, which overlies the Coeymans, is a calcareous sandstone approximately 10 ft thick. At the Nearpass Quarry in northwestern New Jersey, the New Scotland formation rests directly on the Coeymans, but to the southwest they are separated by the Stormville sandstone. The New Scotland consists of about 20 ft of cherty limestone overlain by 160 ft of calcareous shale.

The Becraft (Minisink) limestone, above the New Scotland, is a very fossiliferous, gray, cherty limestone about 20 ft thick. The Becraft is overlain by Port Ewen shale. The Port Ewen is very poorly exposed but, when seen, is a calcareous to siliceous shale about 80 to 150 ft thick.

The only cave found to date in the Helderberg group of New Jersey is located in the Coeymans formation. In New York and Pennsylvania, however, the Helderberg plays an important part in cave formation.

The Oriskany formation, which succeeds the Port Ewen, varies along the strike from a siliceous limestone in the Nearpass Quarry area to a siliceous limestone grading upward into a sandstone near Flatbrookville. It is approximately 170 ft thick and contains three caves. The Oriskany is separated from the next carbonate unit by a thick section of shale known as the Esopus Grit.

The Onondaga limestone, overlying the Esopus - Schoharie rocks, contains much chert in its lower part. The maximum thickness of the Onondaga is close to 250 ft. Three caves are known to exist in the lower part of the formation.

The remainder of the Middle Devonian is represented by non-carbonate rocks containing no caves.

Cavern Development

Table 2 compares variations in passage length, pattern, and type among most of the known carbonate caves. Note that of the 32 caves not developed in the "Kittatinny", 24 occur in the Franklin marble. The median passage length of the Paleozoic limestone caves is almost twice that of either the "Kittatinny" dolomite caves or of the caves in the Franklin marble. Shorter passage lengths in the "Kittatinny" may have been caused by predominance of dolomite in the "Kittatinny", whereas the other cavernous formations (Table 3) are higher in calcium

TABLE 2. A comparison of New Jersey cave passage length, pattern, and type.

Formation	Number of caves	Total passage length	Average passage length	Median passage length	Caves that have branching passages	Caves that have bedding plane or low wide passage
		(ft)	(ft)	(ft)	(%)	(%)
Onondaga	2	420	210	..	..	..
Oriskany	3	235	78	..	..	..
Coeymans	1	170	170	..	..	..
Jacksonburg	2	625	313	..	..	..
Epler	20	793	40	13	33	42
Rickenbach	8	709	89	42	37	100
Allentown	19	2457	129	30	45	50
Leithsville	10	1671	167	68	50	63
"Kittatinny" (undifferentiated)	13	392	30	14	30	..
Franklin	24	2526	105	25	33	40
Limestone caves	8	1450	181	140	17	13
"Kittatinny" caves (dolomite)	70	6022	86	30	39	..

TABLE 4. A comparison of the susceptibility to solution of the "Kittitinny" formations.

Relationship of cavities and overburden with the Cambro-Ordovician Dolomite Formations in the Spruce Run Reservoir Area. — Relationship of cave length and passage type to formation.

Formation	Number of drill holes	Total footage of all drill holes	Number of cavities	Average footage between cavities	Total amount of void penetrated (ft)	Void per total depth (%)	Void per total amount of rock drilled (%)	Average cavity size (ft)	Total amount of overburden cover over rock (ft)	Average depth to rock (ft)	Number of caves	Total length of known cave passage (ft)	Average passage length (ft)	Median passage length (ft)	Predominant pattern: branching vs non-branching. Passages that branch (%)
Epler	66	6,428	90	71	263	4.1	4.6	3	717	11	20	793	40	13	33
Rickenbach	31	3,852	80	48	225	5.8	6.6	3	468	15	8	709	89	42	37
Allentown	65	7,344	127	54	385	5.4	6.2	3	1,093	17	19	2,457	129	30	45
Leithsville	9	1,035	13	80	67	6.5	9.4	5	314	35	10	1,671	167	68	50

carbonate. The Franklin marble presents more of a problem, because it generally contains a higher proportion of calcium carbonate than does the "Kittatinny" (Table 3). One explanation of the low average length of the Franklin caves is that a number of the caves included in the tabulation either are small cavities intersected during zinc mining or are larger caves which have been partially destroyed during quarrying operations. Another reason may be that fine-grained limestones dissolve more readily than coarse-grained ones (Rauch, 1970; Rauch & White, 1970).

The statistics on "Kittatinny" caves (Table 2) are especially interesting, because they show a trend contrasting with that generally found in statistics on caves in limestone rocks. A comparison of the passage lengths found in the different formations shows that the coarse-grained formations of the "Kittatinny" are the most cavernous of all. The coarse-grained Leithsville and upper Rickenbach contain the largest caves. The lower Rickenbach and the Allentown, similar in grain texture, have similar lengths of passage. The lower Allentown is very cavernous. Its caves have the greatest average lengths. One is over 1200 ft long. The fine-grained Epler, on the other hand, contains the shortest caves. Even here, most of the larger caves occur in the coarse-grained portion of the Upper Laminated member. The fine-grained units generally develop fissure passages with little branching development.

Table 4 contains some cavern development statistics based upon the exploratory test drilling program conducted at the Spruce Run Reservoir, Clinton. For this table, the results of 171 drill holes were used, distributed as follows: Epler 66, Rickenbach 31, Allentown 65, and Leithsville 9. Two variables were compared—the cavities in and the amount of overburden above each formation. The average distance between cavities in each formation demonstrates that the Allentown and Rickenbach contain the greatest number of cavities per unit of rock with the Leithsville and Epler containing the least; however, there were very few drill holes in the Leithsville. The ratio of percentage of void to the total thickness of rock penetrated, on the other hand, indicates much the same as does the cave data—the Leithsville is by far the most cavernous, the Rickenbach and Allentown are moderately cavernous, and the Epler is the least cavernous.

In regard to depth of overburden above the formations, the Leithsville is seen to have the greatest, almost double that of the others. The Allentown and Rickenbach have overburdens nearly equal in thickness; the Epler has the least amount of cover. The area was glaciated during one of the earlier glacial periods; therefore, most of the overburden has been derived directly from the parent rock. The figures in Table 4 indicate the amount of solution that has taken place at the surface since early- to mid-Pleistocene time. In accordance with the results shown in Table 2, the Leithsville formation is that which is the most susceptible to solution action.

During construction work at the reservoir (Markewicz, 1958-61), the overburden was removed from the area of the core trench excavation and profiles were drawn showing the configuration of the rock surface. The description which follows is based upon station footage. The section runs from the lower part of the Allentown to the upper part of the Epler.

The trench begins at Station 0+00 ft, in the lower part of the lower (Limeport) member of the Allentown, and continues to station 16+50 ft, cutting across the regional strike at a very slight angle. The surface is relatively smooth, except for one depression about 12 ft deep. This depression resulted from leaching in a small fault zone. From station 16+50 to 17+90, there are numerous holes six to ten feet deep caused by solution in the coarser facies of the dolomite. The section from station 17+90 to 21+10 is missing. The rock surface is very irregular due to a number of faults of small displacement which shattered the rock and allowed solution to progress with ease. From stations 21+10 to 23+90, a zone is crossed in which weathering extends more than 40 ft below the rock surface. The rock surface rises at 23+90, in the vicinity of the Limeport-upper Allentown contact, and contains irregular holes from there to 26+90, where it becomes smooth with a few pinnacles. This bedrock topography continues to 29+30. From 29+30 to 30+00, there is a 35-ft deep hole incised into a fault zone. Beyond this large hole, the surface generally is smooth. At 34+10 the surface slopes toward Spruce Run Valley. At 34+40, the top of the Allentown forms a smooth, rolling rock surface that descends into the stream valley.

The Rickenbach underlies Spruce Run Valley at this point. This rock was not exposed along the traverse segments from 36+00 to 38+50 and 40+60 to 42+00. In these areas are smooth sided rises or ridges. From 42+00 to 43+40, on the west side of the valley, a thick, well-healed breccia and mylonite zone in the upper Rickenbach is in fault contact with the Epler. There are no cavities, even though the Epler is a limestone at this point. The limestone forms a bedrock high to 44+40. From 44+40 to 47+00, there is a gentle rise. The latter part of the traverse segment lies along a gently-sloping, undulating surface. Several large cavities occur in a mylonite zone from 47+00 to 50+00. At 50+00, the Epler dolomite includes thin shaly subbeds containing narrow cavities. From 51+80 to 52+70 there is a large, deeply-weathered zone in a thick, shaly section. Beyond the shaly section is a series of narrow rises and falls in the bedrock surface. A depression occurs from 53+00 to 53+20. The overburden was not removed from station 56+50 to the end of the dam at station 67+50.

A pipeline trench approximately four miles south of Spruce Run Reservoir was dug from the South Branch of the Raritan River at Hamden to Round Valley Reservoir. At the eastern end of the pipeline, this was excavated in the Leithsville formation (Markewicz, 1963) for some 3,200 ft. The excavation ranges from 15 to 35 ft in depth, depending upon its topographic position. Toward the eastern end of the trench, the Leithsville formation, Hardyston quartzite, and Precambrian gneiss are all included in a large fault zone. About 800 ft of breccia-mylonite is present. The stratigraphic top (west end) of the section pertinent to this report lies at station 106+00, either at the base of the Allentown or at the upper limit of the Leithsville.

The Leithsville dolomite is overlain by Triassic shales resting on an eroded land surface probably formed during late Paleozoic time. From station 106+00 to 130+00, the jagged rock surface has a strong pinnacle topography controlled by the lithology of the beds. From station 130+00 to 142+00, the highly faulted and mylonitized dolomite contains slices of Hardyston quartzite and Precambrian gneiss. Precambrian gneiss is thrust over Leithsville Dolomite near station 132+00. Leigh Cave occurs about 150 ft to the north of this station, beneath the trace of this fault. In the Leigh Cave portion of the trench, numerous springs and open cavities were found.

Conclusions

From knowledge of the formations in which New Jersey caves occur some quantitative conclusions can be drawn concerning cave development in these formations. Limestone caves have longer average passage lengths than do caves in dolomite. This is in agreement with the relatively greater lengths of limestone caves in other areas. Little is known about the factors which may control the size and number of caves in the dolomites.

The Leithsville formation has the longest passage length and most solution potential, as shown in Table 4 under the column "Percent of void per total amount of rock drilled". Table 3 indicates that there is little difference in the chemical composition of the "Kittatinny" formations. The principal difference is the rock texture. The Leithsville and parts of the Rickenbach are the coarsest grained. The lower Allentown and lower Rickenbach both have medium to locally coarse textures, which gives them the same solution potential. The Allentown, with its greater relief, is more frequently exposed. Consequently, there is a better chance, physically, of finding caves in the Allentown and, because of its higher elevation, the solution openings generally have been washed free of mud. In contrast, the Leithsville, which has a thick overburden, probably contains many inaccessible caves that are filled with mud. The Epler is the least cavernous formation in terms of passage length. All of the known caves in it are located in the coarse-grained members of the formation. Field-checking of Epler formation caves indicates that the lower part of the Upper Laminated member contains most of the caves. This section locally contains medium-grained rock that is similar to the upper part of the Rickenbach formation.

The Black Jack and Harmonyvale members have no caves listed in the tables, but part of a cave at Sarepta Quarry in the Jacksonburg formation extends into the Harmonyvale. The major portion of the cave is developed in the lowest part of the Jacksonburg limestone, immediately above the unconformity separating it from the Epler. Even though the Black Jack member is very coarse, there is little hope for large caves due to the large amount of bedded and rugose chert present in the formation.

The effect of structure on the development of caves in the dolomite is best interpreted from evidence noted in the Spruce Run Reservoir core trench and the Round Valley pipeline trench. In the larger fault zones, where the rock is similar on both sides of the fault, a great deal of granulation, mylonitization, and recrystallization can be present. This rock in most cases is well healed and few caves will form. On the other hand, faults of limited displacement in carbonate rock tend to have an open fracture system, thus allowing easy leaching. Where major faults occur between two unlike rocks, such as gneiss and dolomite, there is a great deal of breakage and open fracturing due to the differential stress strength of the two different rock types. Fault zones of this type allow easy movement of ground water and there is a good chance for the development of large caves such as Leigh Cave.

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Minimum Diameter Stalactites

Rane L. Curl *

ABSTRACT

Assuming that stalactites grow into the aqueous space available to them at their tips, it is shown how the pendant drop controls the smallest possible equilibrium diameter of a soda-straw stalactite. By dimensional analysis, it is shown that there exists a characteristic Bond Number, Bo = $\rho g d^2/\sigma$, which determines their diameter. From experiments on drops formed on glass capillary tubes of different sizes, it is found that the Bond number for minimum diameter stalactites is Bo = 3.50. This gives a soda-straw diameter of 5.1 mm under ordinary conditions, agreeing with existing observations. Finally, it is shown that the diameter of a *non-equilibrium* stalactite should converge, with growth, in an exponential manner to the minimum equilibrium diameter.

Stalactites in caves are of great interest to the public, to cavers and to speleologists. They are seen as objects of beauty, as subjects for fanciful imagination, as mineralogical curiosities, and as indicators of factors of the cave environment. Except for facets of mineralogy and crystal structure, there does not appear to be much complexity to the story of stalactite morphology. This may be why their literature is relatively scanty. Moore (1962) treated the subject historically and mineralogically, and illustrated the basic features of stalactite crystal structure and growth. The "soda straw" stalactite, illustrated in Figure 1, is the simplest form: a tube of nearly uniform diameter, deposited from a pendant drop. Both Moore (1962) and Goodman (1966) say that the diameter of this tube is equal to the diameter of a drop of water, which seems rather obvious—until it is pointed out that the size of a drop of water depends upon the diameter of the tube from which it hangs. It is my purpose here to explore this "paradox" and to suggest some controlling factors in determining the smallest possible equilibrium diameter of a stalactite. Limiting consideration to the minimum diameter stalactite must necessarily reduce the problem of stalactite morphology to its simplest level. I hope that this may serve as a point of departure for future quantitative studies.

The first feature of soda-straw stalactite growth that seems evident from Figure 1, from the illustrations of Moore (1962), Goodman (1965, 1966), and from the many photographs that have appeared in various publications, is that the growing crystals at the tip are constrained to form within the boundary of the drop surface. That is, they do not appear to distort the shape of the drop by pushing against its surface from within, nor do they penetrate the interface. This observation is not contradicted by the observations of Went (1969), who found that fungus mycelium may guide some stalactite growth. I will therefore assume that the drop surface is a boundary for crystal growth and, conversely, that crystal growth does not directly affect the drop shape except as it determines the size of the tube from which the drop hangs.

The factors affecting the shape of a drop hanging from a rod or tube are the volume of liquid in the drop, its density, the acceleration of gravity, the diameter of the tube, and the surface tension of the liquid. The shapes that may occur are varied and it is useful to observe these on tubes of different diameters. For this purpose, drops were formed slowly on the ends of capillary tubes

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that had been cut off and polished. Figure 2 shows the apparatus used to carry out these experiments.

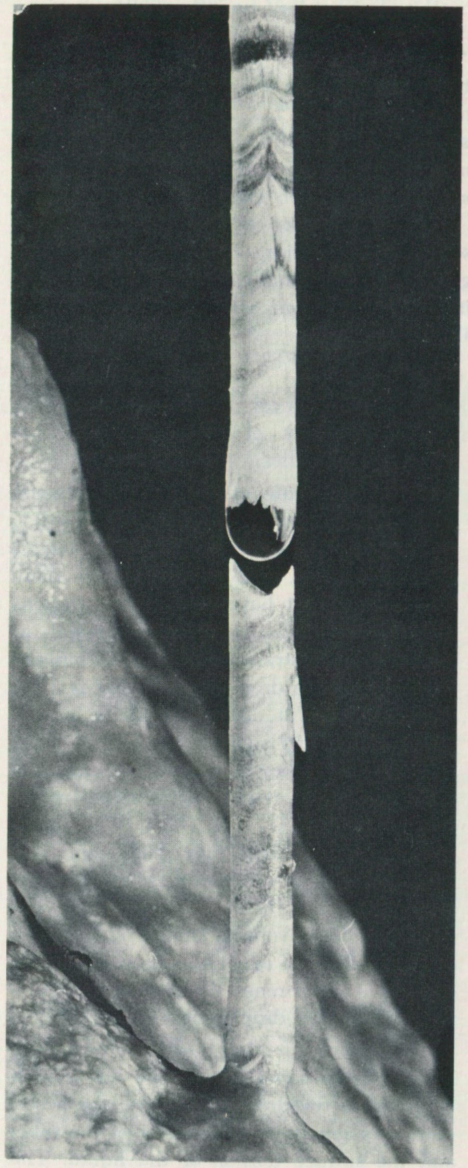

Fig. 1. *Soda-straw* stalactite. The original stalactite apparently broke or dissolved at one point, an extremely unusual event. Photo by Carl Kunath.

Drops were formed on tips made from various diameters of glass capillary tubing. The tubing was connected to a burette, allowing the adjustment and measurement of the water flow. The tip was enclosed within a bottle while in use in order to maintain a water-saturated atmosphere. Microscope slides cemented to the inside and outside of one side of the (square) bottle gave an optically undistorted view of the tip and of its pendant drop. The system was illuminated through an interposed heat-absorbing solution (copper sulfate). The image of the tip and drop was focused upon a flat surface (and enlarged about seven-fold) by a lens and mirror. By placing photographic paper in the plane of the projected image, photographs (negatives) of the drops could be obtained at various stages in their growth. These are shown in Figure 3.

From left to right are shown water drops forming on tips having diameters of 0.311 cm, 0.497 cm and 0.728 cm. Time, and hence drop volume, increases from top to bottom in each column of pictures. We see that on a "small" tip, the drop becomes larger in its maximum diameter than the tip itself, prior to forming a "neck", and at the middle stage bulges outward from its line of attachment to the tip. At this stage, if crystals were growing on the tip, they could grow into the drop and partly outward, *increasing the tip diameter*. Under the conditions of the experiment, therefore, this tube would be smaller than the final minimum size of a soda straw. In speaking here of "minimum diameter", I will generally mean the smallest *equilibrium* diameter, although it is possible for a stalactite to commence growth at a smaller diameter if the initial drop size is controlled by a small enough ceiling projection.

On the larger tip (0.728 cm), the drop always hangs in such a way that the drop surface slopes inward. The middle picture shows the drop at the condition of minimum inward slope. If the drop surface constrains crystal growth, as has been assumed, this tube is too large and growth would lead to a decreasing diameter. The intermediate diameter tube (0.497 cm) appar-

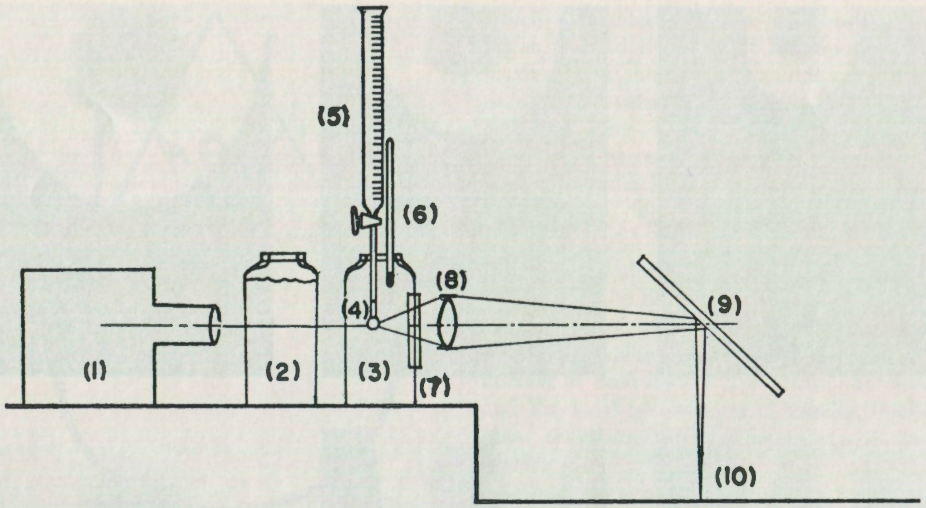

Fig. 2. Experimental apparatus. (1) light source; (2) heat absorber; (3) bottle in which drops are formed; (4) capillary tip and pendant drop; (5) burette; (6) thermometer; (7) microscope slides cemented to bottle side; (8) lens; (9) mirror; (10) surface on which image of tip and drop is brought to focus.

ently has nearly that diameter at which the maximum expansion of the pendant drop leads only to a vertical drop surface at the point of attachment. Under this condition, crystals can only grow vertically downward and the tube diameter will be maintained. This should be the condition for the maintenance of the *minimum equilibrium diameter stalactite.*

The included angle the drop surface makes at the point of attachment of the drop to the tip will be called θ, as shown in Figure 4. As the volume of the liquid in the drop increases, θ may be seen first to increase (Figure 3), then to attain a maximum value, θ_m, and finally to decrease until the drop falls from the tip. θ_m obviously depends upon tip diameter d. The factors determining the drop shape which have already been mentioned must be the same factors determining the angle θ. There must, therefore, exist a functional relation-

$$\theta = f\ (d, \rho, g, \sigma, v) \qquad (1)$$

ship where ρ is the fluid density (g/cm^3), g the acceleration of gravity (cm/sec^2), σ the surface tension of the fluid (g/sec^2), and v the drop volume (cm^3). Because the physics of pendant drops is well known, it is theoretically possible to calculate this relationship from first principles (Adamson, 1967) but this involves complex numerical calculations and has not been done for the present situation. It is simpler, as will be seen, to proceed experimentally.

Dimensional analysis (see Catchpole and Fulford, 1966, for references) then constrains the form of Equation (1) to one involving only dimensionless groups, such as those in

$$\theta = f\ \left(\frac{d^2\rho g}{\sigma}, \frac{d^3}{v}\right) \qquad (2)$$

The first group in the function has been called both a Bond number and an Eötvös number (Catchpole and Fulford, 1966). The use of the former name is older and will be adopted here. Let

$$Bo = \frac{d^2\rho g}{\sigma} \qquad (3)$$

which represents a dimensionless ratioing of gravitational to surface-tension forces. The condition for θ_m is, then,

$$\left.\frac{\partial \theta}{\partial v}\right|_{\theta = \theta_m} = -\frac{d^3}{v^2}\frac{\partial f}{\partial (d^3/v)} = 0 \qquad (4)$$

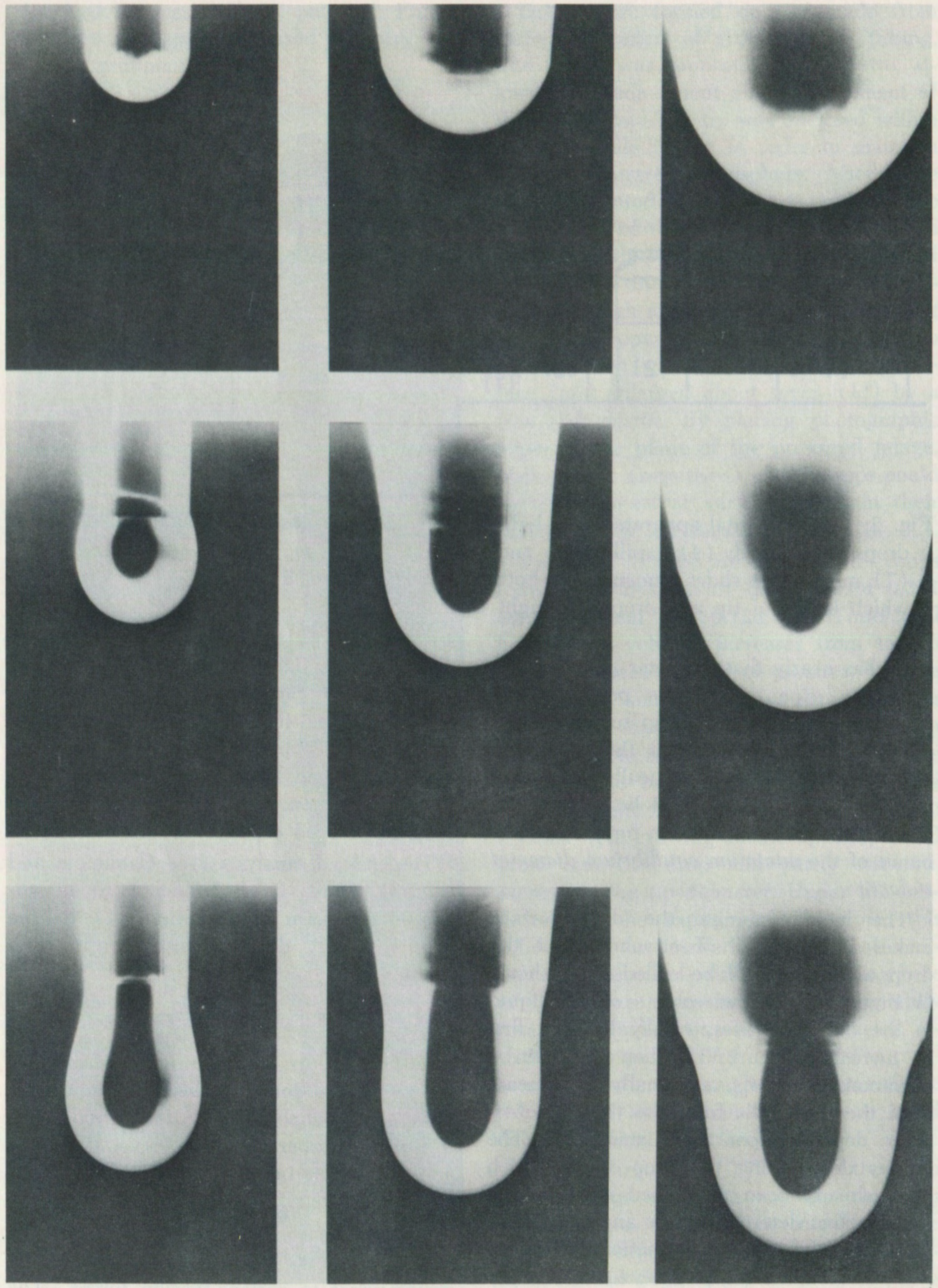

Fig. 3. Images of drops. Columns (left to right): tips of 0.311, 0.497 and 0.728 cm. Rows (top to bottom): liquid remaining after drop detachment, pendant drop at $\theta = \theta_m$, pendant drop just prior to detachment.

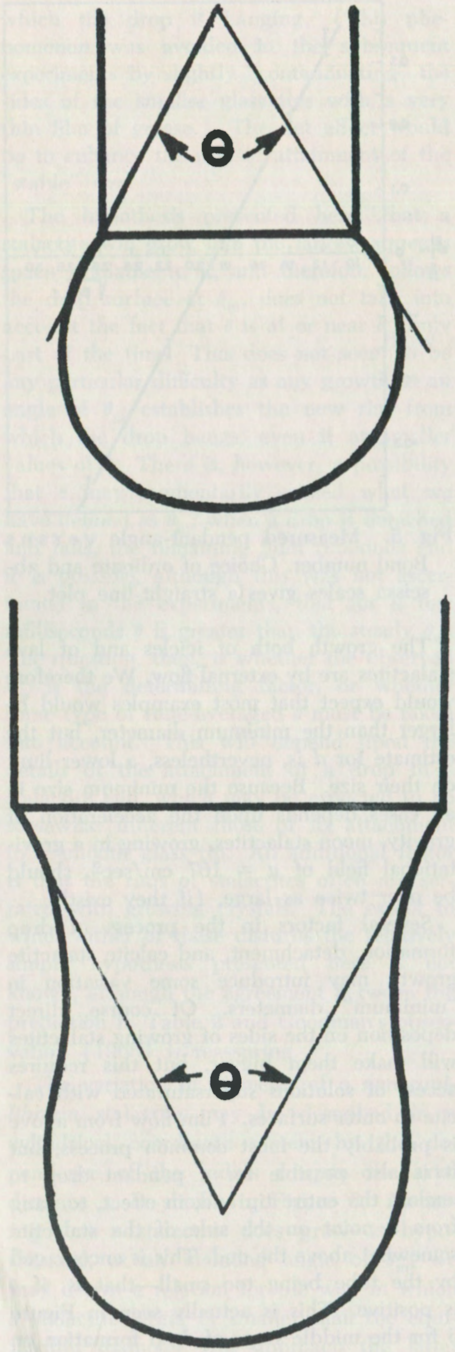

Fig. 4. Definition of angle θ. Above: θ positive. Below: θ negative.

which determines another relationship between Bo and d^3/v. That is, imposing the maximization introduces an additional relationship between the variables θ, Bo and d^3/v, allowing one to be eliminated. We conclude, therefore, that θ_m must be a function of the Bond number alone.

$$\theta_m = f\ (Bo) \qquad (5)$$

The problem, then, comes down to determining the value of Bo for which $\theta_m = 0$ (i.e., when the drop surface hangs vertically at its point of attachment, at maximum θ). When this value of Bo is known, a specification of fluid density and surface tension, and the local acceleration of gravity, determine the associated minimum equilibrium diameter.

Using the apparatus shown in Figure 2, θ_m was determined on a tip of a given size by allowing drops to form slowly (about two per minute), one after the other, and eventually to fall under their own weight. θ was followed on the projected image by holding the edges of pieces of paper tangent to the drop surface at its contact with the tip until the maximum angle was reached. θ_m was measured for two or more consecutive drops at the beginning of a series of about 60 drops, and again at the end. In this way, a standard error of measurement of θ_m could be estimated and confidence intervals (95% C.I.) evaluated. In addition, the average volume of the drops that fell from the tip was determined by measuring the total volume used from the burette and knowing the number of drops formed. The surface tension of the water was then determined using the drop-weight method of Harkins and Brown (1919). The density of water was taken as 0.998 g/cm^3 at 22°C and the acceleration of gravity as 981 cm^2/sec. The results of these measurements are shown in Table 1 and in Figure 5.

Surface tension is sensitive to temperature changes and impurities in the solution. It therefore was determined in the course of the experiments rather than assumed to be the value reported in handbooks. Tap water was used. The temperature in all experiments was about 21°-22°C. The measured values of σ are, happily enough, close to the reported value for pure water of 72.4 g/sec^2 at 22°C. An error analysis, including the effects of errors in the measurements of d and v, indicates a maximum error of about 3% in the determination of the Bond numbers.

In Figure 5, θ_m and Bo are plotted as $tan(\theta_m/2)$ versus $\sqrt{Bo}$. It was found, by trial and error, that this produced an essentially straight line plot. The point where the line intersects the abscissa represents the value of the "minimum stalactite diameter" Bond number, $\sqrt{Bo} = 1.87$, or $Bo = 3.50$. The equation of the line, which will be used later, is given by

$$tan(\theta_m/2) = -0.455\ (\ \sqrt{Bo} - 1.870 \qquad (6)$$

Knowing this minimum equilibrium diameter Bond number, we are able to calculate the minimum d for different circumstances and even for different materials. Examples for "soda straws", icicles and lava stalactites are shown in Table 2. The values are reasonable, although few actual measure-

TABLE 1. Experimental results

d (cm)	θ_m (°)	95% C.I. (°)	σ (g/sec^2)	Bo
0.311	37.1	±1.91	71.6	1.32
0.378	24.2	±0.94	72.7	1.92
0.497	1.8	±0.59	72.6	3.33
0.592	−16.6	±1.78	72.5	4.73
0.728	−39.1	±1.02	71.9	7.22

TABLE 2. Predicted minimum diameters

Type	Temp. (°C)	σ (g/sec2)	ρ (g/cm3)	d (earth) (cm)	d (moon) (cm)
Soda-straw	10	74.2	1.0	0.51	1.24
Icicle	0	75.6	1.0	0.52	1.26
Lava	1400	400	2.6	0.74	1.79

ments have been reported. Goodman (1966) shows two specimens, both almost exactly 0.50 cm in diameter, but does not report the temperature. The surface tension might also depend upon surfactants in solution. No data is available on the composition of lava stalactites, so the properties given in the Table are the density of plagioclase and the surface tension of a nominal blast furnace slag as given by Elliott, *et al* (1963). Lava stalactites are illustrated by Hicks (1950).

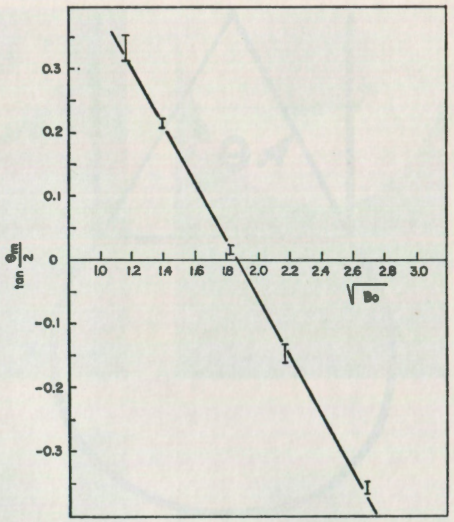

Fig. 5. Measured pendant-angle versus Bond number. Choice of ordinate and abscissa scales gives a straight-line plot.

The growth both of icicles and of lava stalactites are by external flow. We therefore would expect that most examples would be larger than the minimum diameter, but the estimate for d is, nevertheless, a lower limit on their size. Because the minimum size in all cases depends upon the acceleration of gravity, moon stalactites, growing in a gravitational field of g = 167 cm/sec^2, should be over twice as large, (if they exist!).

Several factors in the process of drop formation, detachment, and calcite stalactite growth may introduce some variation in "minimum" diameters. Of course, direct deposition on the sides of growing stalactites will make them bigger, but this requires access of solutions supersaturated with calcite to outer surfaces. Film flow from above is probably the most common process, but it is also possible for a pendant drop to enclose the entire tip and, in effect, to hang from a point on the side of the stalactite somewhat above the end. This is encouraged by the tube being too small—that is, if θ is positive. This is actually seen in Figure 3 for the middle stage of drop formation on the 0.311 cm tip; the light line above the end of the tip is the actual line along

which the drop is hanging. (This phenomenon was avoided in the subsequent experiments by slightly "contaminating" the sides of the smaller glass tips with a very thin film of grease.) The net effect would be to enhance the rate of attainment of the "stable" size.

The hypothesis presented here, that a stalactite will grow into the largest aqueous space available to it, and therefore follows the drop surface at θ_m, does not take into account the fact that θ is at or near θ_m only part of the time. This does not seem to be any particular difficulty as any growth at an angle of θ_m establishes the new rim from which the drop hangs, even if at smaller values of θ. There is, however, a possibility that θ may momentarily exceed what we have defined as θ_m; when a drop is detached and falls, the remaining fluid rebounds and it is possible, although this was not ascertained in the experiments, that for a few milliseconds θ is greater than the steady θ_m. The question, then, is whether the observed θ_m is the determining factor, or whether some type of time-averaged θ must be taken into account. This will depend upon the details of the attachment of a drop to a stalactite end, which may be expected to be somewhat different those of its attachment to a smooth glass tip. An additional factor is that the rims of stalactites often are serrated with growing crystals. The extent to which either of these disturbs the relatively simple hypothesis presented here is not known, although the agreement between the prediction in Table 2 and Goodman's observation (1966) is reassuring.

The variation in diameter of a *non-equilibrium* stalactite may be described in a cylindrical coordinate system by the dependence of the radius r upon the axial distance z, as shown in Figure 6. Assuming that a stalactite does grow as hypothesized, at an included angle of θ_m, we may derive a relation for the way in which a stalactite larger or smaller than the equilibrium diameter will approach the latter value. The identity between $tan(\theta_m/2)$ and the slope of the surface with respect to

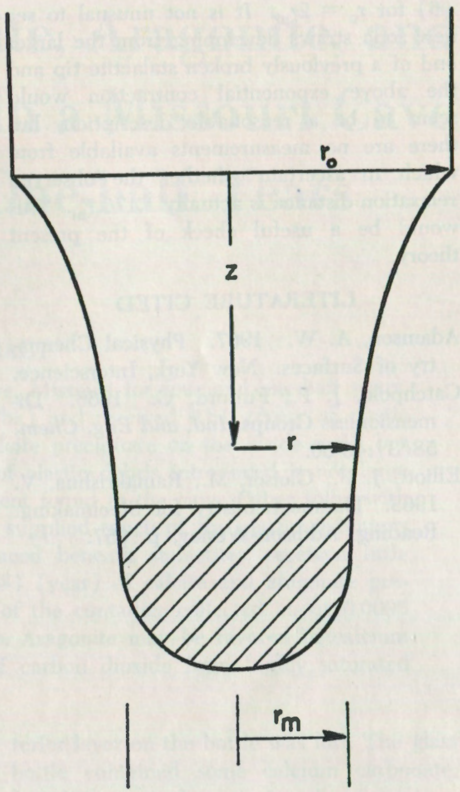

Fig. 6. Convergence of stalactite to equilibrium diameter when initially $r_o = 2r_m$ (pendant drop shaded).

the stalactite axis allows us to write, using Equation (6) and letting $d = 2r$.

$$tan(\theta_m/2) = \frac{dr}{dz} = -\ 0.91\ \left(\sqrt{\frac{\rho g}{\sigma}}\ r - 0.935\right) \qquad (7)$$

which is, fortuitously, a linear differential equation for r. Lettering r_m be the stable radius (obtainable from Bo = 3.50) and r_o the initial radius, the solution to Equation (7) is

$$\frac{r}{r_m} = 1 - \left(1 - \frac{r_o}{r_m}\right) \exp\left(-\ 0.85\ \frac{z}{r_m}\right) \qquad (8)$$

This states that the radius approaches the final radius exponentially. The characteristic "relaxation" distance is $r_m/0.85 = 1.176\ r_m$, which is the distance in which the *departure* from r_m decreases by the factor $e^{-1} = 0.368$. The result is shown in Figure

(6) for $r_o = 2r_m$. It is not unusual to see new soda straws developing from the larger end of a previously broken stalactite tip and the above exponential contraction would seem to be a reasonable description, but there are no measurements available from which to ascertain whether the observed relaxation distance is actually 1.176 r_m. This would be a useful check of the present theory.

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Deposition of Calcite, Aragonite, and Clastic Sediments in a Missouri Cave During Four and One-half Years

Max W. Reams *

ABSTRACT

Two containers were left in Cox Cave, Missouri for four and one-half years. One was placed in a solutional shaft (foiba) and received 0.81 g/year of clastic debris and 0.0011 g/(cm²) (year) of calcite precipitate on the inside walls but little on the outside walls. The volume of clastic debris introduced is not large enough to account for much of the sediment found in the cave. Other foibe with more direct surface connections probably supplied much of the clastic sediments to Cox Cave. The second container, placed beneath stalactites, received little clastic debris; it did receive 0.019 g/(cm²) (year) of calcite and aragonite precipitate on the inside wall. The exterior of the container collected about 0.0095 g/(cm²) (year) of nearly pure aragonite. Aragonite may be favored in calcium carbonate crystallization by rapid loss of carbon dioxide from highly saturated solutions.

INTRODUCTION

On April 28, 1966, two containers were placed in Cox Cave, Pulaski County, Missouri. A clean, 1-gal. plastic bleach bottle with the top removed was rinsed in cave stream water and placed beneath three actively dripping straw stalactites. A clean, 1-gal. glass jar was placed below an actively dripping area of a foiba (vertical solutional shaft formed by water dripping from a resistant bed) (Reams, 1965).

On November 21, 1970, the containers were emptied of most of their water and returned to the laboratory for study. During removal and transportation, most of the delicate precipitates of calcium carbonate on the inside and outside of the plastic bottle broke loose. The inner deposits fell to the bottom of the bottle and apparently received a thin coating of calcium carbonate as the remaining water (a few centimeters deep) evaporated. Much of the exterior layer on the bottle was lost. The glass bottle contained some calcium carbonate, but little was lost during transportation. Evaporation of the remaining water deposited little calcium carbonate.

CLASTIC DEPOSITS

The plastic bottle contained an estimated 0.01 g of silt and clay. The carbonates were slightly stained by ferric iron. Overtopping the container by the cave stream apparently did not occur. Flooding would have introduced clastic debris and probably would have moved the container. Enough sediment apparently is introduced by water dripping from stalactites to color speleothems and to account for part of the clastic material found in stalactites (Mills, 1965). Low-frequency, flash floods are probably necessary to account for the thick clastic layers in stalactites observed by Mills (1965).

Chemical precipitates were hand picked from the glass container placed in the foiba.

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The remaining clastic debris was scraped from the container and weighed on an analytical balance. The clastic debris totaled about 3.6 g or 0.81 g/year. The grain sizes ranged from chert fragments larger than 2 mm to rounded to subrounded 0.5 mm quartz grains to clay. The dominant size range was from coarse-to-medium sand to silt. The sample trapped by the bottle probably was biased because most clay and some silt probably washed over the lip. The bottle intercepted perhaps 20% of the total flow down the foiba. A complete sample with an equivalent sediment loss would have yielded 4 g/year to the cave detrital deposits. No sample was taken of the sediment in the foiba. The efficiency of the trap can be estimated by assuming that the bottle trapped only sand, which makes up five percent or less of some soils developed on carbonates in the Ozarks (Brydon and Marshall, 1958). With five percent trap efficiency, the sediment flux would be 80 g/year. If the sediment were to have 50% porosity and a dry density of about 2.1 g/cm^3 (U. S. D. A., 1966, p. 112), the volume of the sediment flux would be about 80 cm^3/year. If the present flux should be representative of past fluxes, about 8 m^3 of sediment would have been introduced in 100,000 years (probably a maximum time). This is enough to half fill a tube two meters in diameter (the approximate average diameter of the cave) and about five meters long. Cox Cave (several hundred meters long) was nearly filled with clastic sediment as evidenced by fill blockades. The number of foibe transporting sediment into Cox Cave at the above-measured rate, which would have been needed to fill the cave in 100,000 years, is far greater than is known to exist in the cave, given the above influx. Pleistocene climatic influences on sediment influx are unknown.

J. D. Vineyard (personal communication) found a foiba far upstream in Hamilton Cave, Washington County, Missouri, which directly underlies a surface stream. Gravel and other surface debris apparently enter the cave through this foibe during heavy rainfalls. The abundance of gravel in Hamilton Cave tends to support the theory (Reams, 1968) that Missouri cave sediments are transported surface soils which have entered the caves through direct and indirect openings to the surface. Foibe may contribute little or much to the total sediment flux, depending on such factors as surface connection, discharge, available surface materials, filtering by overlying sandstones, etc. Filtering by the Roubidoux Sandstone could be important in Cox Cave.

Carbonate Deposits

Mineralogy and chemical sedimentation rates in the two environments were very different (Table 1). Precipitates in the interior of the bottle in the foiba were dominated by isolated calcite crystals of about equal size. Little material was deposited on the exterior of the bottle, except in patches, probably because of the generally low dissolved-solids content of foiba waters (Reams, 1965). Deposition apparently took place in areas where flow was low (*e. g.*, in the recessed area between the jar threads and on certain spots on the exterior). The interior sedimentation rate is quite low (less than 1/17th), compared with that in the containers in the stalactite area.

The layer deposited on the outside of the bottle left in the stalactite area consisted mainly of aragonite and was deposited at a slightly lower rate than was the inside layer. The interior deposits varied greatly in thickness. Apparently, deposition was greatest just below the water level, where agitation was at a maximum and carbon dioxide was lost rapidly (clusters of calcite crystals were the main deposit in this area). In contrast, the upper four centimeters of the bottle in the foiba were essentially devoid of deposits. The deposition rates generally increase with bottle depth in the foiba area and decrease with bottle depth in the stalactite area. The latter observation was inferred from the deposits after they fell or were knocked loose. Iron staining of the deposits was more prevalent in the foiba experiment.

Calcite made up the bulk of the deposits inside the bottle in the stalactite area, but

TABLE 1. Chemical deposits

Property	Foiba Experiment Outside	Foiba Experiment Inside	Stalactite Experiment Outside	Stalactite Experiment Inside
Thickness of deposit	Patchy, thin	Discontinuous layer; crystals spaced at various distances	Thin, continuous layer 0.027 cm (mean of 17 micrometer measurements, range: 0.0090 to 0.055 cm)	Continuous layer 0.075 cm (mean of 39 micrometer measurements, range: 0.018 to 0.19 cm)
Calculated thickness (based on mass, assuming a density of 2.71 g/cm^3)	Insufficient data	0.0081 cm	Insufficient data	0.31 cm
Mass of deposit	Slight	4.8 g	Unknown	75.2 g
Surface area of deposit	Slight	980 cm^2	Most of exterior	890 cm^2
Rate of deposition	Slight	Depth below water surface (cm) — Rate of deposition: 0–4 — 0.000 g/(cm^2)(year) 4–6.5 — 0.00065 6.5–11.5 — 0.00091 11.5–18 — 0.00073 18–23.5 — 0.0019 23.5–24.5 — 0.0013 Average for whole bottle: 0.0011 g/(cm^2)(year)	0.0095 g/(cm^2)(year) (based on a small fragment)	0.019 g/(cm^2)(year)
Linear growth	Slight	0.0018 cm/year (calc. from mass using 2.71 density)	0.0060 cm/year (based on measured thickness)	0.017 cm/year (based on measured thickness) 0.0070 cm/year (calc. from mass using 2.71 density)
Mineralogy and crystal form	Calcite – most about 0.2 mm or slightly less, elongate. Aragonite – not observed	Calcite – most between 0.2 and 0.3 mm but slightly smaller at top, elongate. Aragonite – not observed	Calcite – not observed, but some found in x-ray pattern Aragonite – most about 0.02 mm laths	Calcite – most 0.2–0.4 mm, elongate Aragonite – base of deposit: most about 0.1 mm; coating calcite: most about 0.1–0.2 mm laths

aragonite frequently occurred as a thin mat of lath-like crystals against the wall, oriented parallel to that depositional surface. Calcite crystals were elongated perpendicularly to the wall. Aragonite also occurred as clusters of radiating needles coating calcite crystals which had fallen from the container walls. Interpretation of these details is difficult because the deposits were not studied in their original positions. The aragonite clusters probably grew as the solution which remained in the bottom of the bottle exaporated. Water clinging to the remaining wall deposits may have evaporated and left small aragonite crystals. Fragments knocked from the wall after the solution had evaporated received no additional coating.

Aragonite-Calcite Problem

Factors affecting calcite and aragonite deposition in cave environments are unclear. Observations made during this study may shed light on the problem. That aragonite was deposited on the outside of the container from the stalactite area and calcite was deposited on the inside suggests several possible interpretations.

Aragonite may have crystallized metastably in both environments, but constant contact with standing water may have caused it to invert to calcite. Calcite would then have dominated until evaporation of the remaining solution. Pobequin (1955) suggested that dry conditions favor aragonite and that wet conditions favor calcite. The relative dryness of the exterior of the container may have produced a supersaturated solution (the thin film of flowing water would allow rapid escape of carbon dioxide) capable of depositing either aragonite or calcite. Kinetic factors may have favored aragonite which, once formed, would continue to be deposited. The continuously wet interior of the container would favor calcite deposition, until evaporating conditions allowed aragonite to form. The much wetter environment of the foiba may have favored calcite on the interior and exterior of the glass bottle. Siegel and Reams (1966) evaporated natural cave waters at 25°C and found only calcite as a residue. Study of artificially prepared "cave" waters indicated that seed crystals may be very important in such environments. Temperature effects would be unimportant in the two environments studied here. Siegel and Dort (1966) observed cave deposits which formed in the last 150 to 200 years. Slowly oozing waters which evaporate quickly tended to deposit aragonite, and rapidly dripping water tended to deposit calcite. A deposit formed in a pool of water was all calcite.

The bottle beneath the stalactites represents the approximate conditions which might be found in a rimstone dam deposit. Detailed comparisons with rimstone dams might be very instructive.

The differences between the plastic container and the glass container may or may not be important factors in determining the mineralogy of the precipitates. The painted exterior of the plastic bottle was the site of aragonite dominance. Unsatisfied charge sites or scratches could influence nucleation.

Deposition Rates

Comparison of the chemical sedimentation rates of Table 1 with data from other caves given in Siegel and Dort (1966) (0.0018 to 0.017 cm/year) are similar. Linear growth is a somewhat deceptive measure. Mass per unit area measurements are more meaningful when considering growth. Unfortunately, such data is difficult to obtain. A long-term study of cave deposits *in situ* would be helpful in understanding sedimentation rates and should improve our knowledge of the factors influencing aragonite and calcite deposition.

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Effect of Small Thrust Faults on Cave Passage Cross-Section

Eberhard Werner *

ABSTRACT

Thrust faults whose displacement may be measured in centimeters and whose lateral extent may be a few hundred meters are very common in caves along the eastern edge of the Allegheny Plateau of West Virginia. These faults exert control over cave passage cross-section. Usually, the cave passage becomes slightly less in height and very much greater in width where the faults intersect the passage. Sometimes, there is an accompanying increase in breakdown.

The control of cave passages by geologic structure has been noted by most investigators of caves and karst. The relationship seen and mentioned most often is that of passage orientation to joints and faults. The effect of geologic structure on cave passage cross-section until recently was generally ignored. Variations in cross-section were assumed to have been caused by variations in lithology. In other words, cross-section was assumed to have been determined by stratigraphic controls, and orientation by structural features.

Recently, however, there have been investigations into the effects of geologic structure on cave passage cross-section. In particular, the effect of minor thrust faults, too small to be shown on the existing geologic maps, has been studied. Eddy and Williamson (1968) first reported thrust faults in Cassell Cave, Pocahontas County, West Virginia. Displacements there range up to 6 m, generally along the dip of the fault, and were sufficient to offset impervious beds, thus allowing groundwater to bypass these beds and to connect two or more levels. Rutherford (1971) noted that many thrust faults are exposed in the caves of Greenbrier County, West Virginia. Effects on cave passage ascribed to the faults are changes in cross-section, abundance of breakdown, and abrupt changes in passage character. Both of the above studies involve relatively complex caves, generally with large passages, secondary filling, and other characteristics of old, well-developed caves.

Thrust faults were observed in a study of the Cloverlick Valley, Pocahontas County, West Virginia. This area is located between those of the two previous studies (Fig. 1). The caves in Cloverlick Valley generally are small and very simple. They appear, for the most part, to be relatively young when compared to the caves described by the two previous studies. Since these caves are simpler, the effects of the thrust faulting should be easier to see in the Clover-

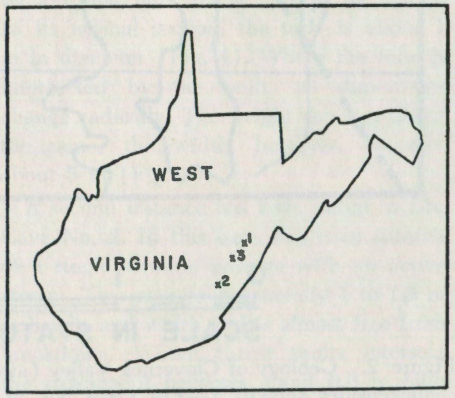

Figure 1. Location map showing study areas: 1—Eddy and Williamson (1968); 2—Rutherford (1971); 3—this paper.

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lick Valley caves than in those of other areas where the developmental history is more complex.

The caves studied are in the Greenbrier series of middle Mississippian age. Most of the caves occur in the Pickaway member, just above an impervious layer near the middle of that member. One cave studied is lower in the series, probably in the Taggard limestone. The caves are located at or near the axis of the Georges Creek (Potomac) syncline in rocks which dip 5° or less (Price, 1929). Figure 2 shows the locations of the caves and the areal geology of the Cloverlick Valley.

The faults usually are seen as slickensides on breakdown blocks and sections of ceilings (Fig. 3). The faults dip 0 to 20° in a direction perpendicular to the axis of the syncline. Most of the dips observed have been towards rather than away from the axis. Displacements have been difficult to measure because very few of the faults can be seen in cross-section, but in two instances the displacement was observed and measured. Movement along the plane of the fault, parallel to the dip, was 10 to 15 cm with a net vertical displacement of about 5 cm. The lateral extents of the faults have not been measured anywhere, but existing observations suggest that they are measurable in tens or hundreds of meters.

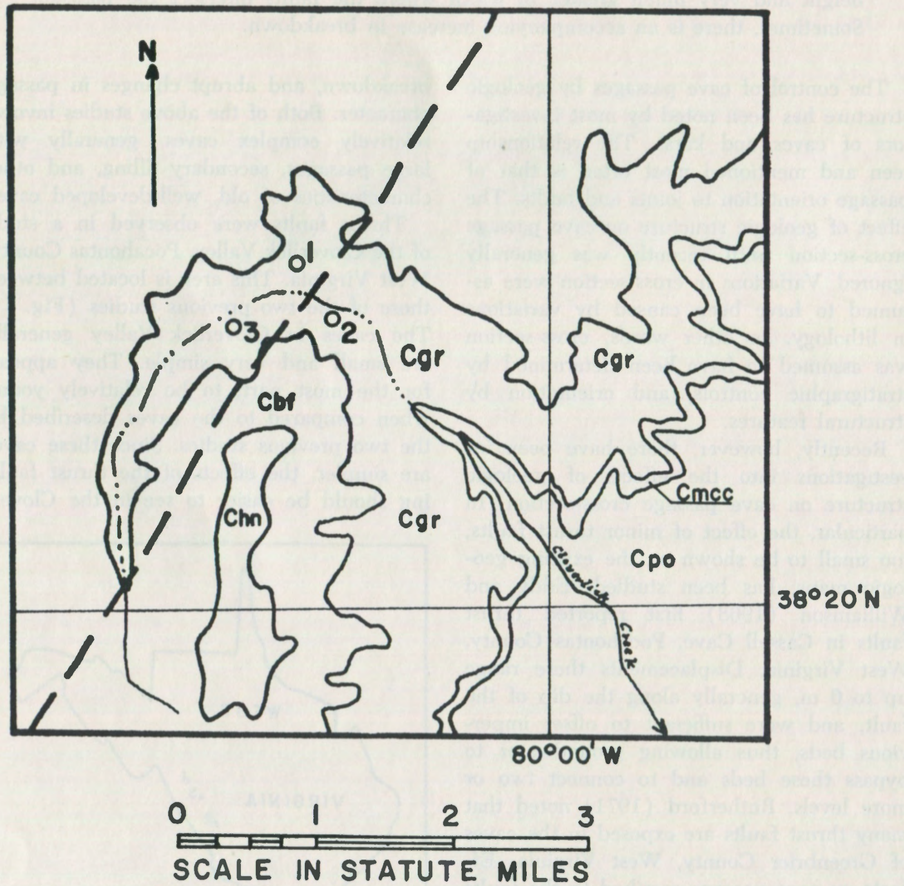

Figure 2. Geology of Cloverlick Valley (after Price, 1929). Location of caves with known slickensides shown: 1—Ebs Cave No. 2; 2—Rigg Cave; 3—Waterfall Cave. Abbreviations for formation names: Chn—Hinton Group, sandstones and shales; Cbf—Bluefield shale; Cgr—Greenbrier limestone; Cmcc—Maccrady red shale; Cpo—Pocono sandstone.

Figure 3. A slickensided ceiling in Rigg Cave. Dip shown is about the maximum found.

Effects of the faults have been observed in two of the caves in Cloverlick Valley. In Rigg Cave (see Fig. 2 for location), there is a solution tube which is intersected by a thrust fault for a part of its length. In its normal section, the tube is about 1 m in diameter (Fig. 4). Where the tube is intersected by the fault, its dimensions change radically. The height remains about the same; the width, however, becomes about 5 m (Fig. 5).

A second instance has been noted in Ebs Cave No. 2. In this case, the cave consists of a single solution passage with an active stream. The passage is generally 1 to 1.5 m in height and width and is almost free from breakdown. Where thrust faults intersect the passage, it becomes about 0.6 m high and 12 m wide (see Fig. 6). In this area, also, there are many large breakdown blocks on the floor.

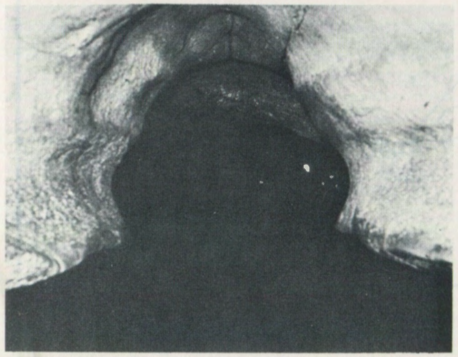

Figure 4. The solution tube in Rigg Cave.

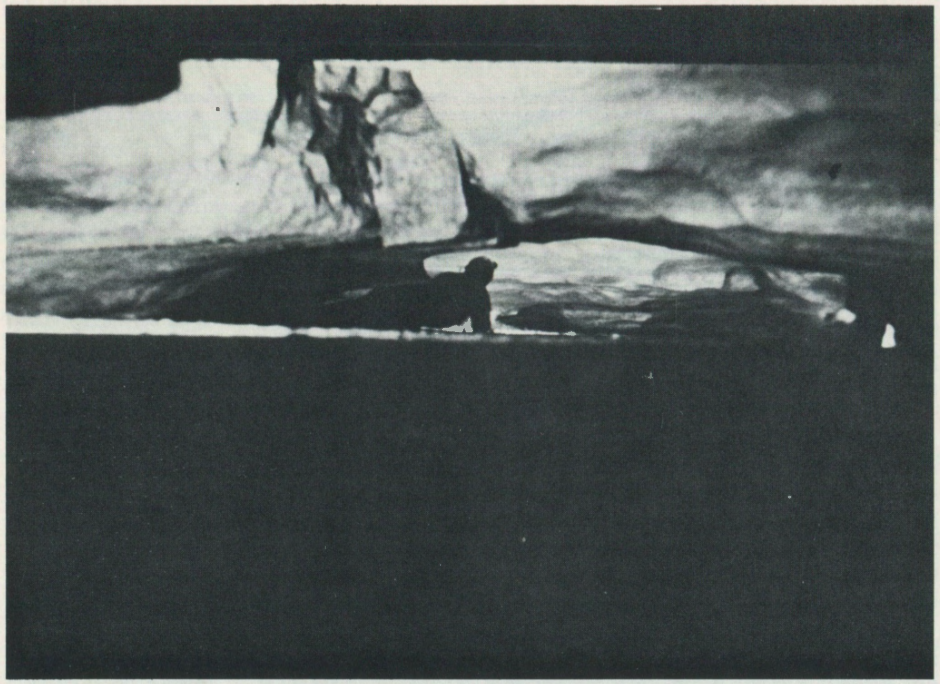

Figure 5. As Fig. 4, but 20 ft farther into the cave.

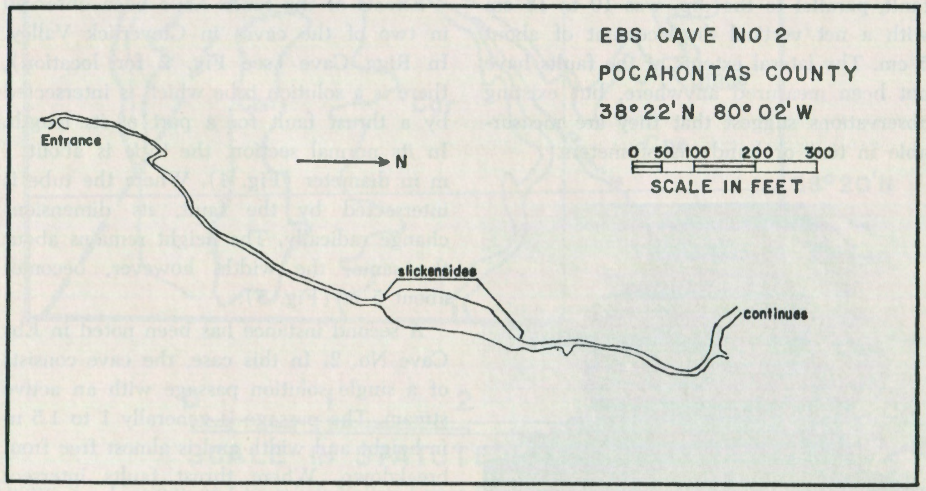

Figure 6. Map of Ebs Cave No. 2.

In summary, thrust faults are common features affecting cave passage cross-section. In relatively young caves, the effect is shown by a change in passage cross-section from one in which height and width are roughly equal to one in which width is several times as great as height. This change probably occurs because the fault weakens the rock, allowing breakdown to occur earlier in passage development. The breakdown then acts to block any water flowing in the passage, thus encouraging widening of the passage at the expense of deepening.

This phenomenon may underlie rooms in older caves which have no obvious explanation, such as those located at passage intersections or being enlarged domepits. The faults themselves may no longer be evident in these older caves if subsequent processes have removed or covered all visible traces of the fault planes.

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