



Cenozoic geomorphic history relating to Lewis and Clark Caverns, Montana
by Richard Bruce Aram

A thesis submitted in partial fulfillment of the requirements for the degree of MASTER OF SCIENCE
in Earth Science

Montana State University

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Abstract:

The structural and erosional history of the London Hills area strongly controlled the origin of Lewis and Clark Caverns. Shearing stresses during the Laramide orogeny created the small, plunging anticline on which the cave was later developed. However, significant ground-water flow and solution, necessary to form the cave, was prohibited during most of the Cenozoic by a thick cover of Tertiary valley fill. A late Tertiary fault elevated the western flank of the London Hills by at least 1,000 feet, but the uplift was quickly buried, allowing a major river to flow across the uplift. During the early Quaternary, the Jefferson River downcut more than 1,400 feet from this earlier level, forming the Jefferson River Canyon.

The resulting topographic relief powered groundwater flow through the Mission Canyon Limestone, causing the caverns to form on the northern lip of the canyon.

Lewis and Clark Caverns formed at the base of the Mission Canyon Limestone, where groundwater flow was concentrated above the less-soluble Lodgepole Limestone. Phreatic water descended across the steeply dipping limestone along many routes on both limbs of the anticline. These bedding plane passages met at the axis of the plunging fold, where trunk passages formed along a major joint system on at least two levels of the caverns, carrying the water to its discharge point in the Jefferson River Canyon.

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CENOZOIC GEOMORPHIC HISTORY RELATING TO
LEWIS AND CLARK CAVERNS, MONTANA

by

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A thesis submitted in partial fulfillment
of the requirements for the degree

of

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in

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ABSTRACT

The structural and erosional history of the London Hills area strongly controlled the origin of Lewis and Clark Caverns. Shearing stresses during the Laramide orogeny created the small, plunging anticline on which the cave was later developed. However, significant groundwater flow and solution, necessary to form the cave, was prohibited during most of the Cenozoic by a thick cover of Tertiary valley fill. A late Tertiary fault elevated the western flank of the London Hills by at least 1,000 feet, but the uplift was quickly buried, allowing a major river to flow across the uplift. During the early Quaternary, the Jefferson River downcut more than 1,400 feet from this earlier level, forming the Jefferson River Canyon. The resulting topographic relief powered groundwater flow through the Mission Canyon Limestone, causing the caverns to form on the northern lip of the canyon.

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CHAPTER 1

INTRODUCTION

Purpose

Up to 90,000 visitors per year have walked through Lewis and Clark Caverns since guided tours began in the 1940's. Despite this great interest in the cave, very little geologic research has been conducted to aid in its interpretation.

This thesis offers a geologic history of the caverns, based on the Cenozoic history of the surrounding London Hills area, as well as new mapping and discoveries in the cave. The author addresses the following questions:

What geologic factors account for the unusual shape and depth of the caverns?

When did the cave form?

What events in the regional geomorphic history caused the cave to form?

Previous Work

Martel (1905) published the earliest known scientific description of Lewis and Clark Caverns in the French speleologic journal, Spelunka. The cave was named Limespur

Cave at the time, and was described as 300 meters deep and 16 kilometers in length. These figures do not match the modern measurements of the caverns, but may prove to be accurate if future exploration connects the cave to the Jefferson River Canyon.

Chesnutt (1908) submitted a written geologic report and photographs to the United States General Land Office, encouraging the government to declare the caverns a National Monument. His actual report has not been located, but is referred to by many letters found during historical research by Mr. Ken Karsmizki (oral commun., 1979).

The Civilian Conservation Corps conducted several studies of the caverns during the time they developed it for tours. Petsch (1936) and Wegemann (1936) each wrote a short report which discussed the geology in general terms, but concentrated on the construction of facilities such as foot trails.

Perry (1946) suggested that the geologic structure and Tertiary history of the region were important in the history of the cave. However, his discussion is very brief.

Kuhns (in press) studied the chemical and clastic sediments of the cave, and derived a generalized history from them. He also made rough surveys of several rooms that

were not mapped by the Civilian Conservation Corps in the late 1930's.

To summarize, no detailed study has been made of the caverns geometry, geologic structure, and relation to the local geomorphic history.

Methods

The geologic structure and distribution of pre-Tertiary rocks have been mapped in detail in the London Hills area (Schmidt, 1975; Berry, 1942). The author compiled these and other maps for Plates I and II. Tertiary deposits have been mapped as a loose group in the Jefferson Island quadrangle, hence the study area (Parker, 1962; Richard, 1966), although they have been classified and mapped to formation rank in adjacent quadrangles (Robinson, 1963; Kuenzi, 1966). The author separated some of the Cenozoic deposits into formations, based on their lithologic characteristics.

The author and others remapped the entire known cave during the time between October, 1977, and May, 1979. Miss Linda Beer conducted a theodolite survey of the tourist trail through the cave, with a closure error of 0.02%. Off-trail sections of the cave were surveyed by Brunton and Suunto compass and cloth tape (Plate III). Strike and dip of the

bedding was added to the map (Fig. 27).

Setting

The London Hills area forms the rugged northern foothills of the Tobacco Root Range of southwestern Montana (Fig. 1). These hills separate two major valleys, the Jefferson and Three Forks. The Jefferson River connects the valleys through a 1400 foot deep canyon on the high western edge of the London Hills. Lewis and Clark Caverns formed on the northern lip of this canyon.

The London Hills form the southern wall of the Jefferson Canyon, and Cave Mountain makes up the northern wall. However, the London Hills, Cave Mountain, Doherty Mountain, and unnamed hills that extend to the north and east shared a common geomorphic history. For this reason, the author finds it convenient to simplify his terminology by calling all of them "the London Hills area" or just "London Hills" (Fig. 2).

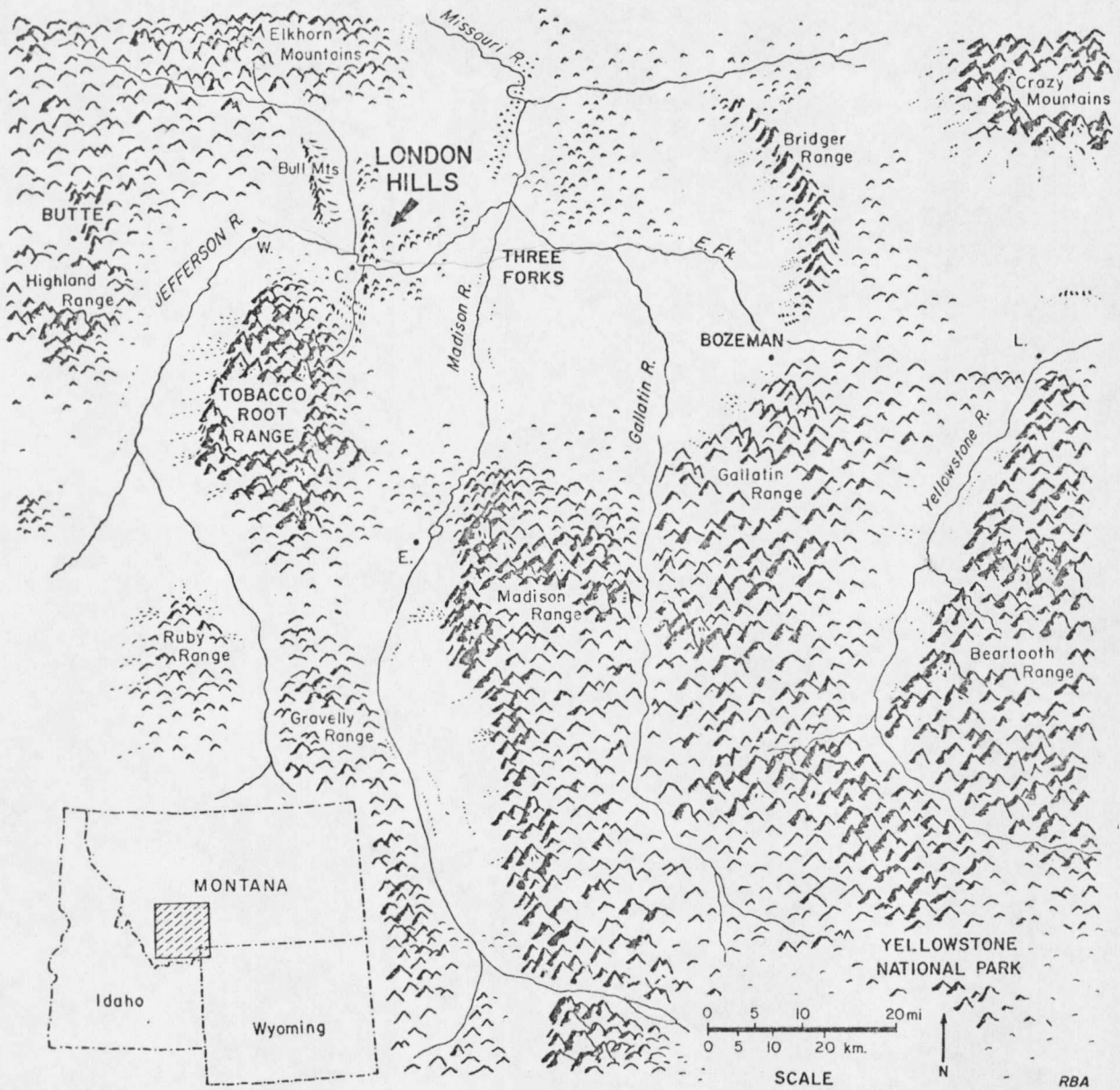


Fig. 1. Location map showing the London Hills of south-western Montana. Abbreviations of towns: Whitehall (W.), Cardwell (C.), Ennis (E.), Livingston (L.).

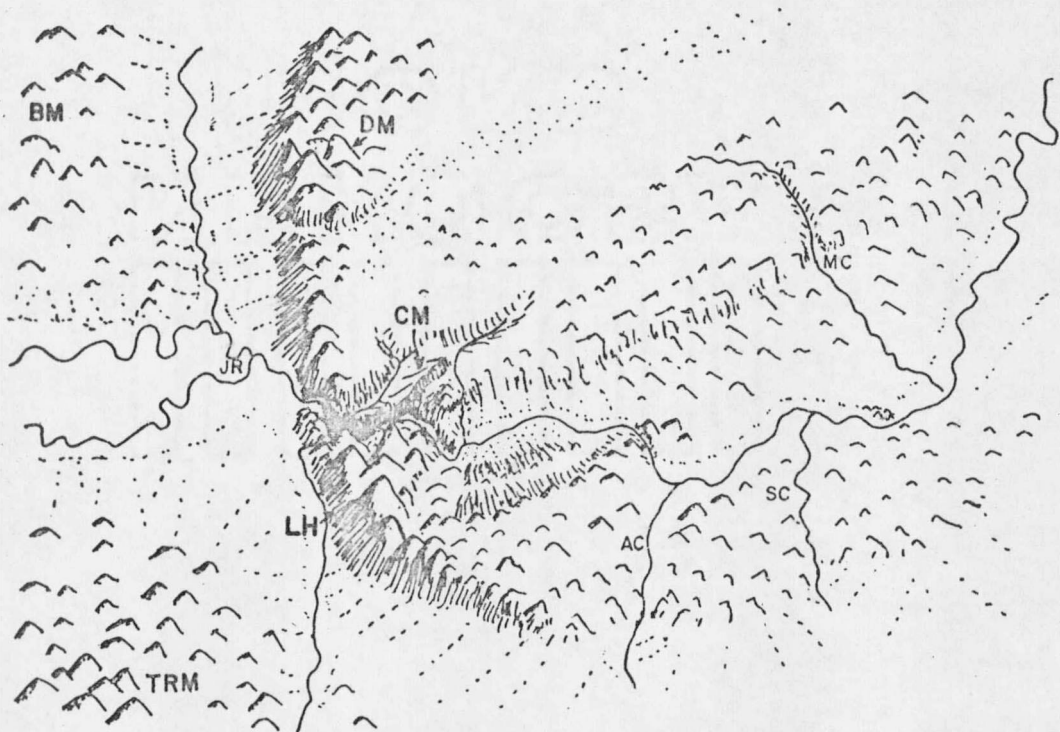


Fig. 2. Definition of the London Hills area as used in this paper. London Hills area includes Doherty Mountain (DM), Cave Mountain (CM), London Hills proper (LH), and the hills reaching east past Antelope Creek (AC), Sand Creek (SC), to Milligan Creek (MC). Not included in the London Hills area are the Bull Mountains (BM), Tobacco Root Mountains (TRM), and valley area west of the London Hills along the Jefferson River (JR). Width of map area equals 20 miles.

CHAPTER 2

GEOLOGIC HISTORY OF SOUTHWESTERN MONTANA

Introduction

Many episodes of high-grade metamorphism took place between 2700 and 1500 million years ago, and obscured the earliest geologic history of southwestern Montana (McMannis, 1965). These metamorphic rocks were buried by as much as 50,000 feet of sediments, called the Belt Supergroup, during the late Precambrian, during the time between 1500 and 850 million years ago (McMannis, 1965; Harrison 1972).

Stable cratonic conditions existed in the London Hills area during most of the Paleozoic and Mesozoic. However, the rocks deposited during these two eras increase in thickness by 300% within 100 miles to the southwest, indicating that miogeosynclinal conditions persisted nearby (Schmidt, 1975). The composition of sandstones changed from quartz to lithic by the end of the Mesozoic, due to immature sediments sweeping off the rising miogeosyncline. The Paleozoic Flathead and Quadrant orthoquartzites, and Mesozoic Swift and Kootenai lithic arenites illustrate the different conditions of deposition. In addition, limestones such as the Madison Group were common in the Paleozoic, but became less common, less pure, and thinner during the Mesozoic (Fig. 3).

Fig. 3. Stratigraphic section of the London Hills area (after Robinson, 1963; Kuenzi and Fields, 1971).

Era	System	Series	Group	Formation	Description
Cenozoic	Quaternary	Age relations complex		Young alluvium	0-50' (?) Rounded gravel and finer deposits on present flood plains of master streams; sub-angular gravel along tributaries.
				Fan & pediment gravel	0-50' (?) Sheets & wedges of colluvial unsorted angular gravel.
				Silt & fine sand	0-15' Unstratified angular eolian silt and fine sand.
				Old alluvium	0-400' Rounded gravel on former flood plains of master streams; subangular gravel in tributary valleys.
	Tertiary	Late Pliocene early Quat.			
		Miocene and Pliocene		Six Mile Creek Formation	0-2400' Arkosic & vitric sandstone & conglomerate
			Oligocene		Renova Formation
				Dunbar Creek Member	0-800' Light-colored tuffaceous siltstone, subordinate sandstone & conglomerate.
				Climbing Arrow Member	0-1,000' Olive bentonitic clay & coarse yellow sand.
Eocene		Malligan Creek Member	0-300' Light-colored tuffaceous lake limestone and interfingering stream-channel sandstone and conglomerate.		
		Sphinx Conglomerate	0-100' Pebbles and cobbles of Paleozoic limestone in a reddish-orange calcareous matrix.		
Mesozoic	Cretaceous	Upper Cretaceous	Elkhorn Mountains Volcanics	9,000'+ Andesite lava flows and volcanic breccia.	
		Lower Cretaceous	Kootenai Formation	650' (?) Orange sandstone with subordinate "salt-and-pepper" sandstone, and varicolored mudstone, siltstone, and limestone.	
	Jurassic	Upper Jurassic	Morrison Formation	100-250' Red, yellow, and brown mudstone, siltstone, and sandstone.	
		Middle & Upper Jurassic	Ellis Formation	75-350' Basal brown pebble conglomerate, brown sandstone, gray limestone, and yellow siltstone.	
Paleozoic	Permian	Lower Permian	Phosphoria Formation	75-200' Dolomite limestone, brown chert, dark phosphatic sandstone, yellow siltstone, and light-colored quartzitic sandstone.	
		Late Mississippian Pennsylvanian		Quadrant Formation	200-400' Light-colored quartzitic sandstone and subordinate dolomite limestone.
			Amsden Formation	250-600' Red sandstone and mudstone grading upward into pink and gray limestone and quartzitic sandstone.	
	Mississippian	Lower Mississippian	Mission Canyon Limestone	600-1,400' Gray thick-bedded limestone.	
			Lodgepole Limestone	350-650' Yellow-gray thin-bedded argillaceous limestone.	
	Devonian		Three Forks Shale	100-400' Upper orange siltstone; medial green shale; lower orange limestone and siltstone.	
		Upper Devonian		Jefferson Dolomite	350-650' Brown fetid dolomite with subordinate black and gray dolomite and gray limestone.
				Waywood Formation	0-200' Yellow, orange, and red calcareous siltstone, grading upward into gray limestone.
	Cambrian	Upper Cambrian	Pilgrim Limestone	350-450' Gray and orange dolomite, subordinate gray limestone.	
			Park Shale	100-200' Greenish gray shale.	
	Precambrian	Lower	Belt	Neagher Limestone	300-500' Gray and black limestone with orange mottling.
				Wolsey Shale	200-400' Olive and brown micaceous shale, with interbeds of glauconitic sandstone and limestone.
				Flathood Sandstone	0-100' Pink quartzitic sandstone.
				Empire Shale	600-1,000' Dark colored shale and argillite
			Lallood Formation	4,000' (?) Brown, coarse, micaceous arkose and conglomerate.	
				Crystalline metamorphic rocks	Thickness unknown- Gneiss, schist, amphibolite, pegmatite, and basic dike rocks.

Violent forces transformed the western margin of the continent at the close of Mesozoic times during the Laramide orogeny. The Cordilleran miogeosyncline, formerly sinking and collecting sediments, began to rise and shed debris to the east. Compressive forces and uplift migrated across the edge of the continent from the west, triggering the retreat of the sea. Great intrusions and volcanoes accompanied the compression. Reverse and thrust faults, as well as folds, crumpled the strata, shortening the crust by as much as 50-75 miles (Robinson, 1959).

Great thicknesses of Cenozoic deposits accumulated in the intermontane basins following the Laramide orogeny. Normal faulting, regional uplift, and superposition of the regional drainage system has since removed a large amount of these sediments.

Events that played important roles in the speleogenesis of Lewis and Clark Caverns will now be discussed in greater detail.

Precambrian Willow Creek Fault

During the deposition of the late Precambrian sediments, a small embayment reached eastward onto the craton in Montana from the Cordilleran miogeosyncline (Fig. 4A). The

majority of the 7,000 foot thickness of these Belt Embayment deposits are mudstones and limestones, indicating deposition in an environment with low water energy. However, the sediments found along the southern margin of the embayment are coarse-grained, including boulder conglomerates (Fig. 7A). These coarse clastics are named the LaHood Formation (Alexander, 1955), and occur in a narrow east-west band, 80 miles long and 2-8 miles wide (McMannis, 1963). The material becomes finer in size to the north and is absent to the south, where the Cambrian Flathead Sandstone rests unconformably on the Precambrian metamorphic rocks. These facts imply that an ancient mountain range, bounded on the north by a normal fault or strike-slip fault, formed the southern shore of the Belt Embayment, and shed coarse sediments into the embayment (Fig. 4B). The fault is named the Willow Creek fault (Robinson, 1963).

This fault can be located by geophysical mapping, but is not exposed at the surface (Davis and others, 1965) (Plate I). Because of this, the origin of the fault is obscure. The Willow Creek fault possibly exploited the foliation of the metamorphic rocks, which dipped to the north in the Tobacco Root Range (Fig. 4B) (Schmidt, 1975).

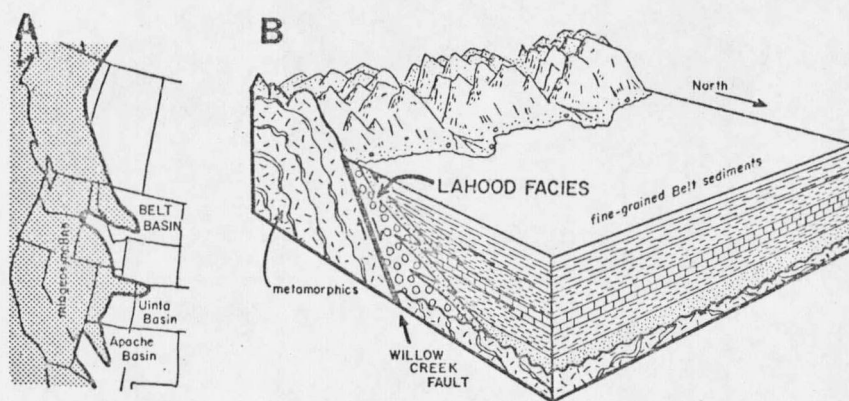


Fig. 4. A. Location of the Belt Basin and Cordilleran miogeosyncline. Stippled areas denote areas of late Precambrian deposition. B. Interpretive drawing of late Precambrian topography and sediment distribution along the southern shore of the Belt Basin. (A. from Harrison and others, 1974).

Mississippian Madison Group

The Madison Group was deposited during a rapid transgression and slow, subsequent regression during the Mississippian Period (Sando, 1976). The thin-bedded Paine Member of the Lodgepole Limestone was deposited first, while the water was deepest. It is dark colored, fine-grained argillaceous limestone (Smith, 1972). Layers of insoluble, clastic sediments separated the carbonate layers.

The Woodhurst Member of the Lodgepole Limestone was deposited when the water began to shallow. Coarse-grained limestone beds became thicker and more abundant, and the insoluble muds were slightly less abundant. The Woodhurst Member is light-colored, with thin to medium bedding (Smith, 1972).

As the sea shallowed, coarse-grained calcite sediments could accumulate without the interruption of the insoluble, clayey sediments that were common during the deeper water deposition of the Lodgepole Limestone. Thus the Mission Canyon Limestone is a thick-bedded, clean limestone. Crinoidal biosparites are the most common rock type in the lower 300 feet of the Mission Canyon Limestone (Sando and Dutro, 1974). Micritic limestone supercedes the fossiliferous deposits, showing that the sea continued to shallow, and restricted the supply of nutrients to animal life. Dolomite and evaporites cap the sequence, and were deposited near the end of the regression.

The retreat of the Madison sea exposed thousands of square miles to subaerial erosion. A great karst plain of caves and sinkholes developed. These solutional features were filled with angular limestone blocks and a clayey matrix. Kaolinite in the matrix suggests that a highly weath-

ered soil had developed (Roberts, 1966) before the deposition of the Big Snowy and Amsden sediments during the late Mississippian and early Pennsylvanian.

Laramide Orogeny

Folds

Two major groups of Laramide folds have been recognized in the northern Tobacco Root Mountains area (Schmidt, 1975). The folds of the first group are large, measuring up to several miles in width. These are gentle and nearly symmetrical. Their axes trend northwest, in response to the regional northeast-southwest compression. Two of these folds make up the London Hills (Plate II). Precambrian through Cretaceous rocks are exposed in the London Hills anticline, and primarily Madison Group rocks are involved in the London Mountain anticline. A third large fold, the Summit Valley syncline, forms the southern boundary of the London Hills, separating them from the Tobacco Root Mountains (Schmidt, 1975).

The folds of the second group are much smaller, and many are asymmetrical and tightly deformed. These folds are either directly or indirectly related to Laramide faults, but the folds of the first group had no relation to indi-

vidual faults (Schmidt, 1975). The Laramide faults probably occurred later in the Laramide orogeny (Fig. 5), because they truncate or offset folds of both groups.

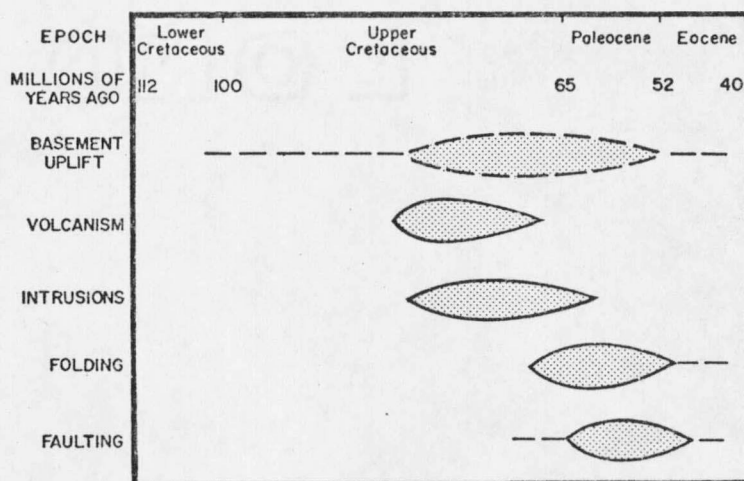


Fig. 5. Timing of Laramide events. (From Schmidt, 1975).

Faults

Most major faults in the northern Rocky Mountains trend northwest-southeast, but several large faults in the London Hills area strike east-west (Plate I). There is a close relation between the location and trend of these Laramide

faults with the Precambrian Willow Creek fault (Robinson, 1963; Schmidt, 1975). The Laramide movement possibly did not reactivate the actual Willow Creek fracture, but instead occurred at the depositional contact of the LaHood Formation and the metamorphic rocks (Schmidt, 1975) (Fig. 6).

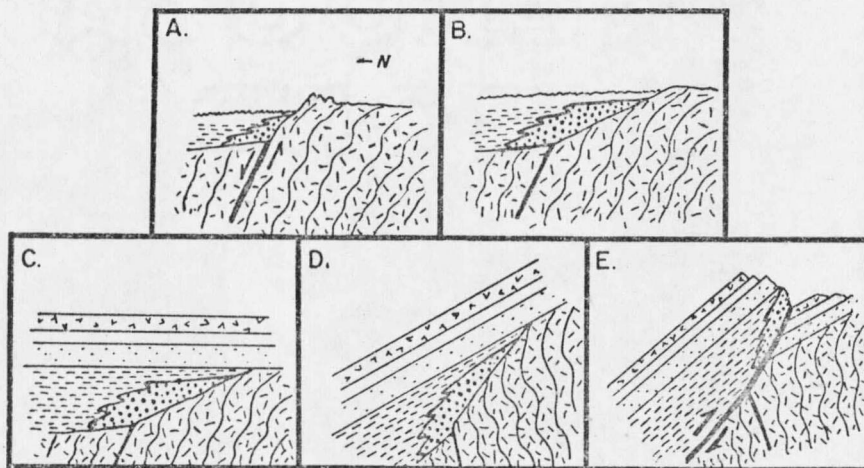


Fig. 6. Genesis of the Jefferson Canyon fault.
 A. Willow Creek fault moves, following foliation of metamorphics. LaHood conglomerates are deposited.
 B. Continued deposition buries the fault after it ceases movement. C. Deposition of Paleozoic, Mesozoic sediments and Cretaceous volcanics. D. Tilting to the north by Laramide orogeny. Willow Creek fault plane now dips to the south. E. North-dipping Jefferson Canyon fault follows the depositional contact at the base of the LaHood Formation. (After Schmidt, 1976).

The Cave fault placed the LaHood Formation against the Mission Canyon Limestone at the northern foot of Cave Mountain (Plate II). The fault splays at both ends, disappearing into the Greer Gulch syncline to the east, and is buried by alluvium to the west. The displacement had nearly equal amounts of reverse and strike-slip movement, and totalled between 3,900 and 7,800 feet (Schmidt, 1976). The offset decreased to the east, where the Jefferson Canyon fault parallels the Cave fault.

The Jefferson Canyon fault trends north-south at its western end, where it had small reverse movement. Across the Jefferson River to the north, the fault abruptly changes direction, striking east (Plate II). The offset increases from about 5000 feet at this bend in the fault to between 24,000 and 69,000 feet, five miles to the east (Schmidt, 1976). The fault had primarily strike-slip movement.

The increase of the amount of offset on the Jefferson Canyon fault to the east, and decrease along the Cave fault to the east, shows that both faults shared the same stress field where they overlapped (Schmidt, 1976). Cave Mountain occupies the sliver between these two faults (Plate II, Fig. 7B).

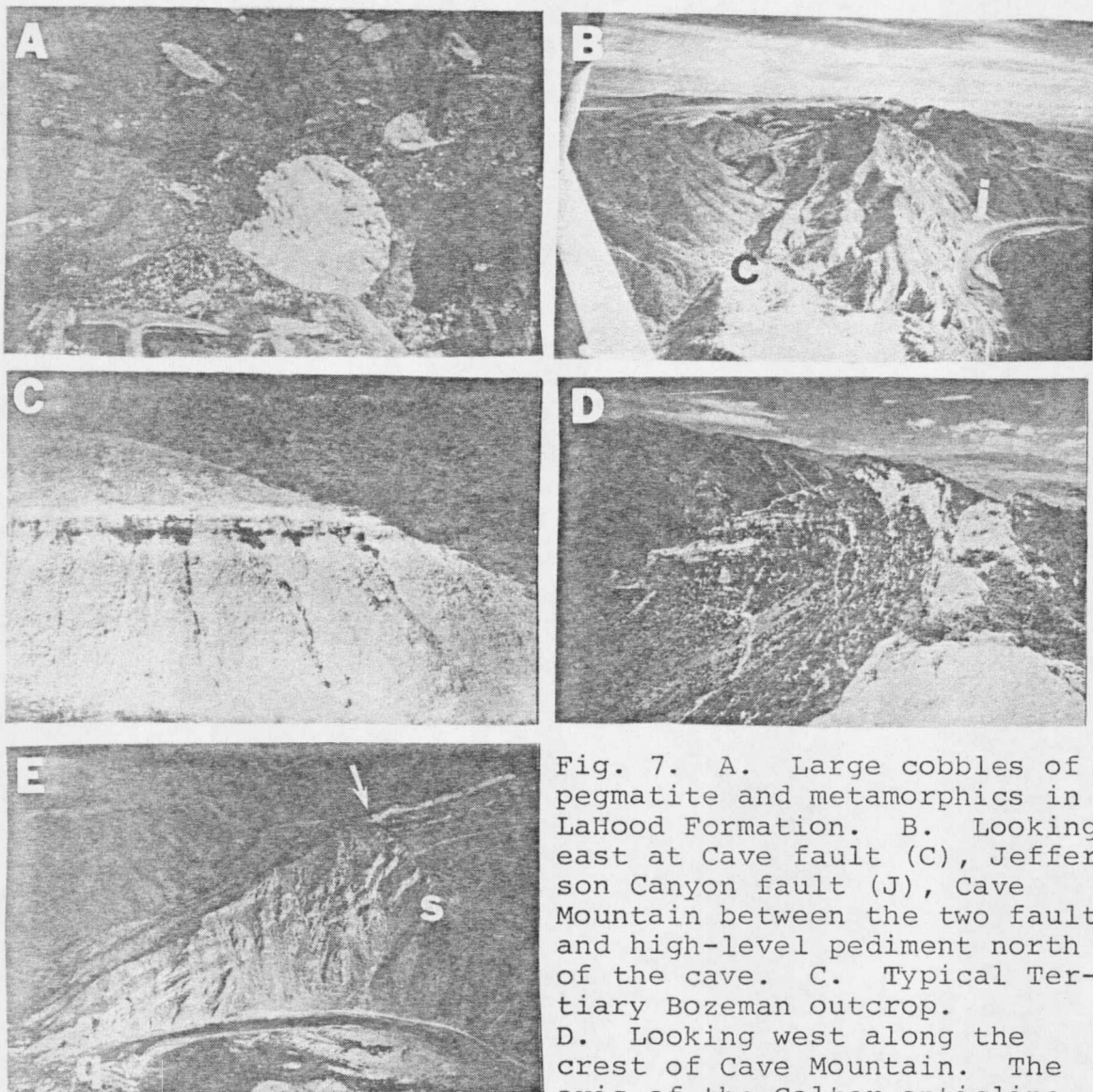


Fig. 7. A. Large cobbles of pegmatite and metamorphics in LaHood Formation. B. Looking east at Cave fault (C), Jefferson Canyon fault (J), Cave Mountain between the two faults, and high-level pediment north of the cave. C. Typical Tertiary Bozeman outcrop. D. Looking west along the crest of Cave Mountain. The axis of the Colter anticline

forms the valley seen in the middle background. E. Continuous outcrops of Madison rocks reaching to the canyon and Limespur quarry (q) from the cave area (arrow). Axis of Potts syncline forms the resistant ridge above J (Jefferson Canyon fault).

Tertiary Bozeman Deposits

Introduction

Laramide structural movements and subsequent erosion created many large basins in southwestern Montana. Great thicknesses of sediments accumulated in these basins during the Cenozoic. The sediments now have a maximum thickness of 7,800 feet in the Jefferson-Beaverhead Basin, and 5,000-6,000 feet in the Three Forks Basin (Burfeind, 1967).

Sediments that were deposited during the time between the end of the Laramide orogeny and the beginning of the Quaternary Period are called the Bozeman Group (Robinson, 1963) (Fig. 7C). Two distinct depositional sequences occurred in many of the intermontane basins. The similarity of rock type and age of these sequences across the entire region indicates that the different basins shared a common geologic history during the Cenozoic (Peterson, 1974).

Renova Formation

Rocks of the first sequence were deposited in a humid climate, so are mostly fine-grained and poorly consolidated (Kuenzi and Fields, 1971). These are called the Renova Formation, and accumulated during latest Eocene to late

Oligocene. Devitrified volcanic ash, as well as calcium carbonate, contributed a light color to the Renova Formation. The mudstones and limestones, common to this formation, were deposited in fluvial floodplains and channels, as well as lakes. The Renova Formation ranges from 0-3500 feet in thickness in the Jefferson Basin, where it is divided into three members, the Bone Basin Member, the Climbing Arrow Member, and the Dunbar Creek Member (Kuenzi and Fields, 1971). In the Three Forks quadrangle, three formations correlate with the Renova Formation: the Milligan Creek Formation, the Climbing Arrow Formation, and the Dunbar Creek Formation (Robinson, 1963).

Between the late Oligocene and late Miocene time, increased through-flowing drainage removed much of the Renova Formation from the region. The resulting unconformity separates the fine-grained Renova Formation from the overlying, coarse-grained Six Mile Creek Formation.

Six Mile Creek Formation

Deposition resumed in the late Miocene, and continued through the middle Pliocene. The drier climate produced coarse sediments which were better consolidated than those of the Renova Formation. The major rock types of the Six

Mile Creek Formation are arkosic sands and gravels. Volcanic ash is common, but the Six Mile Creek ash has fresher shards than those found in the Renova tephra. The rocks were deposited in river channels and floodplains, as well as by sheetwash on alluvial fans, near the edge of the basins (Kuenzi and Fields, 1971). The Six Mile Creek Formation ranges from 0-2,400 feet in thickness in the Jefferson Basin, and is not present in mappable quantities in the western half of the Three Forks Basin (Kuenzi and Fields, 1971; Robinson, 1963).

Late Pliocene-Pleistocene Events

Deposition of Bozeman sediments ceased and erosion began by late Pliocene time. This episode of degradation has continued to the present. Streams have downcut as much as 1,800 feet (Peterson, 1974). The streams cut through the Tertiary sediments, eroding deep enough in places to encounter older, resistant rock below. Seventeen major canyons were cut between the Tertiary basins (Fig. 8).

Normal faults, possibly active throughout the Tertiary, produced the basin and range topography of southwestern Montana. Glaciation modified the Tobacco Root Mountains, but did not extend onto the London Hills. Pediments and terraces planed the basins during the late Pleistocene.

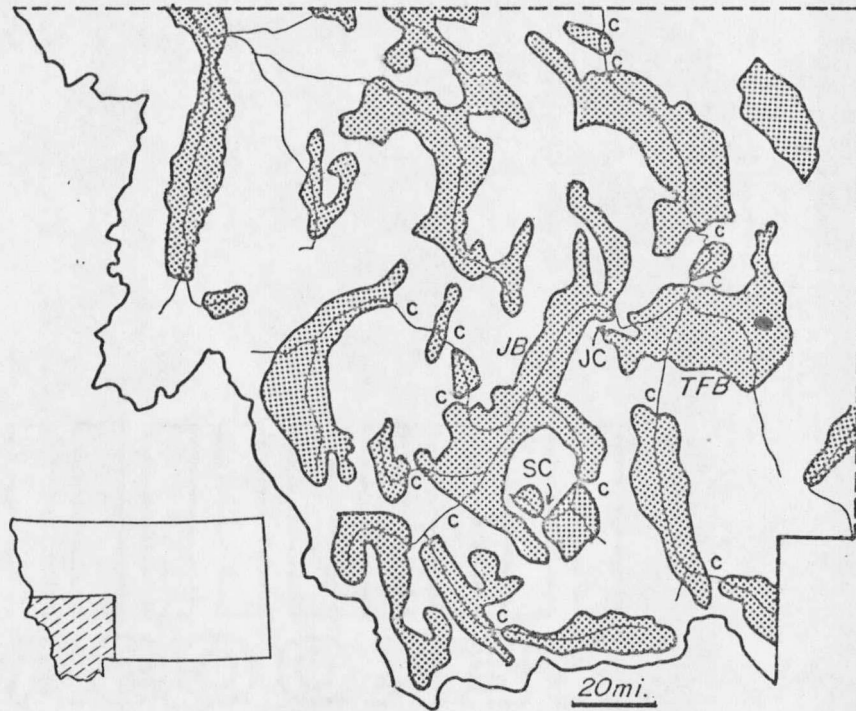


Fig. 8. Distribution of interbasinal canyons (c) along the headwaters of the Missouri River. Stippled areas show Cenozoic basins. Three Forks Basin (TFB), Jefferson Canyon (JC), Jefferson Basin (JB), Sweetwater Canyon (SC). (After Peterson, 1974).

Normal Faults

Late Tertiary normal faults border many of the mountain ranges of western Montana (Pardee, 1950). The Tobacco Root fault, west of the mountain range, and the Jefferson fault, north of the Tobacco Root Range, are younger than the Six Mile Creek deposits (Kuenzi and Fields, 1971). Normal faults offset alluvium that caps the superposed Sweetwater River Canyon (Fig. 8), thus the fracturing probably occurred after the cutting of the canyon (Peterson, 1974).

The western face of the Tobacco Root Mountains was uplifted above the basin by as much as 6,000 feet along the Tobacco Root fault. The offset on the Jefferson fault increased to the east, possibly to as much as 3,000 feet (Kuenzi and Fields, 1971).

Interbasinal Canyons

Although the headwaters of the Missouri River system primarily flow through broad basins that are filled with Cenozoic deposits, they also have cut narrow canyons through the intervening uplands. The canyons cut through rock ranging in age from Precambrian to Pliocene. The Sweetwater Canyon cuts through a basalt flow dated at 4.2 million years

(Peterson, 1974). This maximum age for that canyon has been extrapolated to all of the canyons in the Missouri headwaters, because the basins shared a very similar history during the Cenozoic (Peterson, 1974). Thus the Jefferson River Canyon in the London Hills is probably younger than 4.2 million years.

The canyons were probably cut by the superposition of the rivers due to regional rejuvenation (Peterson, 1974). The individual rivers of the Missouri headwaters system were presumably in grade together before the entrenchment, and are in grade now, thus all of the rivers downcut a similar amount. However, the resulting canyons do not have equal depths, but range from 600 to 1,800 feet in depth. This difference is the result of different thicknesses of Tertiary sediments over the canyon areas (Peterson, 1974). The thick accumulations of Tertiary sediments produced the shallow canyons, while thin blankets yielded deep canyons. According to this reasoning, the Jefferson Canyon area was buried by at least 400 feet of Cenozoic deposits, to an elevation of 6,000 feet, before the canyon was cut.

Quaternary Erosion Surfaces

Pediments cover most of the modern Jefferson Basin. Two surfaces were cut during the interval of time between the Bull Lake and Pinedale glaciations (Reshkin, 1963). Both surfaces slope an average of 3° , and are graded to the former, higher levels of the river.

The older surface, the Vendome pediment, stands an average of 50 feet above the Silver Star terrace. The terrace averages 25 feet above the modern floodplain. Both features are found to the east and the west of the Jefferson River Canyon, but have not been clearly identified in the gorge.

Glaciation

Glaciers of many different episodes covered the mountain ranges of southwestern Montana. Both Bull Lake and Pinedale glaciers reached down the South Boulder Valley from the peaks of the Tobacco Root Range (Roy and Hall, 1979), but no sign of ice cover has been recognized in the London Hills. Periglacial frost polygons are weakly developed on the Quadrant and Phosphoria rocks at an elevation of 6,000 feet in the London Hills, directly across the river from Cave Mountain.

CHAPTER 3

GEOMORPHIC HISTORY OF THE LONDON HILLS AREA

Laramide Structures of Cave Mountain

Regional Dip

The first Laramide deformation of the London Hills area produced broad folds, such as the London Hills anticline and London Mountain anticline (Schmidt, 1975). These folds imparted a steep northerly dip to the strata of Cave Mountain (Schmidt, 1976). Schmidt (1975) mapped a small, overturned syncline on the northern face of Cave Mountain, with an east-west axial trend. The author disagrees; although the beds change from a steep northerly dip to slightly overturned, dipping to the south, no evidence exists of horizontal beds between that would form the trough of the syncline. Rather the beds are vertical between the northerly and southerly dips, forming the overturned northern limb of the London Hills anticline (Fig. 10A&15C).

Colter Anticline

The Colter anticline and Potts syncline formed in response to the same stresses which later caused the Cave fault and Jefferson Canyon fault. The plastic deformation

preceded the brittle rupture (Fig. 9) evidenced by the Jefferson Canyon fault truncating Mississippian rocks along the southwestern limb of the Potts syncline (Plate II). Shearing along the fault planes did not significantly fold the rock further.

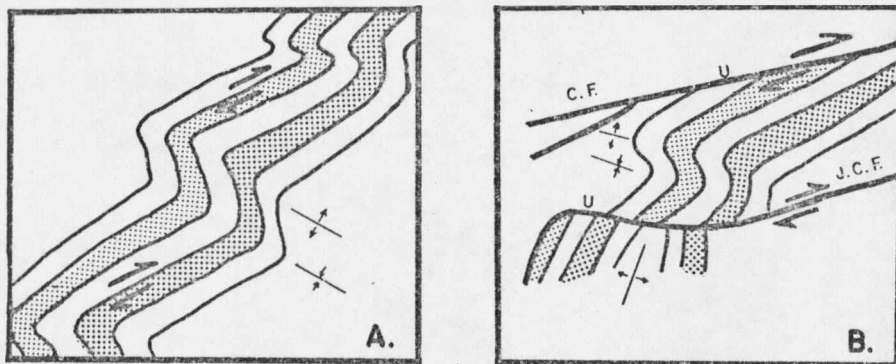


Fig. 9. Genesis of the Colter anticline. A. East-west shearing stresses create the Colter anticline and Potts syncline. B. Continued stress leads to fracturing, forming Cave fault (CF) and Jefferson Canyon fault (JCF).

The Colter anticline affects rocks ranging from the Mission Canyon Limestone to the Jefferson Dolomite, and probably some Cambrian units as well. The angle of dip of the bedding planes is about 54° on both limbs of the anticline, but the direction of dip changes from southwest to north-

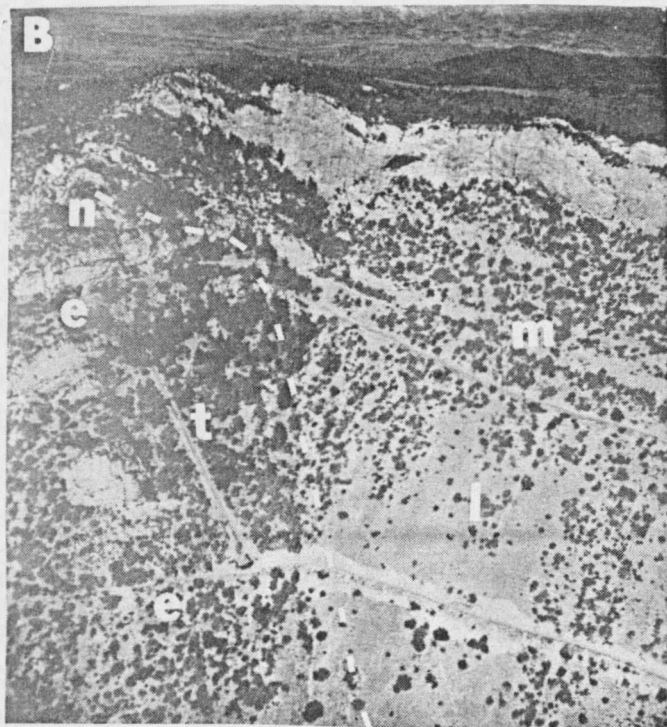
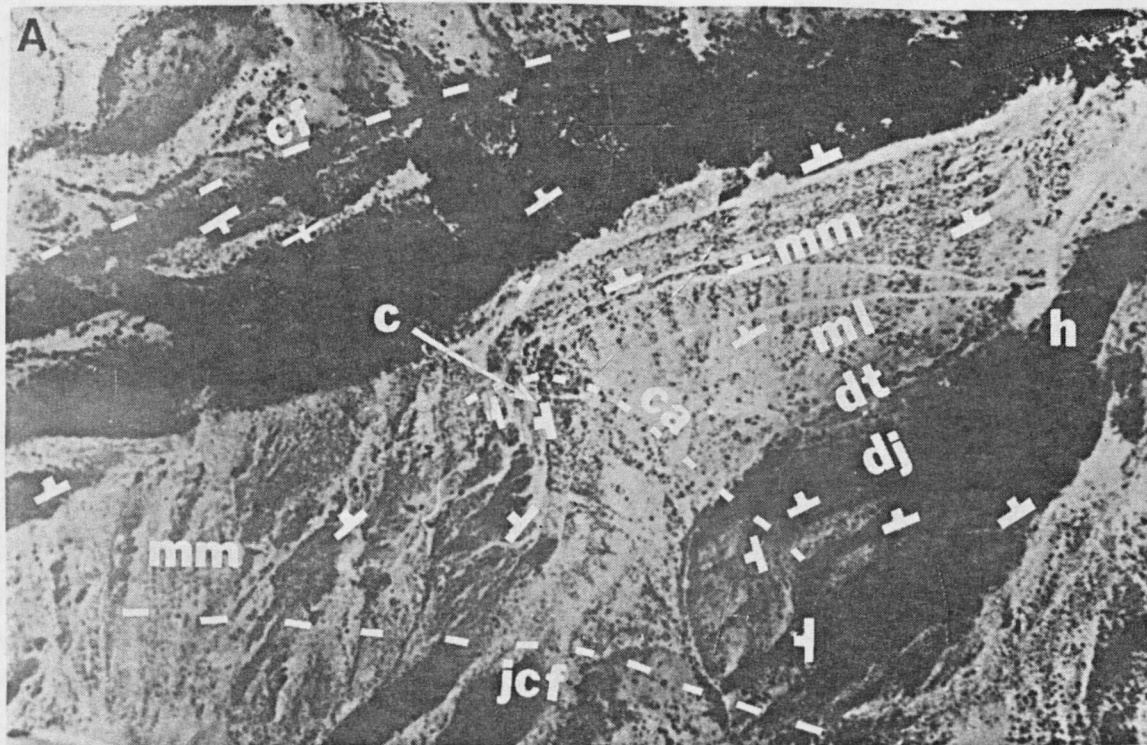


Fig. 10. A. Aerial photograph of Cave Mountain area. Cave fault (cf), Mission Canyon Limestone (mm), Lodgepole Limestone (ml), Three Forks shale (dt), Jefferson Dolomite (dj), Jefferson Canyon fault (jcf), caverns entrance (c), park headquarters (h), trace of axis of Colter anticline (dotted line near c), width of map 7,000 feet. B. Vertical view of axial plane of Colter anticline. Axis (dotted line) runs through the Nasal Passages (n) and near tram tracks (t). Upper entrance and lower exit (e).

west, and the strike changes from N. 20 W. to N. 60 E. (Fig. 10). Outcrops of the Lodgepole Limestone are discontinuous, and so make the change of strike at the fold axis difficult to observe, but the tracks of the vertical tramway for the caverns occupy the approximate position of the axis. The change of strike can be walked out in several other rock formations, such as the upper Jefferson Dolomite and the Mission Canyon Limestone, immediately above the Nasal Passages.

The axial plane of the Colter anticline dips about 70° to the east, from the Jefferson Dolomite through the basal Mission Canyon Limestone, but is nearly recumbent, dipping at $10-15^{\circ}$ to the east, from the caverns upper trail through the Nasal Passages, in the Mission Canyon Limestone (Fig. 10). Small folds near the fold axis occur in the Lodgepole Limestone, and further complicate the structure.

Tertiary Bozeman Group History

Distribution of Sediments

At least four valleys carried Tertiary streams through the London Hills area, and a fifth extended south of the Hills through the Summit Valley. Continuous deposits of Cenozoic materials extend from the Jefferson-Boulder basins to the Three Forks Basin through Negro Hollow (formerly

called Nigger Hollow) and the Eureka Basin, as well as the Summit Valley (Fig. 11). Tertiary materials in three other valleys, Antelope Creek, Sand Creek, and the Eureka Basin, reach across the London Hills from the east, but the deposits pinch out before completely crossing to the Jefferson Basin.

Bozeman Group rocks are absent in the modern Jefferson River Valley between the west end of the Jefferson River Canyon and Milligan Creek (Plate I). Bozeman materials, however, are found in the Jefferson Valley near Milligan Creek, where they are truncated by Quaternary alluvium (Robinson, 1963). Six miles to the southwest, Bozeman sediments are found in the valleys which carry Antelope Creek and Sand Creek, which are tributaries to the Jefferson River Valley. The presence of the Tertiary sediments at both ends of this portion of the Jefferson Valley suggests that the deposits were originally continuous between, but have since been eroded or buried by Quaternary fluvial processes.

The author found no deposits of the Renova or Six Mile Creek Formation along the modern Jefferson Valley west of Antelope Creek. Parker (1961) and Richard (1966) mapped Bozeman Group deposits along the north side of the Jefferson River between Antelope Creek and the Jefferson River Canyon, but the materials are clearly gravels of the Qua-

ternary Vendome pediment. The angular gravels are composed primarily of Paleozoic rocks found directly uphill of the pediment, and bear no similarity to the tuffaceous, fluvial sediments of the Renova or Six Mile Creek Formations. Most authors believed that the Jefferson River Canyon was cut during the Quaternary (Reshkin, 1963; Richard, 1966; Peterson, 1974), thus the narrow, western end of the Jefferson Valley between the canyon and Antelope Creek is probably also Quaternary, and so the absence of Tertiary Bozeman materials there is quite reasonable.

Flow Direction

Parker (1961) and others believed that Tertiary streams in the Jefferson Basin west of the London Hills flowed to the west. Kuenzi (1974) disagreed, finding evidence that the streams flowed east throughout the Tertiary, similar to the modern paths. The author supports Kuenzi's theory of easterly drainage.

Early geophysical work located a buried Tertiary valley in the Jefferson Basin, which was named the Cardwell depression (Parker, 1961). His data showed that the depression shallowed to the east, north, and south. Thus, Parker concluded that the Tertiary stream that eroded this buried

channel drained to the west, because it could not have flowed uphill to the east. Later geophysical studies agreed that the Jefferson and Boulder drainages flowed to the west (Wilson, 1962; Parker, 1962; Richard, 1966; Burfeind, 1967). Richard (1966) believed that the London Hills area served as a major drainage divide, thus Tertiary streams in the Three Forks area, Summit Valley, and Eureka Basin flowed to the east. However, he failed to explain how the former westerly flow reversed directions during the Quaternary.

Kuenzi (1974) believed that streams during the Tertiary flowed to the east, just as they do today. His conclusion was based on detailed analysis of the sediments of the Bozeman Group. However, his findings still do not answer the objections to an easterly drainage raised by Parker (1961) and others. Wilson (1962, p. 39) hinted at the solution,

"...none of these possible outlets could have drained the North Boulder Valley, unless structural movements raised them with respect to the center of the valley, since the time of erosion."

The data, although not the interpretation, of Parker (1961) and others may reconcile with the findings of Kuenzi (1974) if late Tertiary structural movement of the London Hills can be proven.

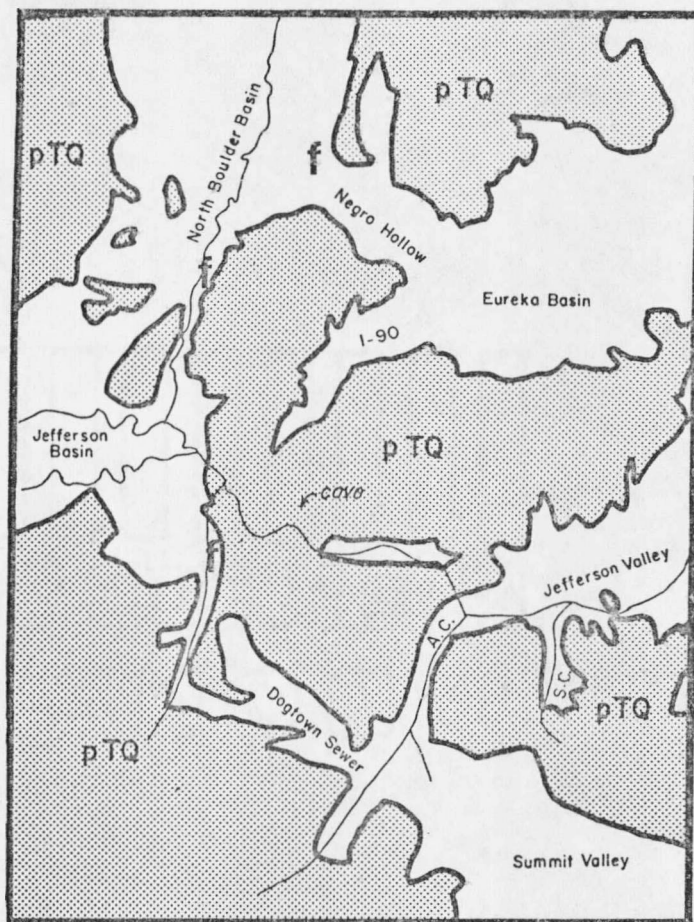


Fig. 11. Distribution of Cenozoic sediments in the London Hills area. White areas show cover of Tertiary and Quaternary sediments. Stippled areas show exposures of pre-Tertiary & Quaternary rocks. Sediments reach across from the Three Forks Basin through Sand Creek (SC), Antelope Creek (AC), and along Interstate 90 from the Eureka Basin, but pinch out before reaching the Jefferson Basin. No such valleys extend east from the Jefferson Basin, indicating tilting to the east from the London Hills area by the Starretts Ditch fault (f) along the western margin of the London Hills area. (After Davis and others, 1965; Richard, 1966).

Starretts Ditch Fault

Introduction

The western face of the London Hills rises sharply above the alluvial plains of the Jefferson and Boulder Valleys, and appears to be an uplifted fault block (Fig. 12A). North of the Jefferson River the break in the topography generally follows the strike of several rock units, and could possibly be interpreted as differential erosion. However, south of the Jefferson River, the break truncates the LaHood Formation, the Cave fault, and the upper Paleozoic sequence (Plate II). Pardee (1950) believed that a normal fault, with an offset of 1,000 feet along the western edge of the London Hills area, created this landscape during the late Tertiary. The author found three outcrops which support this interpretation. The fault will be called the Starretts Ditch fault in this paper, named for an old gold flume that traverses the western flank of the London Hills.

Location I

A linear feature that resembles a fault scarp extends several thousand feet along the foot of the London Hills in sec. 36, T. 2 N., R. 3 W. (Fig. 12B). On the ground, an

abrupt change of slope in the Precambrian Grayson Shale, approximately 30 feet high, marks this feature (Fig. 12C). The bedding of the shale is nearly vertical, and the scarp follows its strike. Slickensides are abundant in the colluvium, but none could be located in place to determine whether the direction of movement matched a normal fault or earlier folding. In places, the Vendome pediment covers the scarp.

In this general location, Alexander (1955) mapped an extension of his "Jefferson Island fault", a Laramide thrust which placed the Grayson Shale over Cambrian units at Doherty Mountain, one mile north of location I. A possible compromise between Alexander's belief that the fault in section 36 is a Laramide thrust, and the author's interpretation that it is a late Tertiary normal fault is that the normal fault exploited an earlier thrust fault.

This situation occurred in the disturbed belt of Wyoming, where the hanging walls, which were the western blocks, were upthrown during the Laramide orogeny, but downdropped along normal faults during late Tertiary time (Blackstone, 1977). This same history may have occurred in southwestern Montana, and may possibly explain why many of the normal faults dip to the west, uplifting the western margin of the mountain ranges.

Other workers have interpreted faults that are quite different than those of Alexander (1955) in the complex structure of Doherty Mountain (Berry, 1942; Perry, 1950). Thus, the fault at location I possibly had no thrust movement. Regardless, the western block at this location is topographically lower than the eastern block, which strongly suggests normal offset.

Location II

Bozeman strata dip as much as 45° to the east into the London Hills front in the SE $\frac{1}{4}$, sec. 14, T. 1 N., R. 3 W. Quaternary alluvium caps the outcrop, so the horizontal distance between the outcrop of dipping strata and the location of the fault cannot be absolutely determined. However, outcrops of the LaHood Formation found along the driveway of Mr. Vern Shaw, approximately 2,000 feet to the ESE, suggest at least 1,000 feet of throw at this location.

The general lithology of the tilted Bozeman materials matches the Six Mile Creek Formation. However, no diagnostic fossils were found to prove its age. The volcanic ash is not devitrified and occurs in distinct, well-consolidated beds, unlike the dispersed and altered tephra normally found in the Renova deposits. Gravel and coarse sand are abundant

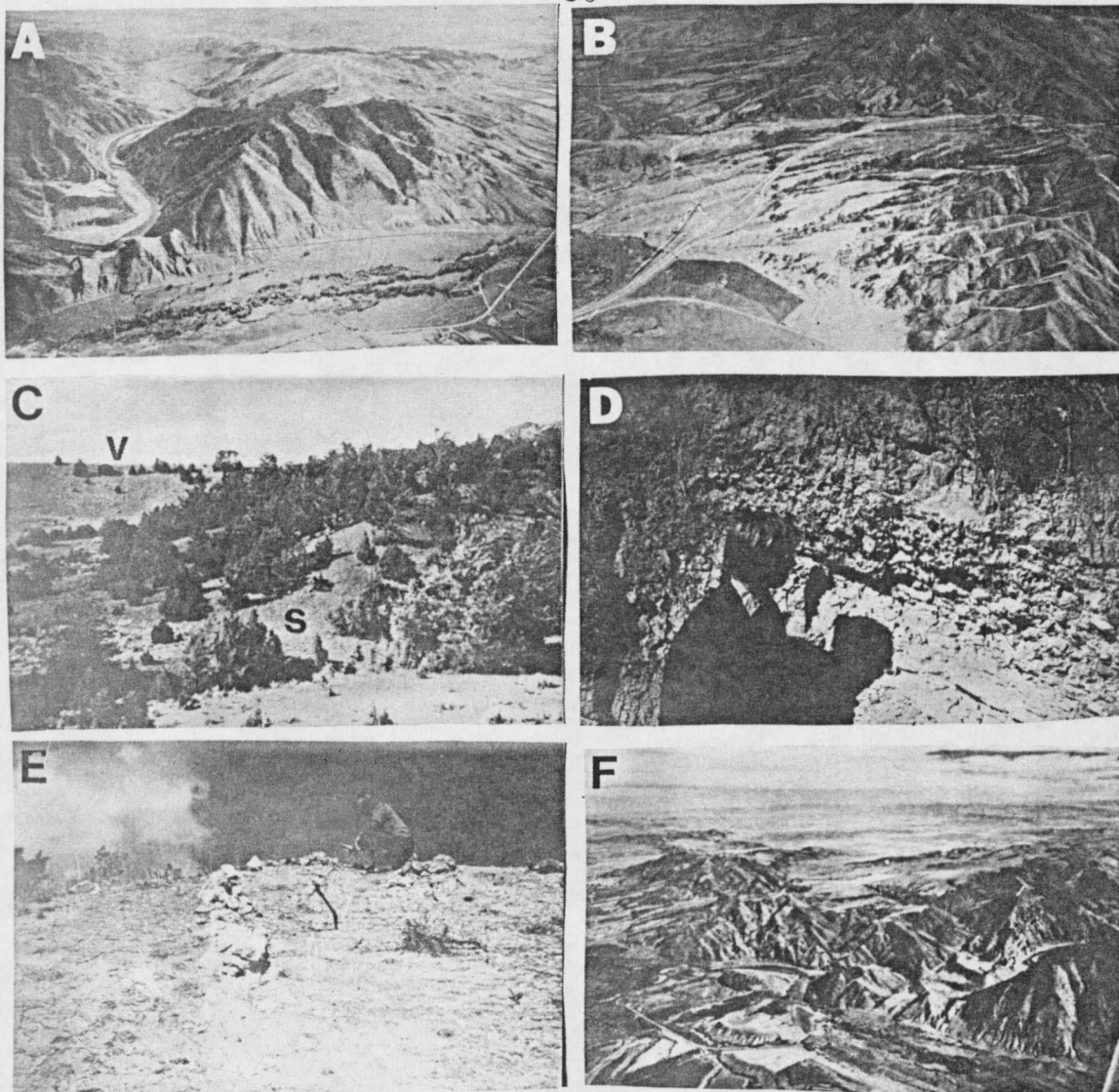


Fig. 12. A. View east at faulted front of London Hills, and meandering Jefferson Canyon. B. View north at fault scarp at Location I. Interstate 90 crosses scarp. C. Tree covered hill forming gentle fault scarp (s) at location I. Vendome pediment (v) caps the scarp. View north. D. Cobbles and sand at location II dipping east into London Hills front. E. Vertical color change marks vertical bedding of Six Mile Creek Formation at location III. F. View northeast at faulted front of London Hills. Abandoned Cenozoic valley which carried Ballard gravels pinches out near the fault. Cave Mountain and Jefferson Canyon are visible south of the valley.

in the upper exposures, and resemble deposits of the Six Mile Creek Formation (Fig. 12D).

Thus, structural movement has dropped the Bozeman materials on the west side of the Starretts Ditch fault at location II. The vertical offset is at least 1,000 feet here, and could possibly be considerably more, if the eastern block moved up as well. The faulting probably took place during or after the deposition of the Six Mile Creek Formation.

Location III

A 500 foot offset of the mid-Tertiary erosion surface had been reported between exposures that lie east and west of the crest of the London Hills near Negro Hollow (Kuenzi and Richard, 1966). They found no evidence of structural movement, and assumed that the difference was the result of deeper erosion on the west side of the so-called drainage divide that was proposed by Richard (1966). Kuenzi (1974) later disagreed with this mechanism, finding that the drainage divide did not exist, though not specifically explaining the reported difference of elevation of the mid-Bozeman erosion surface.

The author found evidence of late Tertiary structural movement nearby. Along the base of the cliffs of the Mission

Canyon Limestone, about one mile north of the Negro Hollow Road, a red-stained breccia zone records the fault movement. The breccia has no clayey matrix, nor partly-rounded limestone clasts, so does not resemble a solution breccia. Dozens of shear planes cut through the rock. The only slickensides that clearly showed relative movement dip steeply to the west.

Within several hundred feet of this broken limestone zone, the dip of the beds at the top of the Six Mile Creek Formation abruptly changes to vertical (Fig. 12E), whereas they dip gently towards the center of the basin for several thousand feet west of the fold. The shear fold is approximately 20 feet wide. Directly uphill from this fold, the Six Mile Creek Formation has been eroded away.

This evidence suggests that late Tertiary normal faulting uplifted the eastern block 500 feet relative to the western block, and fractured the Mission Canyon Limestone. The Six Mile Creek Formation was deformed into a broad drag fold by the movement, and has eroded back from the upthrown block.

Relation of Location II and Location III

At location II, the western block dropped down steeply, but at location III, it was gently deformed upward. The

London Hills front is relatively higher above the alluvial plains at location II than at location III, perhaps reflecting these two different kinds of offset. The faults have not been traced along the entire western front of the London Hills, but a tentative explanation can be offered. The fault in Negro Hollow is not physically continuous with the one adjacent to the Jefferson Canyon, but the two combined into a fault system, together uplifting the western flank of the London Hills area.

Summary of Starretts Ditch Fault

Evidence at three outcrops strongly suggest that the London Hills were uplifted along their western margin relative to the basins. Many other mountain ranges, such as the Tobacco Root, Madison, Bridger, and Beartooth Ranges appear to have been faulted on their western margins. The movement of the London Hills occurred during or probably after the deposition of the Six Mile Creek Formation, but before the cutting of the Vendome pediment in the late Quaternary. This late Tertiary age agrees with other known faults in the region (Pardee, 1950; Kuenzi and Fields, 1971; Peterson, 1974).

Objections raised by Parker (1961) and others that the Tertiary drainage could have flowed to the east are answered.

Bozeman streams flowed to the east, as proposed by Kuenzi (1974), but subsequent structural movement truncated the ancient, buried Tertiary valleys west of the Starretts Ditch fault. This caused the abrupt shallowing of the valleys to the east, as found by Parker (1961) and others.

This interpretation of the Tertiary history of the London Hills better explains the map made by Richard (1966) than did his drainage divide model (Fig. 11). If the London Hills had served as the continental divide, Tertiary drainages would have cut into the western flank of the Hills as they did on the eastern margin. No such drainages exist (Fig. 11). The late Tertiary uplift of the western margin of the London Hills produced the relatively straight boundary. Erosion cut past the bottom of the former Tertiary valleys on the upthrown block, causing them to shallow and pinch out to the west (Fig. 12F).

Ballard Gravels

Introduction

Deposits of an early Quaternary river, probably the ancestral Jefferson River, lie at an elevation that is 1500 feet above the modern Jefferson channel. The deposits are continuous for ten miles, and their composition points to a source area an additional seven miles to the west. For convenience of discussion, the author will informally call these deposits, the "Ballard gravels".

Lithostratigraphy

The Ballard gravels range in size from cobbles to gravel in most outcrops, although boulders and coarse sand also occur. The boulders are most common in the western exposures in sections 5 and 6, T. 1 N., R. 2 W., and section 32, T. 2 N., R. 2 W. (Fig. 13A,B). The coarse sand, including recycled shale clasts, is best exposed in sections 27 and 28, T. 2 N., R. 2 W.

The Ballard gravels are mostly composed of pieces of Elkhorn Mountain Volcanics and Paleozoic quartzites. Andesite porphyry is the most common volcanic material, but basalt and welded tuff are also abundant (Fig. 13C). Occa-

sional pieces of quartz monzonite are also found. Volcanic ash is clearly absent from the Ballard deposits.

Origin

The Ballard gravels were deposited by a large river. Abundant cross bedding, sorting, and rounding of the material prove its fluvial origin. The river carried large boulders that are over two feet in diameter a distance of at least seven miles from Bull Mountain, the source area for the volcanics (Plate I). The area around the Ballard exposures is relatively flat (Fig. 12F), so only a large, efficient river could move boulders across it.

During the deposition of the Ballard gravels, the surrounding Paleozoic rocks were mostly covered by Bozeman Group materials. Thus, pieces of shale of the Renova Formation have been reworked into the gravels, but pieces of the Paleozoic limestones are rare to absent in the Ballard gravels. A six foot-thick block of the shale was found in the gravels in section 27, T. 2 N., R. 2 W., and probably represents a large piece of bank material that was undercut by the river, and dropped into the younger gravels. The ends of this large mass of shale are broken and discontinuous (Fig. 13D), and many gravel-sized pieces of shale can be found above and

below the main mass of shale (Fig. 13E), testifying to the reworking of the older shale. More than 50 leaf fossils were recovered from the shale (Fig. 13F), and include eight genera identified by Dr. Charles N. Miller, Jr., and Ms. Robin Youse of the University of Montana. The flora closely matches the York Valley Ranch collection by Becker (1973) in the Dunbar Creek Member of the Renova Formation of the upper Ruby Valley. In addition, two of the genera, Cercidiphyllum and Zelkova, became extinct in North America before the Pliocene (Miller, 1978, written commun.).

Most of the former Tertiary Bozeman cover has been buried or eroded from the Ballard area. Angular limestone gravels of the Vendome pediment cap many of the remaining Renova outcrops (Reshkin, 1963). In addition, the presence of recycled shale in the Ballard gravels show that the Tertiary cover was once greater than it is today, because no source for the Dunbar Creek shale is known within ten miles of the outcrop containing the leaf fossils. Future detailed mapping of the Cenozoic units of the Jefferson Island quadrangle may locate some Dunbar Creek outcrops, but rapid reconnaissance mapping by the author failed to discover any. However, the Climbing Arrow Member of the Renova Formation directly underlies the Ballard gravels in SE $\frac{1}{4}$, NE $\frac{1}{4}$,

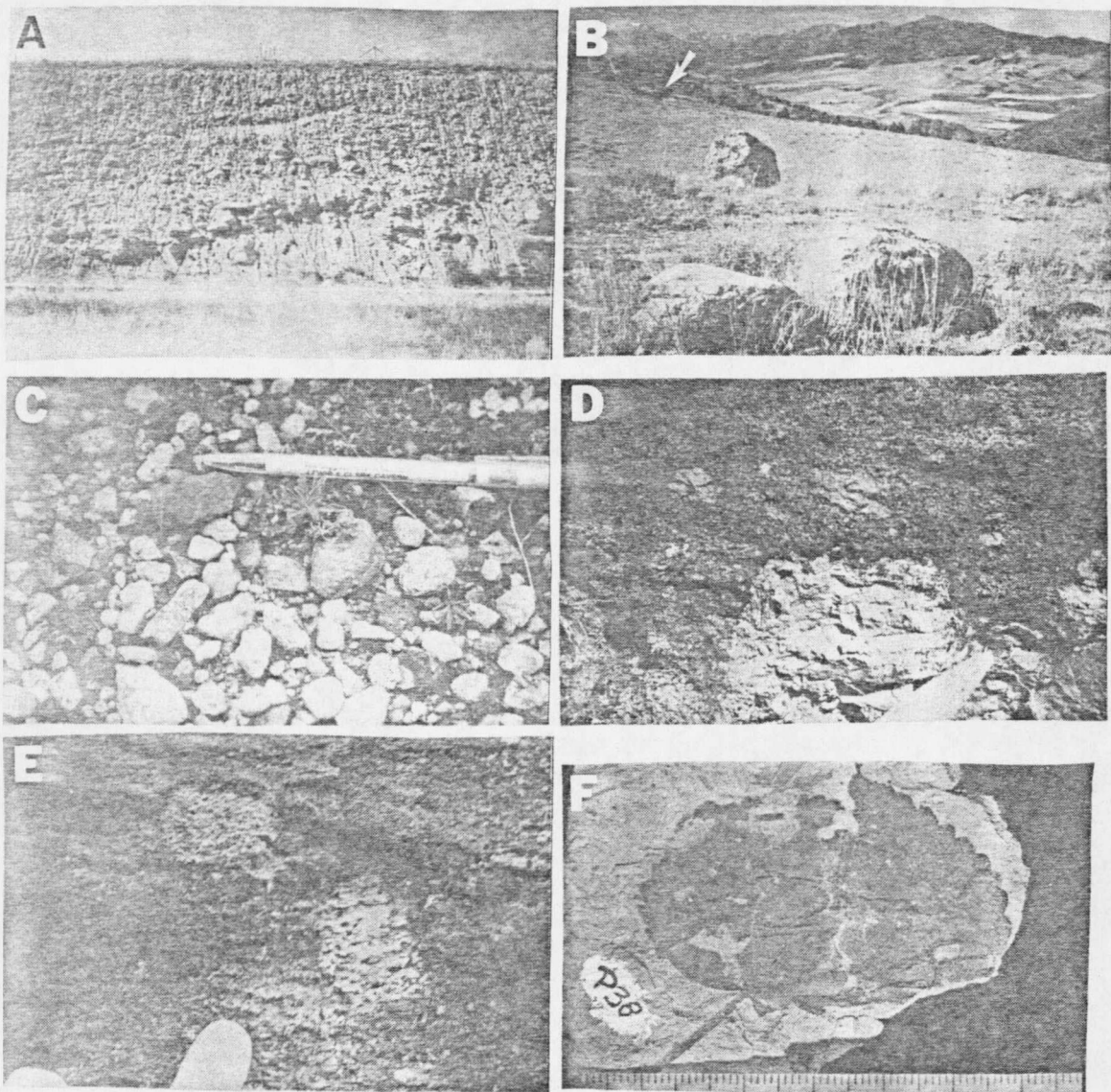


Fig. 13. A. Bedding in boulders of Ballard gravels dipping to east along Interstate 90. B. Looking west from boulder-strewn surface to the Jefferson Basin, 1500 feet below. Jefferson Canyon (arrow). C. Volcanic and quartzite pebbles of Ballard gravels. D. Ragged piece of recycled Renova shale in Ballard gravels (hammer for scale). E. Rounded gravel-sized pieces of Renova shale in Ballard gravels (finger for scale). F. *Cercidiphyllum* fossil from recycled shale.

NW $\frac{1}{4}$, sec. 26, T. 2 N., R. 2 W.

The base of the Ballard gravels is also exposed near their western end, where the gravels rest unconformably on Precambrian and Paleozoic rocks, in NW $\frac{1}{4}$, SE $\frac{1}{4}$, sec. 32, T. 2 N., R. 2 W. The downcutting of the river to these older formations on the western end, but to Oligocene rocks on the eastern end, shows that the Starretts Ditch fault uplifted the western edge of the London Hills before the deposition of the Ballard gravels (Fig. 15C).

The Ballard gravels were deposited during the early Quaternary, and possibly as early as latest Pliocene. The movement of the Starretts Ditch fault occurred after the deposition of the Six Mile Creek Formation, thus the Ballard gravels are post-middle Pliocene. The gravels are covered only by late Quaternary features, such as loess and the Vendome pediment. The gravels are not late Quaternary in age, because they lie at an elevation of 5752 feet in SE $\frac{1}{4}$ sec. 6, T. 2 N., R. 2 W., 1500 feet above the modern Jefferson channel. In addition, the gravels cross the modern drainage divide on the London Hills between tributaries of the Three Forks and Boulder Basins, and thus the gravels pre-date the modern topography.

Early workers assigned the Ballard gravels to the Boze-

man Group without offering any lithologic or paleontologic evidence (Berry, 1943; Perry, 1950; Lowell, 1957; Reshkin, 1963; Richard, 1966). Subsequent definition of the Bozeman Group in adjacent quadrangles (Robinson, 1963; Kuenzi and Fields, 1971) shows that the Ballard gravels do not belong to the Bozeman Group. The very coarse nature of the Ballard deposits is not compatible with the finer sediments of the Bozeman Group, especially the Renova Formation. In addition, typical Bozeman exposures are wide sheets (Plate I), while the Ballard gravels are markedly linear (Plate II). The complete absence of volcanic ash in the Ballard deposits clashes with either the Six Mile Creek or Renova Formation. Finally, the entire one mile exposure of the Ballard gravels in the Three Forks quadrangle was mapped by Robinson (1963) as older Quaternary alluvium, not as Bozeman deposits.

The Ballard gravels were probably deposited by the ancestral Jefferson River. The striking coarseness and great distribution of the Ballard deposits testifies to the large size of the former river. In addition, the author found no other deposits in the London Hills area that could have been from the pre-canyon Jefferson River. However, the absence of pieces of Precambrian gneiss, which are common in the modern Jefferson deposits, is problematic.

The river flowed to the east, similar to the modern Jefferson as well as the Tertiary drainages (Kuenzi, 1974). Reshkin (1963) proposed easterly flow based on the distribution of source rocks for the Ballard gravels. However, the author finds this evidence weak, because Elkhorn Mountain volcanics and Paleozoic quartzites crop out both east and west of the Ballards (Plate I). However, the base of the Ballard gravels is clearly lower on the eastern end than on the western end, indicating an easterly gradient.

The ancestral Jefferson River probably meandered across the low relief plain, switching its position several times. Although the best exposures can be found along Interstate 90 in Cottonwood Canyon and the Eureka Basin, the Ballard gravels also lie along the Negro Hollow Road (sec. 4, T. 3 N., R. 2 W.). Thus the river changed its position several times before cutting the Jefferson Canyon in its present position.

Summary of the Ballard Gravels

The Ballard gravels recorded the presence of a major river that flowed across the buried London Hills. The river post-dated the deposition of the Bozeman Group and movement of the Starretts Ditch fault, but pre-dated the cutting of the Jefferson River Canyon.

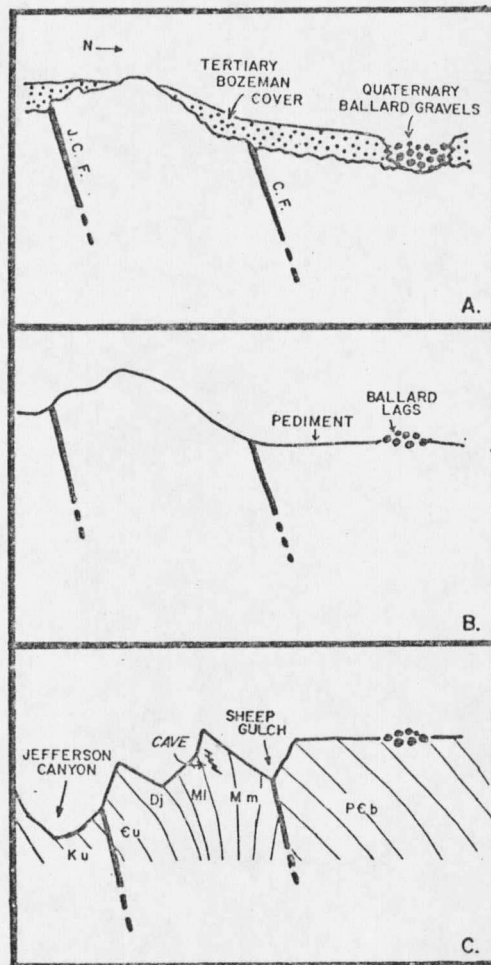


Fig. 14. A. Cross section of early Quaternary landscape, showing pre-Cenozoic rocks (white) mostly buried by Bozeman deposits. Ballard gravels cut through Bozeman cover. B. Erosion removes non-resistant Bozeman materials, porous Ballard gravels remain above the pediment level. C. Modern topography created by deep erosion of Jefferson Canyon and Sheep Gulch. Ballard gravels and pediment surface remaining above the level of the cave show that the cave site was below base level during earliest Quaternary, before the cutting of the canyon. Pre-Cenozoic rocks were tilted during the Laramide orogeny.

High Level Pediment

An early Quaternary pediment also affected the Ballard gravels. The surface cuts rocks from Precambrian to Mississippian, although most of its seven square miles is on the LaHood Formation. Its average elevation is 5,600 feet, and it has 300 feet of relief cut into it by later erosion (Reshkin, 1963). Reshkin (1963) believed that the pediment formed during the late Tertiary, because its elevation and morphology matches with other surfaces of that age. The author proposes that the surface is younger than the Ballard gravels (Fig. 14), so is probably younger than Reshkin's date. The coarse, porous Ballard deposits were more resistant than the surrounding Tertiary sediments, and so remained above the pediment surface. Modern drainages, such as Sheep and Greer Gulch cut the pediment, so the surface is probably older Quaternary.

Jefferson Canyon

Introduction

The Jefferson River Canyon was cut primarily by superposition, but antecedence played a role as well. The entrenchment occurred after the movement along the Starretts

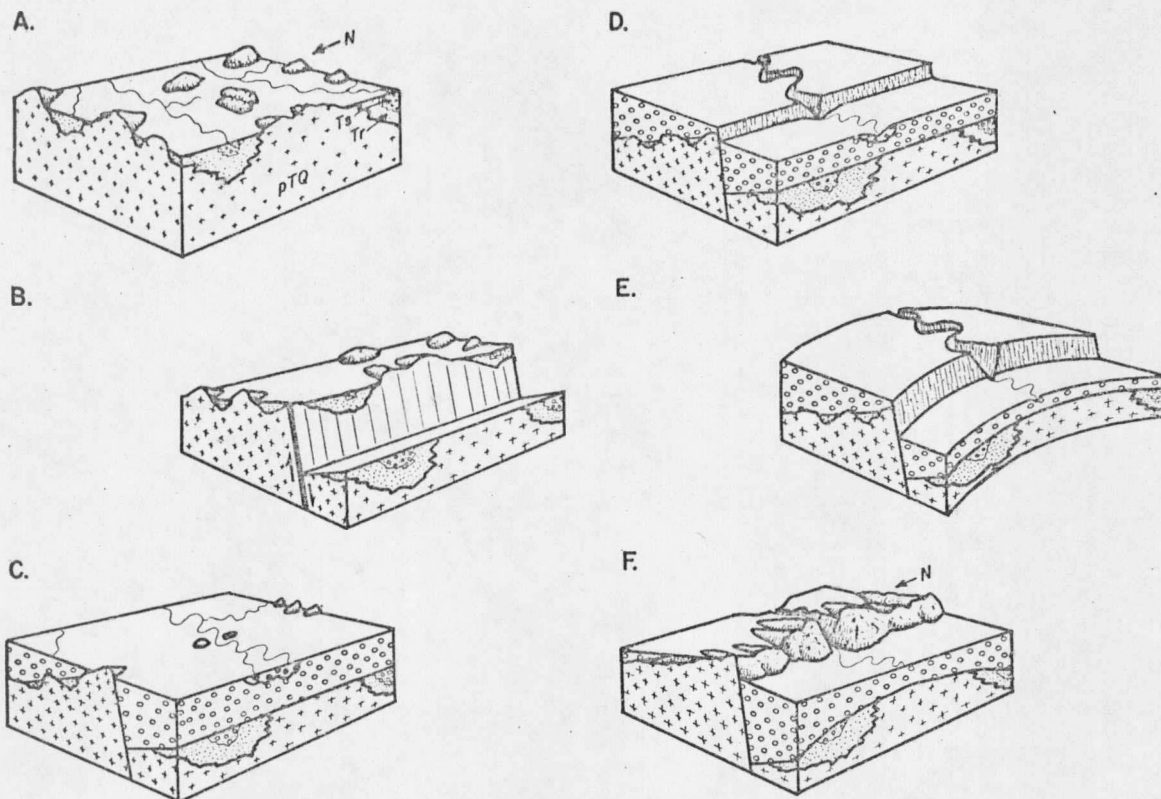


Fig. 15. Cenozoic geomorphic history of London Hills area. A. Deposition of Tertiary Bozeman Group. B. Post-Six Mile Creek movement of Starretts Ditch fault. C. Burial of fault by late Cenozoic sediments, including the Ballard gravels. D. Entrenchment of Jefferson River by superposition; fault scarp partly exposed. E. Arching causes further degradation of river. F. Modern topography of London Hills and Jefferson Canyon.

Ditch fault and deposition of the Ballard gravels, and was an early Quaternary event.

Description

The Jefferson Canyon is six miles long and 1400 feet deep. The meandering course of the canyon is slightly more

sinuous than most of the other canyons along the headwaters of the Missouri River system (Peterson, 1974). The canyon begins in the LaHood Formation near the Starretts Ditch fault, crosses into the upper Paleozoic rocks at the Cave fault, and traverses the Madison Group for most of its course (Fig. 16A). The canyon ends where the river crosses a Laramide fault zone, herein named the Managers fault (Plate II) (Fig. 16B), and follows the strike of the less-resistant Cretaceous units for three miles.

Two small canyons occur downstream from the main Jefferson Canyon. The river turns sharply to the south, cutting through a large hogback of Pennsylvanian and Mississippian rocks. The eastern end of this hogback is folded into the shape of the letter "J", and is called the "J Hill" by some local geologists. For convenience of discussion, the author will call this short canyon, the "J Canyon" (Fig. 16C). Four miles downstream from the J Canyon and two miles past the town of Sappington is another canyon, cut primarily through the Madison Group. The author will refer to this as the Sappington Canyon (Fig. 16D).

Method of Entrenchment

Headward erosion: Three different processes could have cut

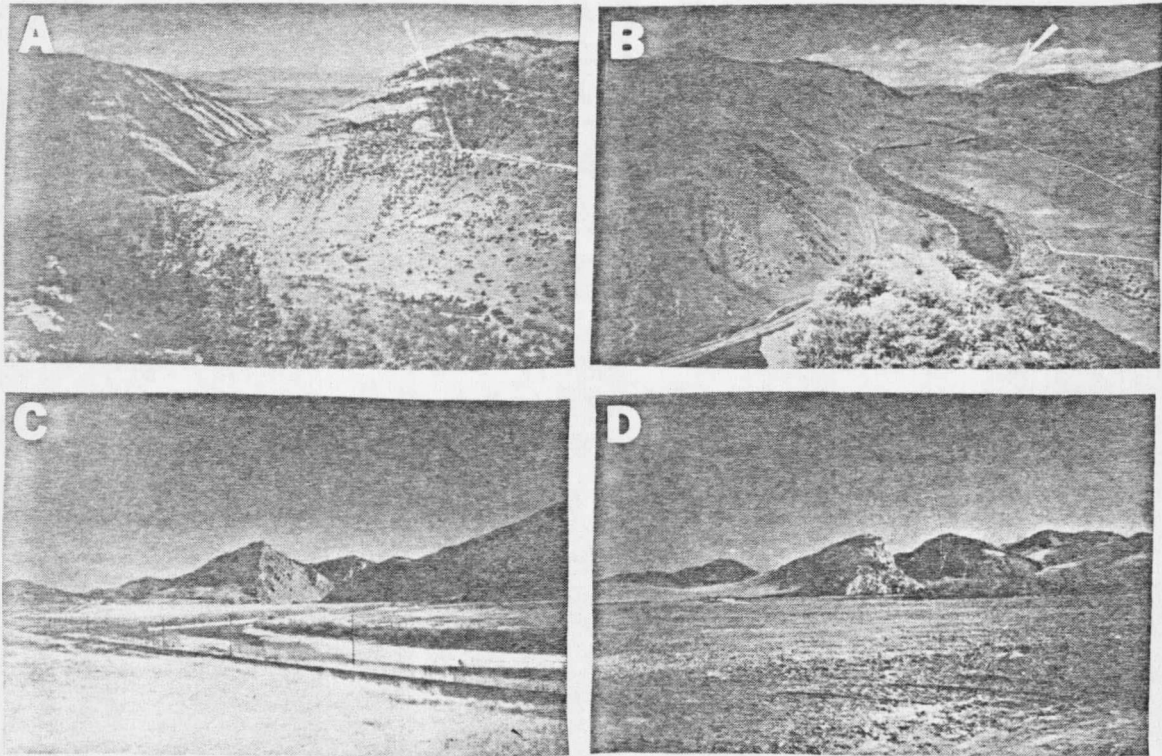


Fig. 16. A. Looking west into the Jefferson Canyon, cut primarily through the Mission Canyon Limestone. Caverns entrance (arrow) near top of vertical tram tracks north of canyon. B. Looking east along dipping Cretaceous units from the top of the J Canyon towards the Jefferson Canyon. Managers fault uplifted Madison rocks on south side of river at mouth of the canyon (background). Cave Mountain (arrow). C. The J Canyon, looking east. D. The Sappington Canyon, looking east. Trees mark the river position. Note the low, flat topography north of the canyon which was not exploited by the river.

the Jefferson Canyon and other canyons on the Missouri headwaters, headward erosion, superposition, and anteceden-

(Peterson, 1974). Headward erosion is unlikely. Tertiary deposits extend completely through the London Hills area in Negro Hollow, and around the southern edge of the Hills through Summit Valley. However, the river cut through the more resistant Mission Canyon Limestone and LaHood Formation, rather than these non-resistant units. In addition, the meandering path of the canyon suggests superposition or antecedence instead of headward erosion (Lowell, 1957; Peterson, 1974).

Superposition: Most authors favored superposition over antecedence for the Jefferson Canyon, but offered little evidence, (Lowell, 1957; Reshkin, 1963; Thornbury, 1965; Richard, 1966; Peterson, 1974). Fields (oral commun., 1979), however, cited the presence of late Tertiary sediments at an elevation of approximately 6,600 feet on the downthrown block, about five miles west of the Starretts Ditch fault. This shows that Cenozoic valley fill covered much of the uplifted London Hills area, allowing the superposition of the Jefferson River.

The abundance of other canyons in the area suggests regional rejuvenation, rather than many, localized uplifts. At least 17 major canyons have been cut by the headwater

ivers of the Missouri River (Fig. 8) (Peterson, 1974).

Smaller canyons, such as the Sappington Canyon and J Canyon, show additional evidence for superposition. Both of these canyons formed without any apparent local, late Cenozoic movement. The upstream end of the J Canyon is the resistant dip slope of the Quadrant Formation. Downstream, the river enters the old Tertiary valley of Antelope Creek. The Sappington Canyon cuts through an erosional remnant of the resistant, vertically bedded Madison Group.

The Ballard gravels also support superposition. The ancestral Jefferson River flowed across the buried Starretts Ditch fault, depositing the Ballard gravels. The gravels rest unconformably on Precambrian and Paleozoic rocks near the fault, but overlies Tertiary Renova deposits five miles west of the fault. These relationships show that the Starretts Ditch fault had already moved, exposing the older rocks near the fault, before the Ballard gravels were deposited and before the Jefferson Canyon was cut (Fig. 15C).

Antecedence: Late Quaternary arching of the canyon area aided in the entrenchment of the Jefferson River Canyon (Fig. 15E). Older pediments along the North Boulder Valley recorded tilting away from the canyon (Richard, 1966).

The older, higher pediments have a component of dip to the north, opposite to the grade of the modern river, especially when viewed from the east. Younger pediments slope less to the north, and the youngest surfaces follow the southerly grade of the river. In addition, Richard (1966) found that tributaries of the North Boulder River and Jefferson River are slightly barbed away from the canyon.

The distribution of the Ballard gravels also proves antecedent arching. Two miles north of the Jefferson Canyon, the base of the Ballard gravels is found at an elevation of 5,700 feet, but the gravels are at least 800 feet lower eight miles north of the Canyon in Negro Hollow. Both elevations of the gravels were found very close to the Starretts Ditch fault.

Thus the canyon area was uplifted relative to areas north and south of it by at least 800 feet during the late Quaternary, and possibly early Quaternary. The arching affected locations both east and west of the Starretts Ditch fault, such as the Ballard gravels and the pediments on the west side of the North Boulder Valley, so was not directly related to that fault.

Summary of Method of Entrenchment: The Jefferson Canyon was cut by both superposition and antecedence. The antecedent structure was only local, and did not affect the 16 other major canyons on the Missouri headwaters. Superposition was probably more important than antecedence for the Jefferson Canyon, because the other major canyons were cut with no apparent antecedent control.

Structural Control of the
Site of the Jefferson Canyon

The position of the superposed Jefferson Canyon was not completely arbitrary, but was influenced by weaknesses along the ancient Willow Creek fault. The anomalous east-west trend of adjacent basins, the Three Forks Basin and the north-eastern portion of the Jefferson Basin (Fig. 8), was caused by the Jefferson Canyon fault and buried Willow Creek fault (Robinson, 1963; Kuenzi and Fields, 1971). Precambrian movement of the Willow Creek fault weakened the rocks of the area, and the zone was reactivated along the Jefferson fault (Kuenzi and Fields, 1971), as well as possibly during the Laramide. The Jefferson Canyon is between and in line with these two east-west basins (Plate I). An additional fault

may coincide with the position of the river in the Jefferson Canyon (Schmidt, oral commun. 1977). Some features, such as the small anticline which is cored by the LaHood Formation in sec. 20, T. 1 N., R. 2 W., do not match well across the river (Plate II). The author agrees that a fault may be present, but was unable to prove its existence.

The possible presence of a fault does not prove headward erosion, but only shows some slight structural control of the location of the canyon. The ancestral Jefferson River flowed along several paths, evidenced by the presence of the Ballard gravels in Cottonwood Canyon and Negro Hollow, but the river finally settled in its present position, where the fractured rock was slightly less resistant. The canyon was then cut by superposition.

Age of the Canyon

The author found no deposits of the Tertiary Bozeman Group along the course of the modern Jefferson River from the Starretts Ditch fault to Antelope Creek, although several earlier authors mistakenly mapped Quaternary pediment gravels as Bozeman materials (Parker, 1961; Richard, 1966). If the canyon formed during Bozeman times, Renova or Six Mile deposits would probably be preserved. Their absence suggests

that the canyon formed during the late Cenozoic. Most authors agreed with this evidence and date (Lowell, 1957; Reshkin, 1963; Richard, 1966; Peterson, 1974).

The Ballard gravels were deposited at their high elevation before the canyon was cut (Fig. 14). If the canyon had existed during the deposition of the Ballard gravels, the high level river would not have flowed parallel to the canyon for a distance of eight miles, but would have cut a direct path to the canyon, the local base level. The age of the gravels is early Quaternary, thus the canyon is a Quaternary feature. This fits within the age date of Peterson (1974), who extrapolated the maximum age of 4.2 million years of the Sweetwater Canyon to all other canyons on the Missouri headwaters. Late Quaternary features, such as the Vendome pediment and deposits of Recent volcanic ash occur less than 100 feet above the modern drainage. Thus the 1400 foot deep Jefferson Canyon was cut during the early Quaternary.

Summary of London Hills History

The Willow Creek fault formed a mountain range along the southern shore of the Belt embayment during late Precambrian times. Most Paleozoic and Mesozoic rocks, such as the

Mission Canyon Limestone, do not record any effects of the fault, so it was probably mostly inactive during these eras. The weakness was reactivated during the Laramide orogeny, however, influencing the Cave fault and Jefferson Canyon fault. Just before these two faults moved, shear folding deformed the strata of Cave Mountain into a plunging anticline and syncline.

Great thicknesses of sediments accumulated in the basins of southwestern Montana after the Laramide orogeny. Fine-grained sediments were deposited, later partly eroded, and finally partly covered by coarse-grained sediments. The London Hills area was relatively low during the deposition of these Bozeman Group sediments, allowing easterly-flowing streams to bury much of the area.

Shortly after or near the end of the deposition of the Bozeman Group, the Starretts Ditch fault uplifted the western edge of the London Hills area. Sediments accumulated in the dammed drainages west of the fault, eventually filling to the top of the London Hills and continuing the flow to the east. The Ballard gravels were deposited by the ancestral Jefferson River across the London Hills during the early Quaternary. After changing its position several times, the river downcut in its present position, powered by regional rejuvenation and local arching of the canyon area.

CHAPTER 4

SPELEOGENESIS

Introduction

Mankind has shown interest in caves throughout history, from the so-called "cave man" to Tom Sawyer. Scientific concern, however, had a later start, beginning in the 1700's with Nagel (Sweeting, 1973). The outstanding karst area of the Slavic countries naturally sparked European studies, thus intense work began in the late 1800's with Cvijic, Martel, and others (Jennings, 1971). Research in the United States has accelerated in the past 20 years, but probably still lags behind European accomplishments.

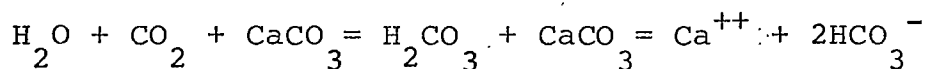
The field of speleology, the science of caves, involves many disciplines: geology, botany, anthropology, and others (Ford and Cullingford, 1976). This paper is concerned only with speleogenesis, the geologic origin of caves. This also is a multi-faceted field, including hydrology, structural geology, petrology, and geomorphology.

Chemical controls

Importance of Carbon Dioxide

Calcite, the major mineral of limestone, dissolves

readily in acidic water. Although sulfuric acid plays a role in some karst areas (Jagnow, 1978), carbonic acid is the most common acid in the solution of limestone caves (Moore and Nicholas, 1964). The atmosphere and soil furnish rain and groundwater with carbon dioxide, creating carbonic acid. The acid breaks the solid calcium carbonate into aqueous ions.



Addition of carbon dioxide drives the equation to the right, and more calcite is dissolved. Loss of carbon dioxide leads to the precipitation of calcite from the solution.

At least two factors encourage a high percentage of carbon dioxide in groundwater, low temperature and high organic activity in the soil. Cold water can hold more carbon dioxide in solution than can warm water (Thrailkill, 1968), illustrated by the rapid loss of carbonation from warm root beer. However, this effect is quite small in comparison to the partial pressure of carbon dioxide in the air in contact with the karst water (Jennings, 1971). The amount of carbon dioxide in the free atmosphere is commonly about 0.03%, but usually is about 1-2% in the air spaces of soil which contain plant materials (Jennings, 1971).

After collecting carbon dioxide from the atmosphere and soil, the water can dissolve limestone.

The groundwater normally reaches saturation with respect to calcite before leaving the limestone, unless the path through the limestone is short of the water moves rapidly through an efficient conduit. Thus solution is very slow during the early stages of cave development, because the groundwater can only move slowly through the small cracks, the the rate of solution increases as the passages enlarge.

Karst groundwater may deposit calcite if it enters an air-filled cave. Without significant plant activity, the cave atmosphere has a low level of carbon dioxide, thus the groundwater is not in equilibrium with it. The water must release some carbon dioxide to the cave atmosphere, and so loses some of its acidity, and deposits calcite. This process may increase the amount of carbon dioxide in a cave, and cause the incoming water to remain acidic, thus preventing further deposition of calcite. Consequently, these secondary mineral deposits, called speleothems, are common near cave entrances, but are rare deep inside of the cave, where the air is trapped (Moore and Nicholas, 1964).

Mixing Corrosion

The solutional aggressiveness of a saturated karst water can be restored or increased by the process of mixing corrosion, commonly called mischungskorrosion in the literature. A mixture of two or more solutions that have different temperatures or concentrations, can have greater dissolving power than either of its components (Picknett and others, 1976, in Ford and Cullingford). On a phase diagram, the field boundary between undersaturated and supersaturated conditions is a curved line, concave towards the field that is undersaturated with respect to calcite (Fig. 17). Any mixture of different solutions falls on a straight line between them, thus moving the product into a less saturated condition.

Simple cave passages may be filled with only a single body of groundwater, thus the mixing effect is most important in compound caves where passages join together, mixing different bodies of water. This is often important at the boundary between the phreatic and vadose zones, where vadose passages feed many different solutions into the phreatic zone (Thrailkill, 1968). Within the phreatic zone, large rooms may form where mixing takes place at the junction of several phreatic passages.

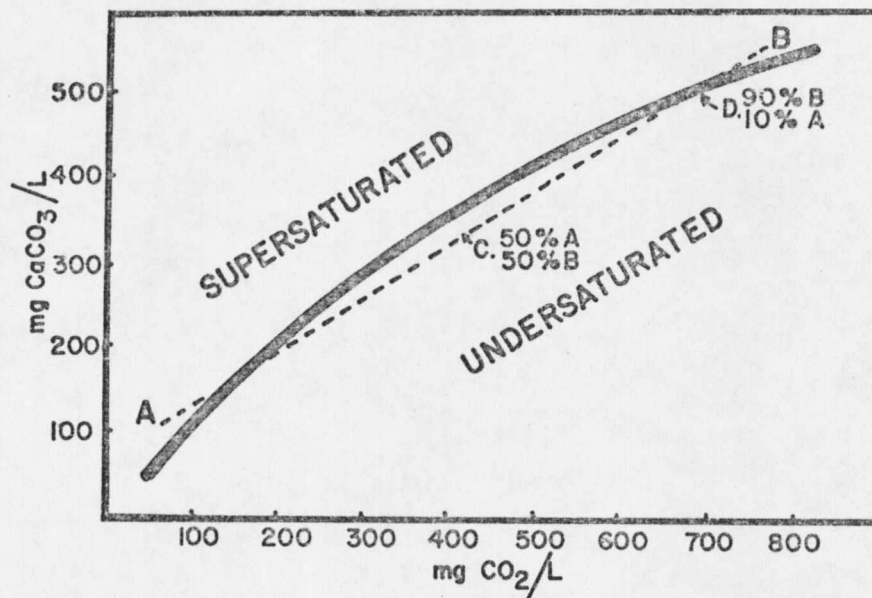


Fig. 17. Diagram explaining mixing corrosion. The solutional aggressiveness of one or both components of a mixture can be increased by mixing. Solutions A and B both fall in the supersaturated field, but a mixture of the two can be made less saturated by mixing 1 part A with 9 parts B (point D). The mixture may reach into the undersaturated field with a 1:1 mixture of A and B (point C). (After Picknett and others, 1976).

Hydrologic Controls

Groundwater Zones

Introduction: There are three zones in groundwater that are important in cave solution: the vadose zone, the phreatic zone, and the water table. Vadose water only partly fills

pores or caves in the rock, thus air occupies part of the vadose zone. In the phreatic zone, water completely fills all of the spaces. The flow of water is powered by the weight of the overlying water, called the hydrostatic pressure. The boundary between the vadose and phreatic zones is called the water table.

Each of these three zones have been cited as the most important in influencing cave development (Fig. 18). Early workers reasoned that the greatest solution takes place in the vadose zone, where the water is most acidic (Malott, 1937). Others thought that groundwater circulates to great depths, forming caves deep in the phreatic zone (Davis, 1930; Bretz, 1942). A third group proposed that caves form along the water table, because it was the most direct route through the rock from the recharge to discharge point (Swinerton, 1932).

Vadose and Phreatic Features: Bretz (1942) presented field evidence that cave solution occurs in both the vadose and phreatic zones. Phreatic and vadose caves can be distinguished by such things as the shape of cave passages and the pattern of the cave systems (Fig. 19).

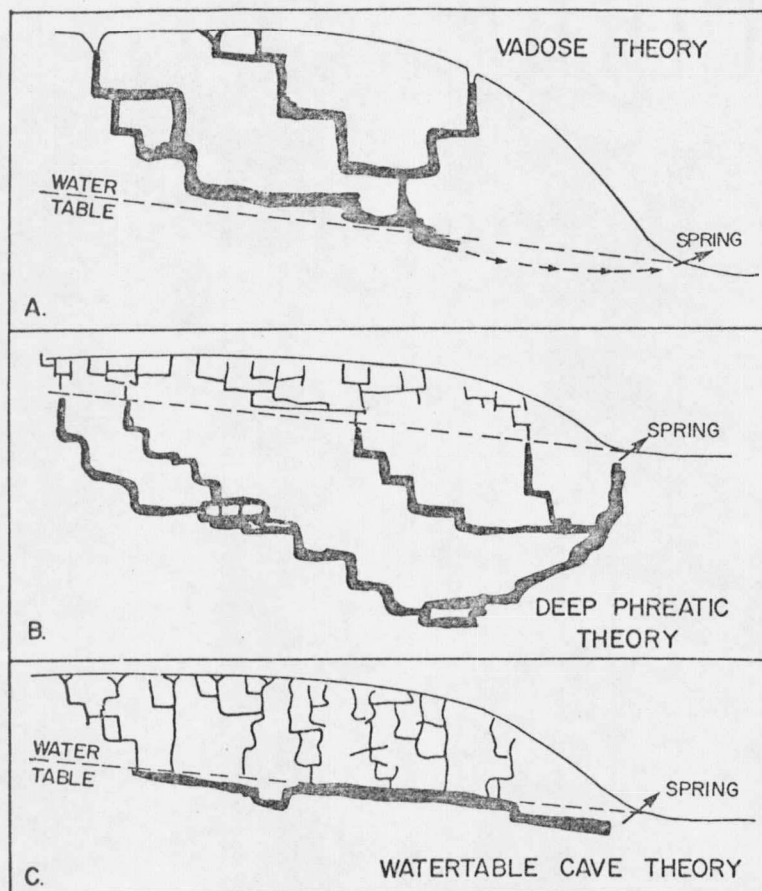


Fig. 18. Longitudinal section of sample caves that formed in each of the three groundwater zones. (From Ford and Ewers, 1978).

Vadose water, by definition, does not fill its conduit completely, thus erosion takes place only on lower walls and by a vadose stream (Fig. 19A). In addition, vadose waterfalls rapidly enlarge their path into a large shaft, called a domepit. Vadose water flows downhill, following the steepest possible path.

Phreatic water fills its conduit completely and is under hydrostatic pressure. Thus water can be forced to exploit every possible route, uphill as well as downhill.

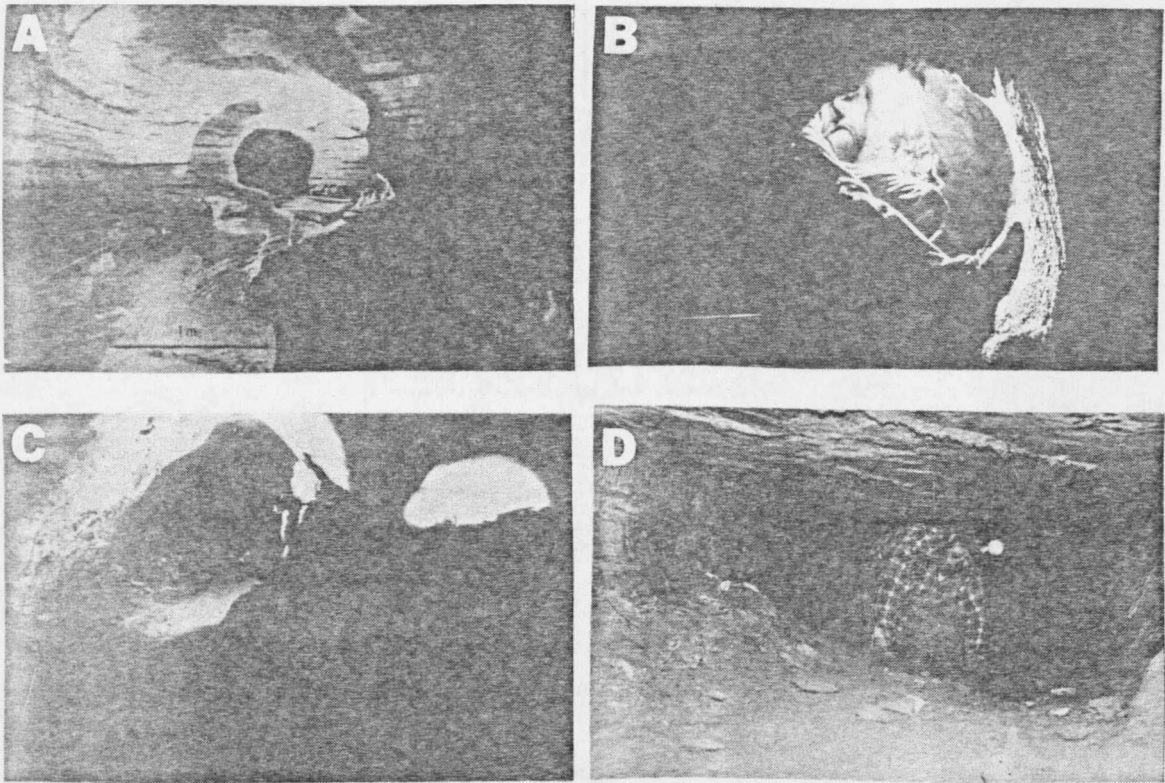


Fig. 19. A. Meandering vadose channel cut into a former phreatic, elliptical passage in horizontal limestone. (From Ewers, 1971). B. Looking upward into six-foot deep ceiling tube. Fish Tunnel, Lewis and Clark Caverns. C. Bedrock span dividing phreatic passage into two separate sections. Fish Tunnel, Lewis and Clark Caverns. D. Bedding plane ceiling in horizontal limestone (Sullivan's Cave, Indiana. Photo by author).

Small tubes may grow into the walls or ceilings (Fig. 19B). The flow may split, leaving a bedrock partition between two halves of the flow (Fig. 19C). Many interconnecting side channels may form, creating a maze. Phreatic passages often have an elliptical cross section, because solution attacks both the ceiling and the floor (Fig. 19A).

Water table caves are also called shallow phreatic caves, because phreatic features can be found in them. No sharp dividing line can be drawn between the water table and deep phreatic caves. In general, water table caves maintain a fairly constant and flat gradient, while deep phreatic caves consist of deep loops, so the water flows both up and down. These different caves are recognized in longitudinal cross section (Fig. 18).

Vadose caves may temporarily assume phreatic conditions, thus mixing the features described above. Rapid snowmelt or high rainfall may deliver more water to the vadose conduit than it can discharge. When this happens, the cave water fills to the ceiling, and may build a significant hydrostatic head. Elliptical cross sections and maze patterns may form in vadose caves by this process (Palmer, 1975). In addition, aggradation of sediments may stop the downcutting of a vadose stream and actually raise its bed, allowing the stream

to cut into the ceiling. However, this will likely produce a narrow, meandering channel in the ceiling, rather than eroding the entire ceiling as phreatic water does (Ewers, 1972).

Causes of Vadose Caves: Four factors encourage a significant amount of solution in the vadose zone: high permeability of the limestone, a mature phreatic cave system, high topographic relief, and a large surface stream feeding the karst conduits. The first three conditions encourage thick vadose zone by lowering or flattening the water table. The permeability can be increased by a large amount of solution in the phreatic zone, and by rapid enlargement of vadose passages due to high topographic relief. Ford (1971) suggested that a 12:1 ratio of horizontal to vertical distance from recharge to discharge is usually enough to promote significant vadose action. The lowering of the water table may drain many of the phreatic routes, exposing them to vadose waters. Former phreatic passages that have been modified by vadose streams are called drawdown caves (Ford and Ewers, 1978).

The formation of vadose caves is also favored by a large surface stream flowing into the limestone. The large

volume of acidic water and its rapid movement can rapidly dissolve cave conduits. Thus, a surface stream, fed by a large catchment area, that enters the limestone through a large, efficient opening, such as a sinkhole, may enlarge a drawdown cave or perhaps initiate its own path through the limestone, forming an invasion cave (Ford and Ewers, 1978).

Causes of Deep and Shallow Phreatic Solution: The density of joints in a limestone determines where the greatest flow of groundwater will be. Most of the water can follow the most direct path through the limestone, the water table, if there is a large number of fractures crossing the bedding plane. On the other hand, a small amount of joints gives the water few options, thus it follows an indirect, deep phreatic route. Thus the fissure frequency determines the depth within the phreatic zone that groundwater follows (Ford, 1971).

The relation of fissure frequency to the depth of a phreatic route was partly uncovered by Thrailkill (1968), in his investigation of the apparent water table control of Mammoth Cave in Kentucky. He analyzed the distribution of water flow through pipe networks of different geometries,

comparing the deep and shallow routes. A 1:1 ratio of horizontal paths to vertical paths, analogous to bedding planes and joints, showed a marked advantage for the water table route (Fig. 20A). However, he believed that a different ratio, 100:1 was more realistic for the poorly jointed limestone of the Mammoth Cave region. This ratio produced only an insignificant difference in the amount of flow between the shallow and deep routes (Fig. 20B). Thus Thrailkill looked to mixing corrosion and other processes to explain the water table control of Mammoth Cave.

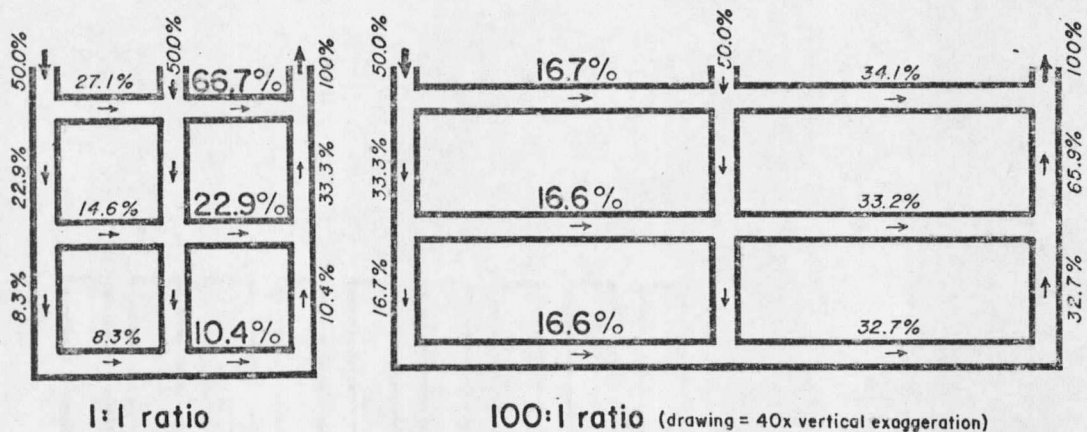


Fig. 20. Flow analysis through pipe networks. Shallow route receives relatively high percentage of flow with horizontal:vertical ratio of 1:1, but has only a negligible advantage with a 100:1 ratio. (After Thrailkill, 1968).

A four state model has been used to compare the deep and shallow phreatic routes (Ford, 1971). State 1 has the

minimum number of joints that will allow flow through the rock from sink to spring. This corresponds to Thrailkill's 100:1 ratio, and produces deep phreatic loops along the joints and bedding planes (Fig. 21). State 4 has a high fissure frequency, so that groundwater can choose the most direct path from the many available routes. The resulting state 4 caves maintain a low and fairly constant gradient. State 2 and 3 caves are transitional forms between the two end members, states 1 and 4 (Fig. 21).

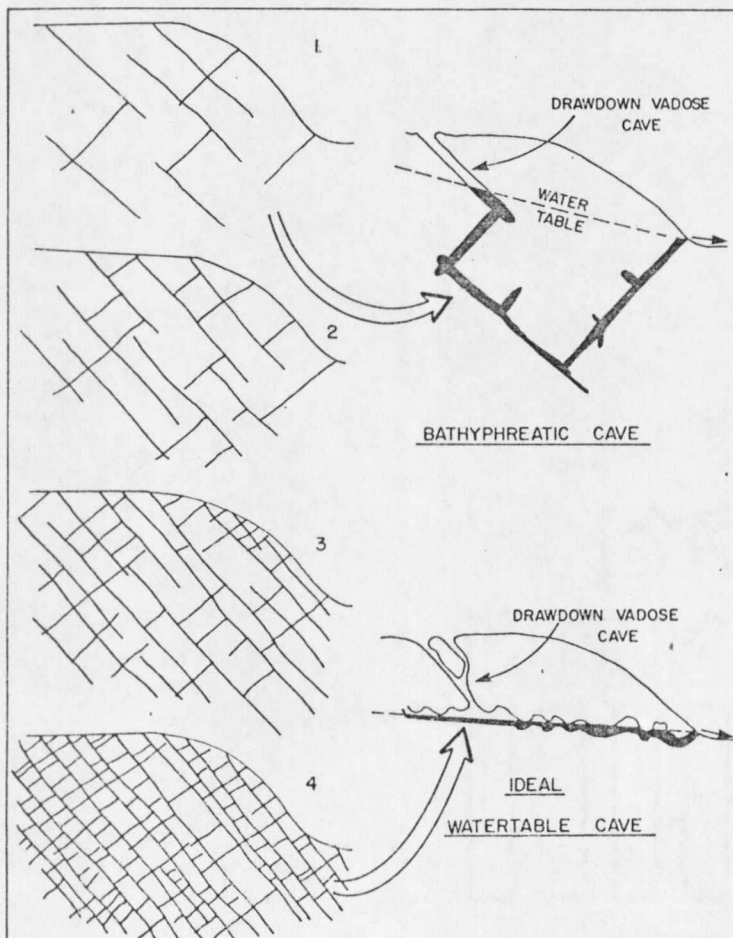


Fig. 21. Longitudinal sections of state 1 and state 4 caves and relation of fissure frequency to the cave geometry. (From Ford and Ewers, 1978).

Actual field data for this model is scarce. State 4 caves may require penetrable beds spaced no more than 35 feet apart, and joints within 50 feet of each other (Ford and Ewers, 1978). The gradient of a state 4 cave is very low, perhaps less than 1° (Davies, 1960), while state 1 caves have much steeper gradients. The thickness of the solution zone of a water table cave is approximately 50-60 feet (White, 1960). Phreatic loops of state 1 range up to 300-400 feet in amplitude (Ford, 1971).

Patterns of Cave Systems

Introduction: An integrated system of cave passages can form in two basic patterns, a branchwork or a maze (Palmer, 1975) (Fig. 22). Most of the passages of a branchwork system intersect in the downstream direction, forming tributaries to one or several large, trunk or primary passages. Passages of a maze cave intersect in many directions, forming a large number of closed loops (Fig. 22).

Experimental models (Ewers, 1972) as well as study of actual cave patterns (Palmer, 1975) show that the branchwork pattern is the most common cave form. Maze patterns require special conditions to form.

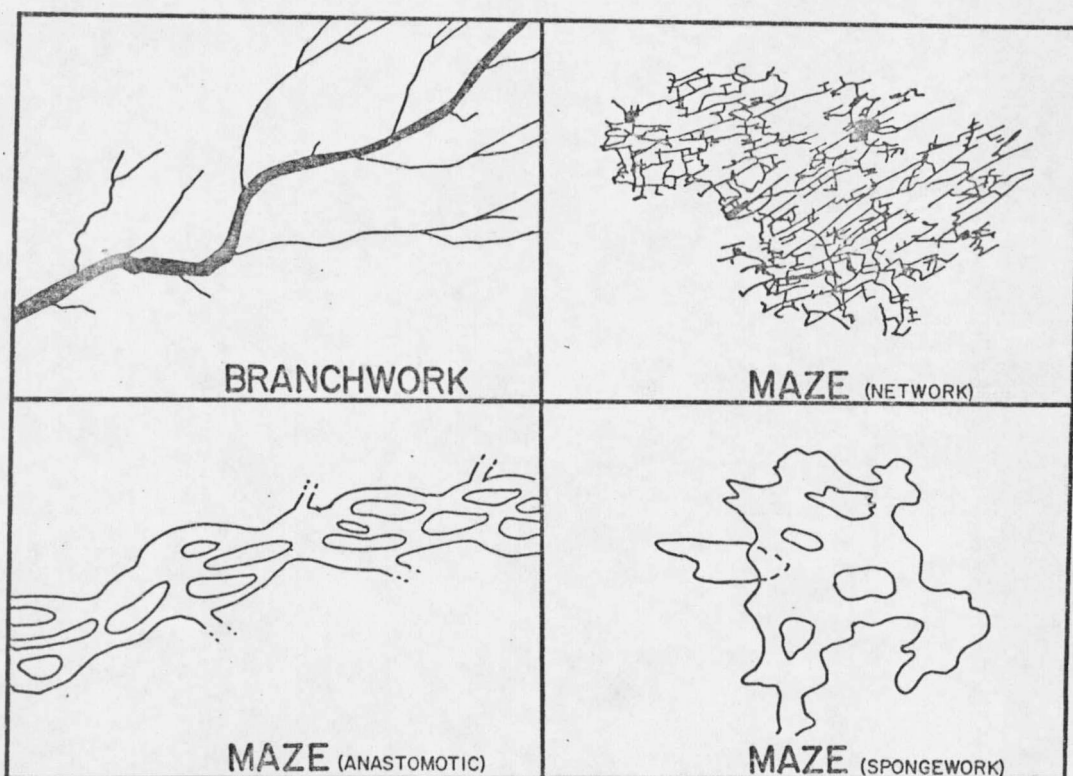


Fig. 22. Representative plan views of branchwork and maze caves. Network mazes consist of a regular, angular grid of passages. Anastomosing mazes are many meandering passages. Spongework caves are highly irregular. (After Palmer, 1975).

Branchwork patterns: The tendency for branchwork patterns to form can be shown in salt-block models (Ewers, 1972). Water flows through tiny, planar cracks in the soluble salt, which allows detailed study of the evolution of the solution conduits. Although many solution channels begin to radiate from the recharge points, one primary tube soon dominates from each recharge point (Fig. 23.2). Each of the primary

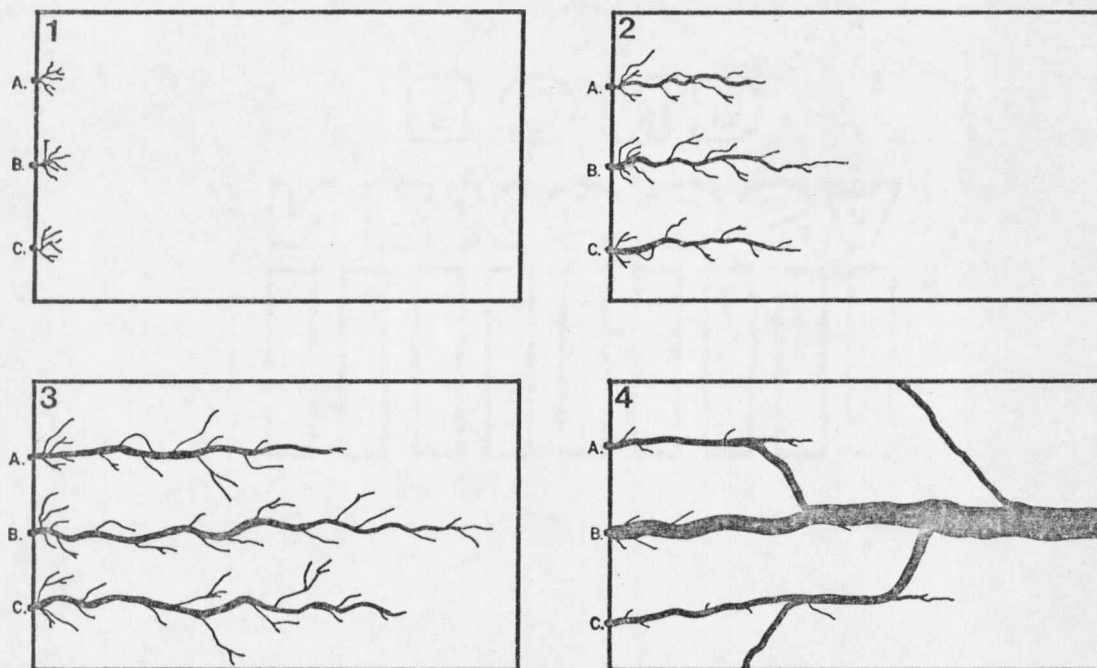


Fig. 23. Plan views of evolving cave passages. 1. Initial passages form in random directions. 2. Major passages soon develop in the direction of steepest hydraulic gradient. Minor random passages also grow. 3. Continued growth of passages towards recharge point. 4. First passage to reach discharge quickly enlarges, dropping the hydrostatic head. Other passages are pirated from original flow towards the evolving trunk passage. (After salt block models of Ewers, 1972).

tubes grows towards its discharge point, along the path of the steepest hydrologic gradient. Thus the pattern of the primitive cave passages consists of many parallel conduits forming in the direction of the steepest gradient.

The pattern changes once one of the solutional conduits grows all of the way to its discharge point. The integrated passage can efficiently carry water through the system, and so it reduces the hydrostatic pressure of the system. The

other primary tubes, which have not connected to the discharge point, are no longer powered by a large hydrologic gradient, and so abandon their path towards their original discharge points. They instead cut a path to the integrated primary conduit (Fig. 23.4). This forms a branchwork pattern of a large, trunk passage, fed by smaller, tributary passages.

Several factors encourage the trunk passage to enlarge more rapidly than the tributary passages. Early in the solutional history, the trunk will be the first to enlarge sufficiently to allow turbulent flow. Turbulent flow can erode the limestone four times faster than laminar flow (Palmer, 1969), and begins once the conduit has enlarged to a diameter of two to five inches (Ford and Ewers, 1978). In addition, the large diameter of the trunk passage allows efficient water flow, so only a small hydrostatic head is required to move water through (Palmer, written commun., 1978). A large hydrostatic pressure is needed to carry water through the smaller, tributary passages, thus they carry less water flow and enlarge still slower than the trunk passage. Finally, the combined flow from the many tributaries through the trunk enlarges it further. Thus most cave systems form a branchwork pattern, consisting of a large, domi-

nant passage, fed by secondary passages.

Maze Patterns: In a maze pattern, the tendency to form a trunk passage is suppressed, and many interconnecting passages carry the flow. This may occur if the water that enters the limestone is scattered over a very large number of recharge points, rather than a smaller number of points (Palmer, 1975). Many maze caves in the Midwest formed where sandstone, which scatters the water, caps the limestone. Where the sandstone cover is thin or absent, sinkholes form distinct input points, and branchwork caves occur (Palmer, 1975). This dispersal also occurs in areas of high relief, where the limestone is exposed as a resistant peak. The convex shape of the peak spreads the water much more than would a flat plain with sinkholes. Thus, maze patterns often form in high, limestone peaks (Palmer, 1975).

Maze patterns also form by floodwater recharge. Caves may receive more water from major rainstorms than they can handle, thus the water accumulates and builds up hydrostatic pressure (Palmer, 1975). The resulting insufficient permeability of the major passages allows the smaller, less efficient routes to be exploited and enlarged. Constrictions in the cave conduits, such as damming by sediments or less-

soluble zones in the rock, can cause or increase this effect.

Conclusion: Up to 75% of known caves formed in a branch-work pattern (Palmer, 1975). Laboratory experiments confirm that this pattern should be most common (Ewers, 1972). Maze patterns form if many recharge points feed the cave system, or if a high hydrostatic pressure is present (Palmer, 1975).

Lithologic Controls

Three factors of the petrology of limestone affect cave formation; the amount of insoluble residue, bed thickness, and primary and secondary porosity and permeability. In general, caves form in relatively pure limestones. Insoluble materials such as silica and clay retard both solution and water flow by choking the cave passage.

Beds of insoluble materials may concentrate water flow, thus causing the local formation of a cave passage. The fine muds may stop the downward flow of water, and concentrate the water in the carbonates directly above the mud layer. In addition, shale is physically eroded faster than limestone. The removal of shale as clasts may quickly open up subsurface routes along the bedding plane (Sweeting, 1973).

These insoluble layers cause cave formation in thin-

bedded limestones (Thornbury, 1969). However, the many thin beds allow many routes for solution, thus preventing concentrated flow along one or several beds that would enlarge easily into a cave. Thick bedded, pure limestones would have fewer favorable zones, thus concentrating the flow (Ford, 1976). In addition, thick beds are stronger than thin beds, so that caves can continue to grow, rather than be choked off by collapse. Hence, massive limestones should yield more caves than thin-bedded ones. Few caves in Montana formed in the thin-bedded Lodgepole Limestone, but numerous caves occur in the massive Mission Canyon Limestone (Campbell, 1978).

Karst limestones generally have low primary, but high secondary porosity and permeability (Thornbury, 1969). Water flow through rocks with large and interconnected pores, such as a Cenozoic limestone, would not be concentrated, and thus would probably not favor cave development. However, water can only enter a tight, massive limestone through cracks in the rock. These secondary features concentrate the flow, thus can enlarge into cave conduits.

Structural Control

Structural Elements

A cave is generally formed along several structural elements (Ford and Ewers, 1978). Joints, faults, and bedding planes are common elements, although each of these can be further subdivided.

The passages of some caves, such as Jewel Cave in the Black Hills, follow only joints (Deal, 1962). The joints commonly occur in a system, with several predominant directions, thus providing two or more structural elements. These caves commonly form an angular grid pattern of cave passages (Fig. 24A). The individual cave passages are generally narrow and elongated along the direction of the joint.

Other caves, such as Timpanogos Cave in Utah, follow faults. Although fault passages may resemble joint passages in cross section, their map pattern, a simple linear form, is different (Fig. 24B). Water flow may be encouraged through faults by porous breccias or physical openings within the fault plane (Ford, 1976, in Ford and Cullingford), but may be retarded by secondary mineralization (Ford and Ewers, 1978). Sweeting (1973) believed that faults more commonly influence cave passages than do joints, although North Ameri-

can literature generally seems to favor joints. This probably reflects the difference of structural setting between the central United States and Europe.

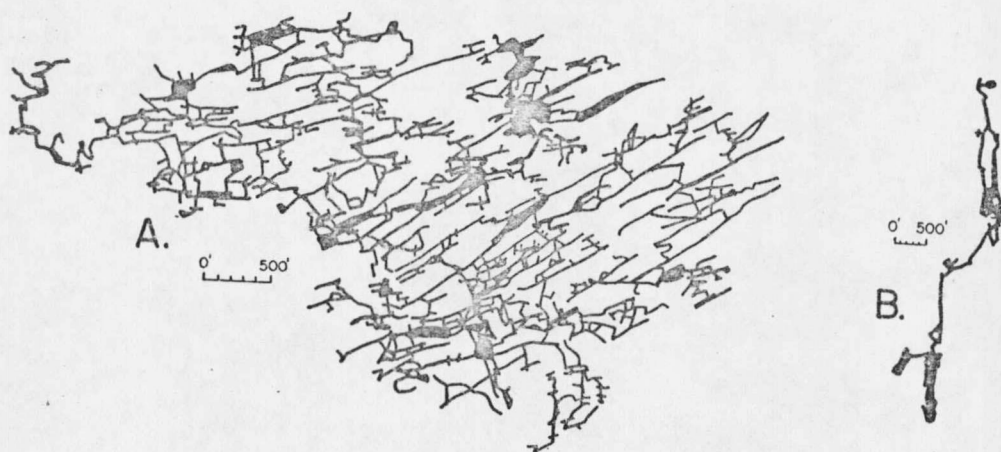


Fig. 24. A. Complex plan view of Jewel Cave, developed along a joint system. B. Simple linear pattern of Timpanogos Cave along two parallel faults. (From Conn and Conn, 1977, and White and Van Gundy, 1974).

Mammoth Cave in Kentucky shows very little influence of joints or faults (Ford and Ewers, 1978). Most of its passages meander along horizontal bedding planes. This bedding control does not show well on plan views, because the plane of the map parallels the plane of the bedding. However, the effect can be seen in cross section (Fig. 19D).

Bedding planes should be the most common structural element, because they continue throughout the limestone mass (Ford and Ewers, 1978). Joints are much shorter, possibly only extending tens of feet (Palmer, 1969), and faults are not always present. However, it is unlikely that the water can follow a single bedding plane from sink to spring, because the dip of the bedding is usually different from the hydrologic gradient (Ford and Ewers, 1978). Thus water must use a combination of several beds and joints along its flow path through limestone.

Dip Control

Caves in horizontal limestone have a different shape than those in steeply dipping limestone. American literature has been unduly influenced by studies of Mammoth Cave in the flat-lying limestones of Kentucky (Ford, 1971), and relatively little work has been done with caves in steeply dipping limestone.

The direction of cave growth, and the morphology of individual passages, partly depends on the amount of dip of the bedding. Caves in flat-lying limestone generally follow the dip, although it is very small (Ford, 1971). Subsurface drainage along the strike is common in steeply dipping lime-

stones, so that caves are elongated in this direction (White, 1960). Some dip passages do occur in steeply dipping limestone, but their shape is different than those in horizontal strata. Single dip tubes grow in limestone with dips greater than 5° , but bands of anastomosing tubes are found in flatter-lying limestones (Ford and Ewers, 1978).

Caves in flat-lying limestone naturally assume the morphology of water table caves, despite their fissure frequency. The soluble rock may commonly extend for 10 kilometers laterally, but have a depth of less than 100 meters (Thrailkill, 1968). Low fissure frequency would produce deep phreatic loops, but the magnitude of these loops would be small, because the limestone mass is long but thin (Ford and Ewers, 1978).

Deep caves may form in steeply dipping limestones, where the water moves down the bedding plane (Ford and Ewers, 1978). Thus deep phreatic caves may form, regardless of fissure frequency, if the water is trapped within the limestone by surrounding impermeable rocks. However, water table caves can form across the limestone beds if the water is unconfined, and if the folding which imparted a steep dip to the rock, increased the fissure frequency. This is common in the folded Appalachians (Davies, 1960).

Physiographic Controls

Relation to Surface Streams

Caves can only form if the surface streams are deepening their channels. This down-cutting provides a steep topographic and hydrologic gradient, so that subterranean flow is sufficient to dissolve the limestone. Thus caves usually form along river valleys. Caves are much rarer in the uplands away from the surface drainages (Davies, 1960; Miotke and Palmer, 1976). Few passages in the 180 mile long Mammoth Cave-Flint Ridge system of Kentucky cross completely under an upland; most are concentrated in the valleys where they discharge (Miotke and Palmer, 1976). Most of the cave passages in the folded Appalachians slope gently towards the surface rivers (Davies, 1960).

Age of Caves

Most caves formed in the geologic past. Many authors show evidence that caves formed during the late Mississippian Era, at the top of the Mission Canyon Limestone (Roberts, 1966; Campbell, 1977). Nearly all of these caves are filled with sediment. Late Mississippian or Pennsylvanian fossils in the fill may show that the cave formed and was open during

these times. This date may also be proved if the sediments show tilting by the Laramide orogeny (Campbell, 1977).

Most known caves formed during the late Tertiary and Quaternary. Several features of Carlsbad Caverns show that it formed previous to the present topography (Horberg, 1949). The modern surface drainages have cut into the abandoned phreatic cave passages, and have deposited speleothem fragments as surface gravel. In addition, the vast size and depth of the cave do not match the present, smaller drainage.

The surface drainage history, recorded in terraces, often allows the dating of cave events. During a pause in the downcutting of the surface rivers, the cave waters follow the same cave level for a long time, creating water table passages (Davies, 1960). The major levels of passages in the Mammoth Cave-Flint Ridge system have been correlated to terraces of the Ohio River system which formed during the interglacial periods. In addition, cave sediments can be correlated with episodes of deposition during glacial advances (Miotke and Palmer, 1976).

Miscellaneous Cave Features

Cave sediments

Caves naturally collect the sediments transported by karst water. Although few sediments can be carried when the passages are very small, sedimentation can occur throughout most of the cave's history. Cave sediments can be classified as allochthonous, autochthonous, or chemical (White and White, 1968).

Allochthonous sediments originate from outside of the cave, and are transported largely by fluvial processes, although downslope movement and glaciers may deposit debris at the cave entrance. These sediments come from other rocks in the drainage basin, and may include organic remains as well.

Autochthonous sediments come from sources and processes within the cave. Insoluble materials from the limestone, such as mud and chert, and guano from bats and other animals may accumulate in place, or be moved by stream water. Blocks of limestone, called breakdown, fallen from the ceiling or walls, are common in many caves.

Breakdown is released by many mechanisms. These include wedging in joints by ice or sulfate minerals, undercutting of the walls by vadose streams, and buoyancy loss due to

drainage of the cave (White and White, 1969). Breakdown reaches a depth of 135 feet in Wyandotte Cave in Indiana (Thornbury, 1969). The collapsed ceiling can be stabilized by acquiring an arched cross section. The process may continue, filling the room to the ceiling with rubble or by reaching the surface, forming a sinkhole (Fig. 25).

Most chemical deposits in caves, called speleothems, are composed of calcite. Although the majority of speleothems form after the cave has been drained, some deposits form in the phreatic zone. These are characterized by angular crystals, such as dogtooth spar. Many Black Hills caves have crystal coatings up to a foot thick, with individual crystals as large as goose eggs (Moore and Nicholas, 1964). Such large deposits required unusual temperature and chemical conditions (White, 1976, in Ford and Cullingford), but spar deposits are common in many caves.

In the vadose zone, loss of carbon dioxide is the most common mechanism for speleothem deposition. Evaporation also occurs, but most caves are too humid for this effect to be significant (Moore and Nicholas, 1964). There are dozens of speleothem forms, such as stalagmites, helictites, and others, each with a special origin. The reader can find detailed accounts of these in Hill (1976), White (1976, in Ford and Cullingford), and Kuhns (in press).

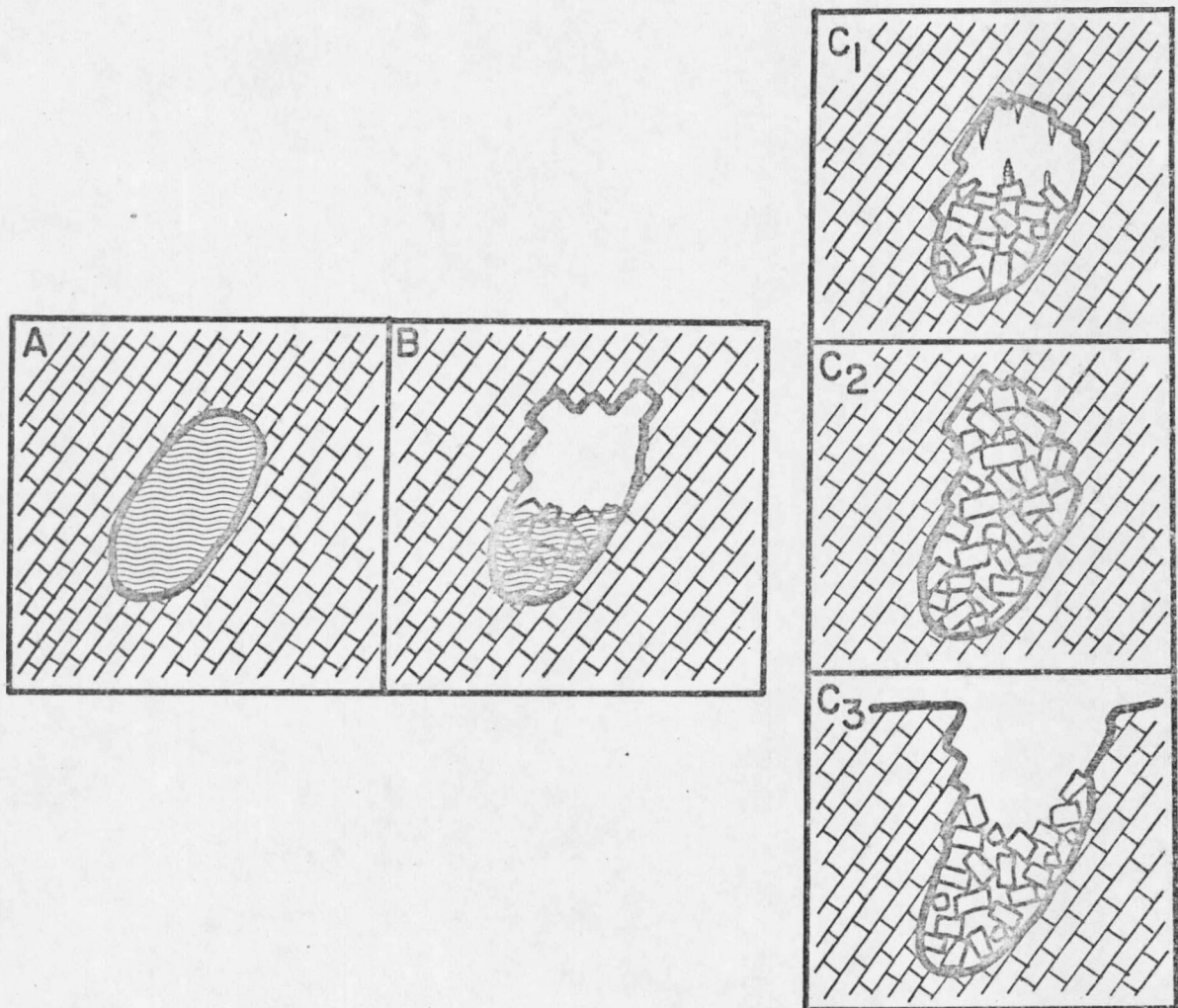


Fig. 25. Generation of cave breakdown. A. Solution of the caverns removes the bouyant force of the water, and the ceiling begins to collapse. Three resulting situations may occur: C1. Breakdown may cease due to the formation of an arched ceiling. C2. The breakdown may fill the passage to the ceiling, holding the roof up. C3. The collapse may reach the surface, forming a sinkhole.

Cave Winds

The movement of air can be felt in most caves (Wigley and Brown, 1976, in Ford and Cullingford). The winds can be caused by the resonance effect, barometric changes, or the chimney effect (Moore and Nicholas, 1964). Resonance winds in caves are caused by outside winds blowing across the mouth of the cave. This type of cave wind rapidly reverses directions, with cycles of several hours or less.

Barometric pressure changes create an imbalance of air pressure between the outside air and the cave atmosphere. These winds are strongest if the cave has a great volume and a small entrance. Barometric cave winds reverse in cycles of hours to several days, similar to barometric changes. However, the depth and size of the cave may cause the wind changes to lag behind the actual pressure change by several hours.

Temperature differences power most cave winds (Tuttle, 1977). The air temperature inside of a cave remains fairly constant due to the thick cover of bedrock. Compared to the outside air temperature, the cave air is warmer during the winter and colder during the summer. Thus, caves with two entrances at different elevations behave like chimneys, leak-

ing air out the top during winter, and losing air out the bottom during the summer (Fig. 26). These winds may flow in the same direction for months during the hottest and coldest parts of the year, and reverse within a single day during fall and spring.

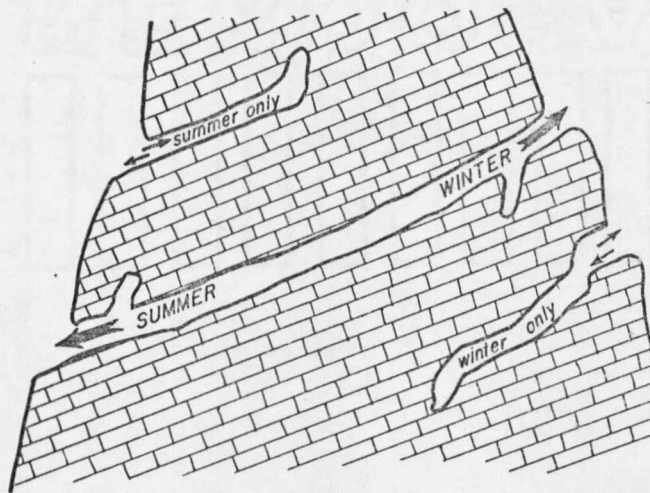


Fig. 26. Thermal circulation of cave winds. Air flows into the upper entrance and out the lower entrance of a two entrance cave during the summer, and flows in the opposite direction during winter. One entrance caves allow less wind circulation. Passages reaching up from their entrance move air during summer, descending passages leak air during the winter. (After Tuttle and Stevenson, 1977).

Single entrance caves may carry winds due to temperature effects as well (Tuttle, 1977). A cave that drops be-

low its entrance will allow slight winds during the winter, as the cold outside air sinks into the cave. Caves that rise above their entrance allow weak winds during the summer (Fig. 26).

Summary of Speleogenesis

Four elements are required for cave formation (Thornbury, 1969). First, soluble rock must be present at or near the surface. Deep wells drilled in the folded Appalachians rarely encountered caves where the limestone was buried by more than 100 feet (Davies, 1960).

Second, at least moderate rainfall is needed as the solute. Vegetation and soil development are also encouraged by significant rainfall, thus more carbon dioxide is available to form the solvent, carbonic acid. Ten to twelve inches of yearly precipitation is the approximate minimum required for karst development (Sweeting, 1973).

Third, the primary porosity and permeability of the limestone should be low, relative to the secondary characteristics. The chalk regions of England and France show that dispersed water flow through porous limestone does not develop large or numerous caves (Thornbury, 1969). Concentrated flow and solution along joints, faults, and bedding

planes leads to the formation of a sizable cave.

Fourth, the surface drainage must be actively entrenching into its valley. This increases the hydrologic gradient, and allows the water to rapidly flow through the rock.

CHAPTER 5

FEATURES OF LEWIS AND CLARK CAVERNS

Passages and Rooms

General Statement

Although Lewis and Clark Caverns has less than two miles of passageways, it is nearly 600 feet deep. This depth, tabulated during the thesis work, will make the cave about the 20th deepest in the United States, according to a recent list by Sprouse (1978). Future exploration and mapping may lead to the discovery of passages that connect the cave to the Jefferson River Canyon, perhaps doubling the present known depth.

Two natural entrances occur within 100 feet of each other at the top of the main cave. The Civilian Conservation Corps created a third entrance near the bottom of the cave, blasting a 538 foot long tunnel through the limestone. Another entrance, not yet positively located, allows wind circulation through the cave.

The plan view of the cave presents a bewildering pattern of passages (Plate III). The depth of the cave and the steep dip of the limestone superimposes rooms of the three levels of the cave on the plan view. In addition, cave passages

formed along at least five structural elements, the dip of the bedding, two directions of strike of the bedding, and two directions of joints.

Strike Passages

Most of the major passages in Lewis and Clark Caverns follow a component of the strike of the bedding. Only one of these, Uncle Sams Room, is horizontal, and thus follows the true strike. The remainder of the strike passages descend across the strike with a steep gradient, commonly about 28° , towards the axis of the Colter anticline (Plate IV). The map orientation of these passages is up to 10° from the true strike. Nonetheless, the bedding plane exerts the major influence on these passages, and their trend is closer to the strike than to the dip.

Three levels of strike passages follow bedding planes that strike N. 20 W. (Plate IV). The Garden of the Gods parallels Uncle Sams Room-Princess Palace, as well as the passage that connects the entrances to the Cathedral Room. Minor passages follow this same orientation, and include a tube above the Tapestry Wall in the Cathedral Room, and the area around Grannies Breakfast Nook (Fig. 27).

Northeast of the Cathedral and Paradise Rooms, the

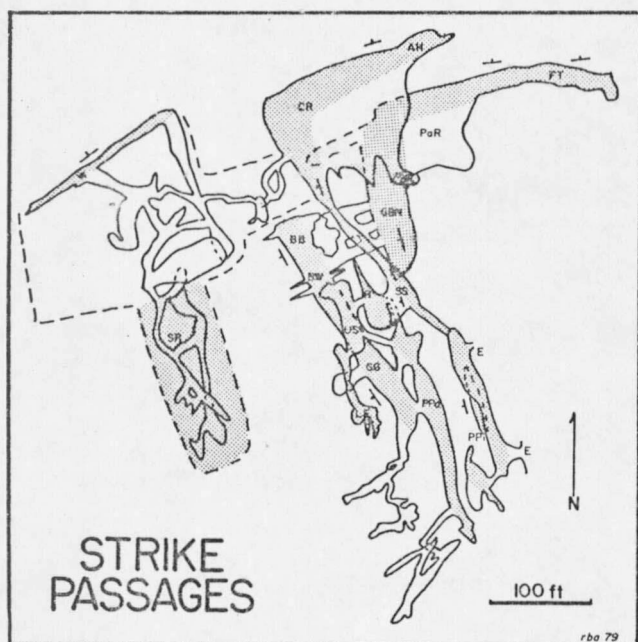


Fig. 27. Map of strike passages (stippled areas): Fish Tunnel (FT), Ash Hole (SH), Popcorn Room (PCR), passage from Cathedral Room (CR) through Spiral Stairway (SS) to the entrances (E), Grannys Breakfast Nook (GBN), Piggy Pits (PPI), Princess Palace (PPa), Rats Roost (R), Garden of the Gods (GG) through the Brown Waterfall Room (BW) and Brownies Battlefield (BB), Uncle Sams Room (US), Little Eagle Alley (LE), the Wall (W) in the Wind Tunnel, and possibly the Snow Room area (SR) of the Wind Tunnel.

strike trends ENE, ranging between N. 55 E. and N. 73 E. Two major strike passages, the Fish Tunnel and the Ash Hole, are found here. The Popcorn Room, extending east from the Cathedral Room, is a minor strike passage.

The bedding plane, which dips at 54° in nearly all parts of the cave, can easily be seen forming the slanting ceiling and wall of many of these passages. This is best viewed in the entrance area, and above the Spiral Stairway Shaft, and between the Brown Waterfall Room and Brownies Battlefield (Fig. 28A&B).

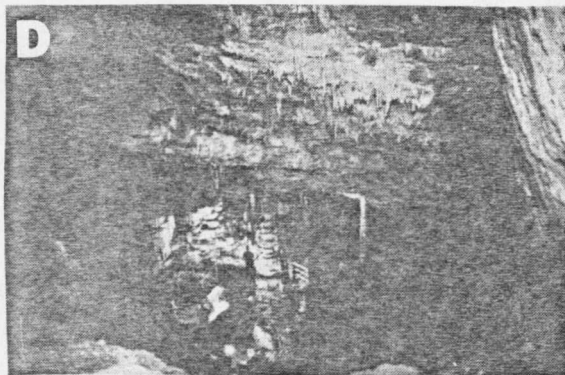
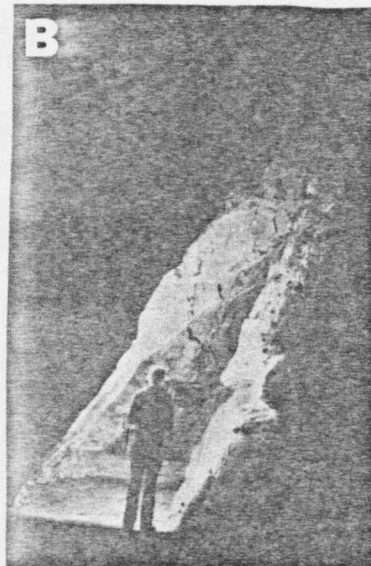
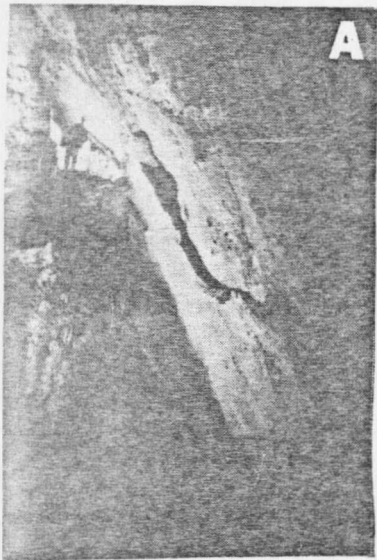


Fig. 28. A. Bedding plane of the Mission Canyon Limestone dipping steeply to the southwest (right in picture) along strike passage. Person in background for scale. Spiral Stairway Room. B. Cross sectional effect of bedding plane dipping to the left in strike passage. Entrance area. C. 100 foot deep dip passage between two major strike passages. Spiral Stairway Shaft. D. Looking from strike passage (Fish Tunnel) into junction room (Paradise Room). Bedding plane dips steeply to right.

In other locations, the bedding plane is obscured by joints. The overall trend of Uncle Sams Room and the Garden of the Gods follows the strike of the Bedding, but joints intersect portions of the bedding planes, producing locally vertical or stepped walls.

Dip Passages

In four locations, cave passages are elongated along the dip rather than the strike. Two such passages descend about 20 feet from the floor of the strike passage near the cave entrance, but widen along the strike at their base, forming small rooms. The dip passages, called the Piggy Pits, were partly filled with debris by early workers at the cave, so their exact size and shape is difficult to determine.

The third dip passage, the Spiral Stairway Shaft, drops 90 feet down the dip, connecting two major strike passages (Fig. 28C). Although the Piggy Pits do not continue above the strike passage, the Spiral Stairway Shaft extends at least 30 feet into the ceiling, narrowing from a diameter of 25 to 8 feet. Dangerous climbing conditions prevented the exploration of this area.

The entrance to the Nasal Passages, another dip section,

lies 125 feet above the entrance to the main cave. This is currently a separate cave, and is not physically connected to the main cave, but probably connected to the Ash Hole during speleogenesis. The Nasal Passages are 67 feet deep and probably extend deeper, but are filled with sediments. The passages roughly follow the plunge of the Colter anticline, trending NW and descending at an average angle of 45° . The dip of bedding planes can be seen in the Nasal Passages, and both dip and strike are clearly visible outside and above them.

Junction Rooms

The two largest known rooms of the cave are the Cathedral Room and Paradise Room. The rooms formed on the axis of the Colter anticline, where the major strike passages meet. Thus both rooms have similar shapes (Fig. 29). Deep breakdown floors, large stalagmites, and the abundance of other speleothems, typify these rooms (Fig. 28D).

Joint Passages

Several passages and rooms follow major joints rather than the bedding planes. The walls of these joint passages usually are vertical, compared to walls that dip at 54° along

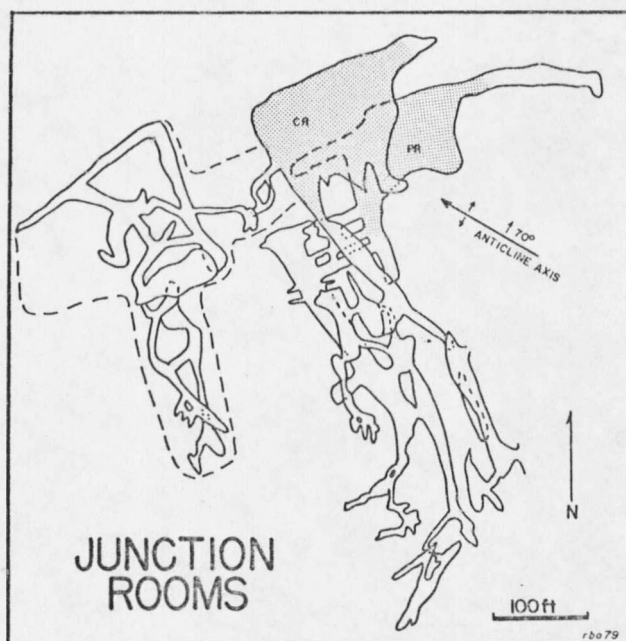


Fig. 29. Map of junction rooms (stippled areas) Cathedral Room (CR) and Paradise Room (PR). The axial plane dips about 70° to the northeast (Fig. 10A), thus the lower junction room, the Paradise Room, lies northeast of the Cathedral Room.

the bedding plane, as seen in the strike or dip passages. In addition, their map orientation differs sharply from that of the bedding plane passages.

The Princess Dungeon Room off the Princess Palace trends NE, while the Princess Palace follows the strike of the bedding at N. 20 W. (Plate III). Several sets of crawlways, as well as the main room, follow the same NE orientation. In addition, joint controlled walls can be seen at several points. The Bacon Room also follows a NE trend, perhaps

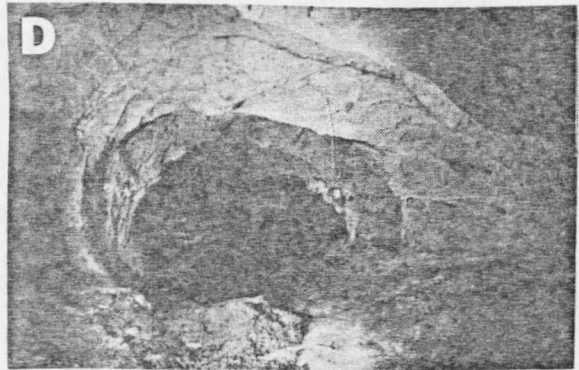
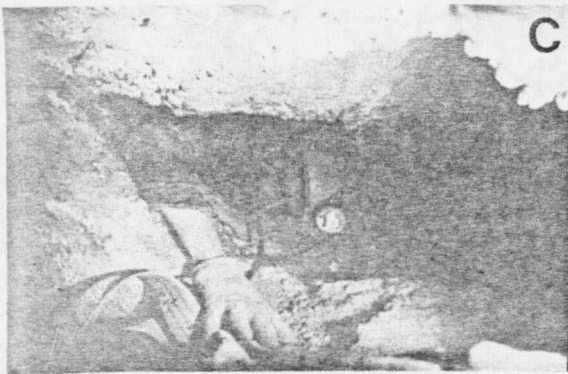
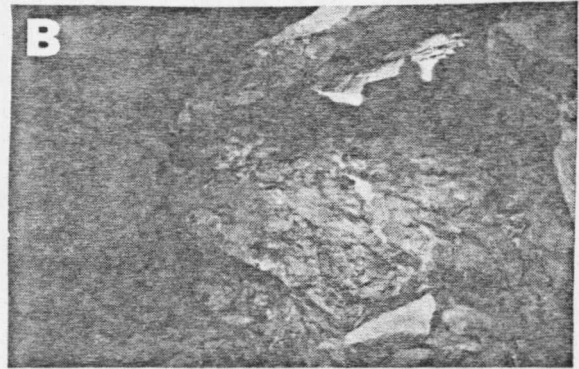
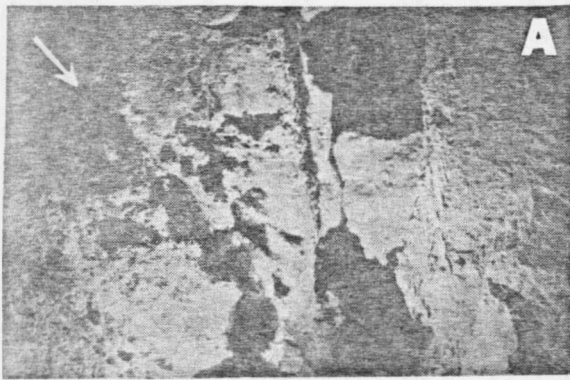


Fig. 30. A. Passage developed along vertical joint, cutting across different beds (arrow points along dip). Grand Canyon Room. B. Large room (seated person for scale, upper right) completely enclosed in breakdown. Shining Mountain Room, Wind Tunnel. C. Smaller rooms in breakdown prevent complete exploration. Wind Tunnel. D. Phreatic elliptical cross section. Upper Nasal Passages. E. Wall pockets, average width three feet. Spiral Stairway Shaft.

along the same joint system. Extensive breakdown has modified the Bacon Room area, so its geologic relations are difficult to observe.

On the lower level of the cave, major joint passages intersect the long strike passage, called the Garden of the Gods, at the Brown Waterfall Room. At the intersection, the passage becomes larger. The strike passage abruptly ends in the next room, Brownies Battlefield. These two joint rooms are elongated slightly in an east-west direction, and have very high ceilings and deep breakdown floors (Fig. 30A).

Many small passages extend off the strike passage, but pinch out in 20-30 feet. These side passages are tall, narrow cracks, and probably formed along individual joints. The Brown Waterfall Room and Brownies Battlefield developed along many joints, which are seen in the ceiling, and as vertical, east-west walls which parallel the small, side passages. Adjacent to these two large rooms, other similar large rooms continue to the east. These rooms, the Horses Barn and Grand Canyon Room, have breakdown floors as deep as 65 feet.

On the upper level of the cave, the Wind Tunnel-Hells Half Acre extend west from the Cathedral Room. Only one wall in the entire area has been identified as bedrock in

place. This wall forms the northern boundary of the Wind Tunnel area, and can be traversed for a distance of 150 feet horizontally, and 100 feet vertically. The remainder of the Wind Tunnel area appears to be continuous breakdown.

Breakdown Rooms

Although breakdown floors are common in most of the passages of Lewis and Clark Caverns, many rooms of the cave have walls and even ceilings that are made of breakdown. A single wall may be made of only one block, or of many blocks. The shape of these rooms is highly irregular in all three dimensions, and they interconnect and overlap with numerous other "rooms". Thus, it is nearly impossible to draw an adequate map of these areas. The breakdown rooms are analogous to cave openings found in deep talus piles at the base of mountain slopes.

Many such rooms have been named in the deep breakdown floors of the Paradise Room and Grand Canyon Room. These smaller rooms include the Goats Walk, Milk Room, Pinecone Room, Glow Room, Flathead Cherry Passage, and Ray Kellys Tunnel. The walls that separate the larger rooms of the area, such as Brownies Battlefield, the Horses Barn, the Grand Canyon Room, and the Paradise Room, are also possibly break-

down blocks. These walls are discontinuous and penetrated by several small passages. The routes which connect the large rooms show no evidence of solution, but rather resemble other smaller gaps in the breakdown. The very large blocks that make up these walls have not moved far, because the strike and dip of their bedding planes matches from wall to wall.

The hundreds of small and large rooms of the Wind Tunnel area combine to form a breakdown maze. Over a dozen traversable routes reach the Snow Room. The largest known room in the Wind Tunnel, the Shining Mountain Room, is about 120 feet by 80 feet in size (Fig. 30B&C). Its walls and floor are clearly breakdown, but its ceiling is possibly in place, perhaps a continuation of the bedrock wall which forms the northern boundary of the Wind Tunnel.

Maze Passages

Several sections of the cave deviate from the normal, single conduit to a swarm of smaller, interconnecting tubes. These maze sections have elliptical cross sections, wall and ceiling pockets as well as tubes, and bedrock spans (Fig. 19B&C, 30D), but speleothems are relatively rare. The mazes are localized along portions of both strike and dip passages and are found in those that have the steepest gradients.

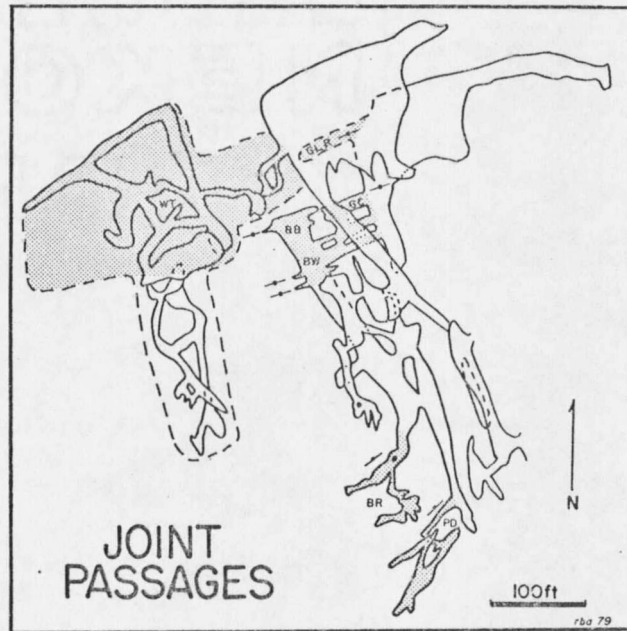


Fig. 31. Map of joint passages (stippled areas): Glow Room (GLR), Grand Canyon Room (GC), Brownies Battlefield (BB), Brown Waterfall Room (BW), Bacon Room (BR), Princess Dungeon (PD), and most of the Wind Tunnel area (WT).

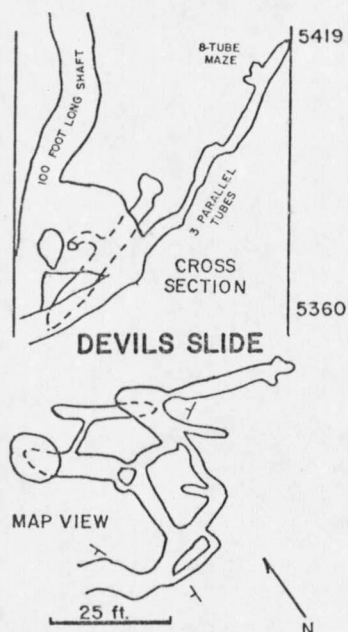
Some of the maze passages are not fully developed, and perhaps should be called "proto mazes". They show features such as pockets and tubes, which are the result of phreatic solution under a large hydrostatic head (Bretz, 1942). Continued solution and pressure may have caused these features to grow into interconnecting tubes, thus a "mature maze". Phreatic pockets can be seen in the bottom of the

Spiral Stairway Shaft and portions of Uncle Sams Room that connect to the shaft. The hollows are nearly round and do not resemble scallops from uni-directional, turbulent water flow (Fig. 30E). Many pockets also occur in the Fish Tunnel, where bedrock spans are also seen (Fig. 19C).

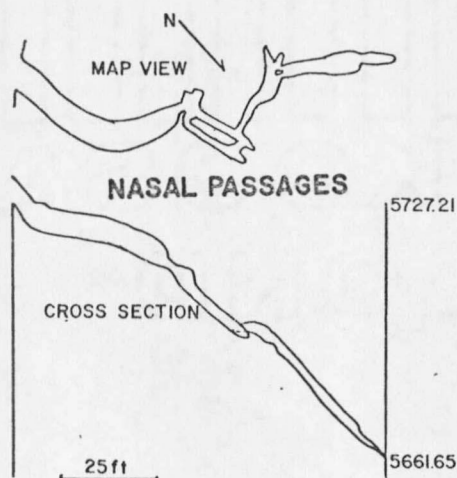
The Nasal Passages are two long, parallel, elliptical tubes which are connected at their tops and bottoms (Fig. 32). As mentioned on page 98, the bottom of the Nasal Passages is filled with sediments. However, an adjacent tube continues below for 30 feet, before pinching out (Fig. 32).

The best developed maze passages in Lewis and Clark Caverns occur in the Devils Slide area, above the Paradise Room and Grand Canyon Room. Three traversable, adjacent tubes run parallel to each other, and many interconnecting tubes reach off from them (Fig. 32). Downstream, the tubes combine into a single passage, which intersects the junction of the wall and ceiling of the Grand Canyon Room. Upstream, the Devils Slide area continues as a nearly vertical tube, splitting into nearly ten tubes, most of which are too small to traverse. A large vertical shaft, the Link Shaft, connects to one side of the area, but shows no maze features.

Fig. 32. Maze sections of Lewis and Clark Caverns.



Devils Slide made of many inter-connecting and blind phreatic tubes. Strike of bedding changes 70° within the maze, showing that the maze formed on the axis of the Colter anticline. No vertical exaggeration on cross section.



Nasal Passages made of two main, parallel tubes, and other phreatic passages. Change of strike of bedding cannot be observed inside the passages, but is well exposed outside, immediately above them. No vertical exaggeration on cross section.

Cave Sediments

Very few insoluble sediments are found in Lewis and Clark Caverns, but vadose silt, and smectite muds occur in localized areas. The major sediment is breakdown, which was discussed on pages 101-102.

Vadose Silt

Four feet of silt collected in the Cathedral Room, just below the entrance to the Ash Hole which is completely filled with the sediment. The silt coats the loose popcorn walls of the Link Shaft below, which reaches down 125 feet past the Devils Slide to the Paradise Room. The silt fills the floor of the Pinecone Room area below.

The silt is composed largely of volcanic ash, although phlogopite flakes are common. The material is tan-colored, but locally white, where pure ash makes up the laminae (Fig. 33A&B). Similar ash can be seen on the walls of the Nasal Passages, where angular limestone gravels are included in the silt. Gravels are also seen in the Ash Hole, but not in the Cathedral Room deposits. However, vadose speleothems are found in both the Ash Hole and Cathedral Room silt.

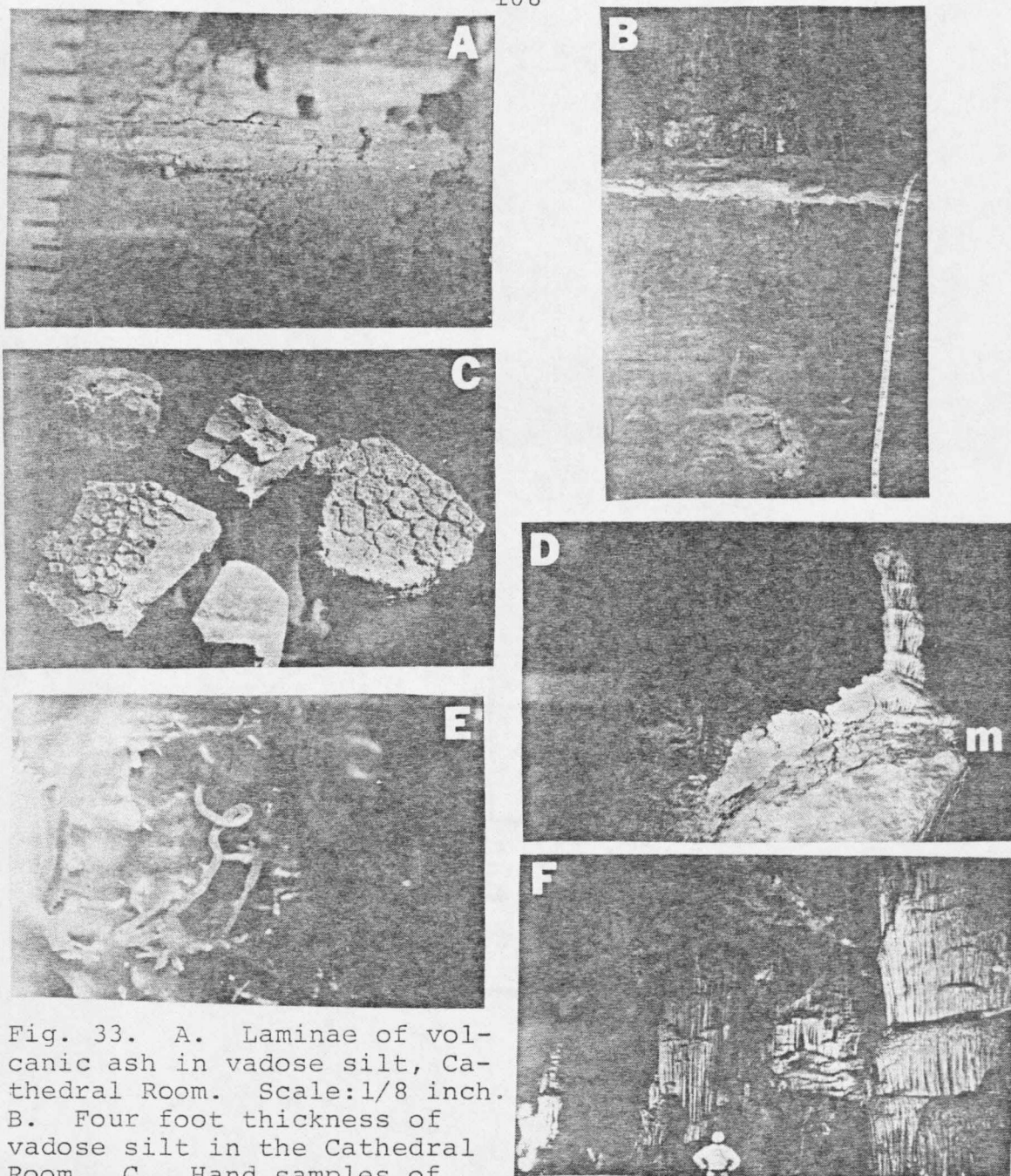


Fig. 33. A. Laminae of volcanic ash in vadose silt, Cathedral Room. Scale: 1/8 inch. B. Four foot thickness of vadose silt in the Cathedral Room. C. Hand samples of smectite muds, showing dessication cracks. Samples average four inches in width. D. Dark-colored smectite mud on top of isolated breakdown block, and below thick cover of speleothems. The muds (M) are 8 inches thick. Paradise Room. E. Curiously spiralling helictite. Width of view about 2 inches. Pinecone Room. F. Large columns. Paradise Room.

Wilma Karsmizki recovered a metapodial bone from a Recent deer (Melton, oral commun., 1979) while digging through the Ash Hole silt.

Smectite Muds

Light-colored muds composed primarily of smectite and calcite occur in scattered portions of the lower half of the cave. Quartz and illite are minor constituents (Berg, written commun., 1978). The muds are commonly light brown, but orange, red, and black colors can also be found. Although some of the muds are slightly loose and damp, most are lithified. Dessication cracks appear on most surfaces, even across the laminar beds of muds (Fig. 33D).

As described by Kuhns (in press), the muds are found on and between blocks of breakdown (Fig. 33C). However, none occur in the breakdown areas of the Cathedral Room and Wind Tunnel, but are only found in the Princess Palace, and rooms on the lower level of the cave.

Speleothems

Secondary mineral deposits blanket many of the walls, floors, and ceilings of the caverns. Calcite is the most abundant mineral, but aragonite is found in some of the

coralloid speleothems (Kuhns, in press). Hematite and limonite are common coloring agents, especially in the Brown Waterfall area, and pyrolusite and rhodochrosite occur in the Snow Room area of the Wind Tunnel.

Although the cave has many large stalagmites and stalactites, erratic speleothems (Hill, 1976) are found in unusual abundance. Helictites and shields are common, but always spectacular (Fig. 33E). Subaqueous forms can be found in small spots throughout the cave, but many are covered by the vadose speleothems. Many portions of the cave, especially those deep inside of Cave Mountain, still host actively growing speleothems (Fig. 33F).

Cave Winds

Wind flows through nearly every room in the caverns, but is strongest at the entrance to the Wind Tunnel. During the summer, the wind flows into the cave from both entrances at the top of the tourist trail, and enters the Wind Tunnel and Bacon Room. The air flows from these passages and out the entrance during the winter. During fall and spring, the winds reverse within a single day, if the outside temperature changes greatly. This suggests air flow by the chimney effect, and requires the presence of another

entrance, one that is lower than the two at the top of the cave. The exact location of this entrance is unknown.

The Exit Tunnel of the cave does not carry the cave wind. The Civilian Conservation Corps installed doors at both ends of the tunnel, and the doors are never opened at the same time, preventing winds that could dry out the Paradise Room. In addition, the Exit Tunnel is on the east side of the cave, but the Wind Tunnel and Bacon Room, which together carry the main air flow, are on the west side of the cave.

The unknown entrance is probably a small cave passage found in the Limespur quarry, one mile W.S.W. of the cave. The passage follows a joint to the E.N.E., and can be explored for about 100 feet. However, it sucked air during winter and blew air during summer, indicating that it has another entrance at higher elevation. Preliminary air-tracing experiments suggest that the two may be connected.

CHAPTER 6

SPELEOGENESIS OF LEWIS AND CLARK CAVERNS

Lithologic Controls

Lewis and Clark Caverns formed in the lower portion of the Mission Canyon Limestone, and the upper 30 feet of the Lodgepole Limestone (Fig. 34A). The change of lithology at the formation contact probably determined the stratigraphic position of the cave. The top of the Lodgepole Formation is marked by 20% insoluble material, but the basal beds of the Mission Canyon Limestone generally contain less than 2% insoluble material (Sloss and Hamblin, 1942). This insoluble layer at the top of the Lodgepole Limestone retarded the downward movement of groundwater, and concentrated both the flow and solution at the base of the Mission Canyon Limestone.

Cave development up-section in the Mission Canyon Limestone was possibly limited by a zone of solutional breccia. The breccia was tentatively identified 450 feet above the base of the Mission Canyon Limestone at the Logan section, 20 miles east of the cave (Sando and Dutro, 1974). The breccia is well exposed along the northern walls of the Jefferson Canyon. Numerous shallow holes, produced by collapse after solution, as well as strong limonite staining

mark the outcrops of the breccia. Solution breccias are typically poorly jointed, and contain high amounts of clay and insoluble chert and dolomite, thus resist cave formation (Campbell, 1977).

Despite the abundance of speleothems, the petrology of the wallrock can be observed in numerous places in the caverns. Gray biosparite is the most common lithology (Fig. 34B). In many locations, such as the Shining Mountains Room in the Wind Tunnel and Uncle Sams Room, the fossil fragments are resistant, but the matrix and cement are quite loose, making the overall hardness of the limestone less than that of a fingernail. The fossils are much harder to see where this same lithology is exposed outside of the caverns or in artificial tunnels of the cave. In these locations, the rock is quite hard.

Yellow micrite is also commonly seen in the cave, although less often than the biosparite. Chert occasionally occurs as small, round nodules in the micrite. The micrite and sparite can be easily recognized in the artificial tunnels, just below the cave entrance and between the Cathedral Room and the bottom of the Spiral Stairway Shaft.

Although individual beds are quite visible in the biosparites, the beds are all similar, and cannot be correlated

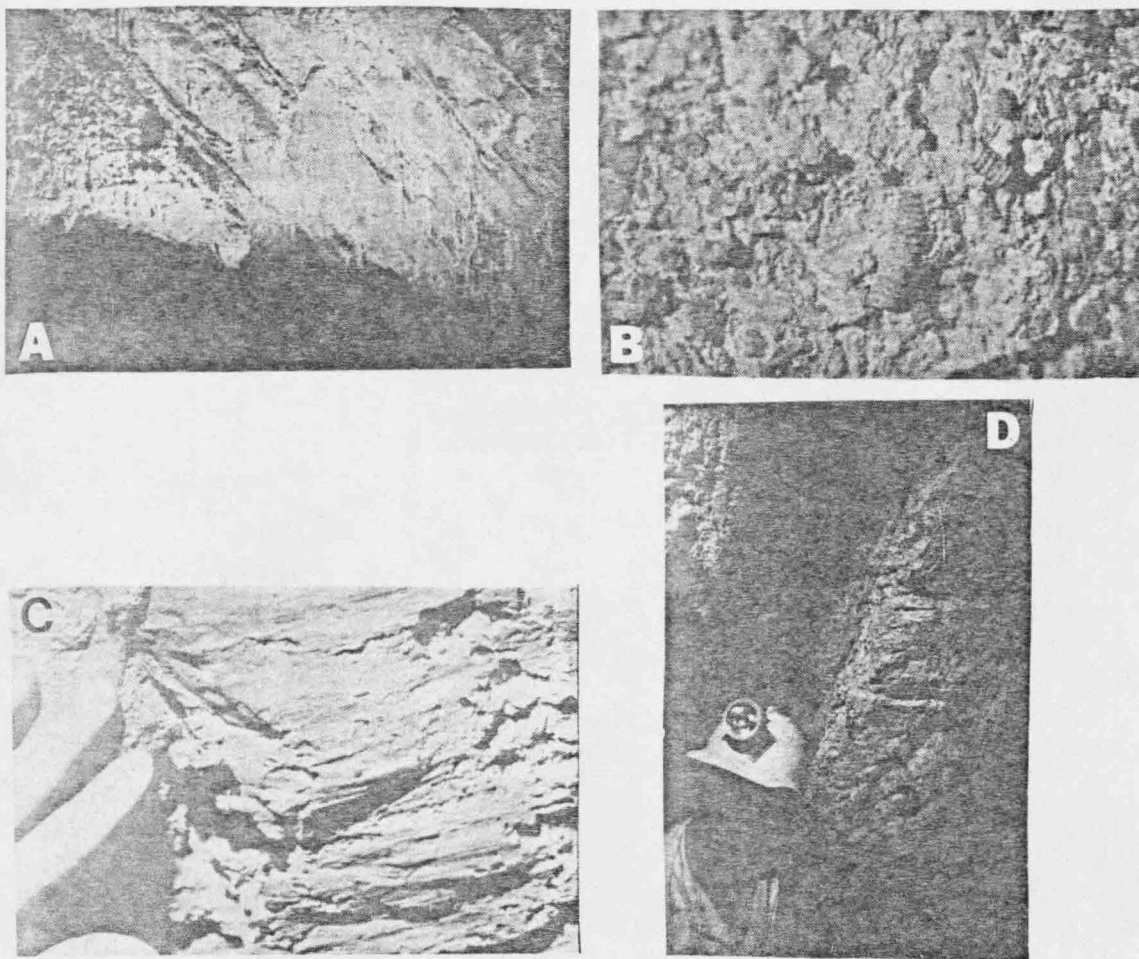


Fig. 34. Thin beds of the uppermost Lodgepole Limestone exposed in the Paradise Room. Beds average 4-6 inches in thickness. B. Crinoidal biosparite, showing preferential solution of matrix and cement rather than fossil fragments. Width of view: 2 inches. Uncle Sams Room. C. Slickenslides. Above entrance to Sample Room. D. Six-inch thick fault gap. Dense breccia occurs within shadowed fault. Northern end of the Grafitti Wall Hallway, Wind Tunnel.

between individual rooms or passages. However, it is clear that the passages formed along several different beds, rather than along the same bedding plane. The Garden of the Gods formed directly below Uncle Sams Room, not downdip, thus in a different bedding plane. Similarly, Uncle Sams Room and Little Eagle Alley formed in different beds, because they lie at the same level, next to each other. Other examples include the Ash Hole-Popcorn Room, the Fish Tunnel-Ash Hole, and the Garden of the Gods-Grannies Breakfast Nook area (Plate III).

Structural Controls

Introduction

The major structural feature of Lewis and Clark Caverns is the Colter anticline. Strike passages formed on both limbs of the anticline, and met at the fold axis. Joint passages developed at the axis along fractures caused by the folding.

Anticline Versus Syncline

Cave formation was favored on the Colter anticline, but apparently was retarded on the Potts syncline. The scarp slope of the beds on the anticline forms a shallow, but prominent valley down the southern face of Cave Mountain

(Fig. 7D), but the syncline forms a resistant ridge (Fig. 7E). Thus, surface water was funnelled into the anticline, but the water was dispersed on the syncline.

Most passages on the plunging anticline followed the strike, and only a few are along the dip. Dip passages would diverge away from the anticlinal axis, forming many small passages, but the strike passages would converge together at the axis, where large, efficient trunk passages could form. Greater fracturing along the axis also increased the permeability, encouraging flow towards the axis. If any cave development occurred on the adjacent Potts syncline, dip tubes would probably predominate, because they would carry water to the axis more efficiently than strike passages.

Bedding Plane Control

Faults along the bedding planes probably encouraged the formation of many passages along the beds rather than along joints. The tight folding of the Colter anticline caused the beds to slide past each other, forming faults with unknown, but probably small, displacement. These faults do not offset bedding, so can only be recognized by the presence of fault breccia or slickensides. Thin, but continuous breccias can be seen in Shining Mountain Room of the Wind Tunnel, and between the Sample Room and the Discovery Hole

entrance, along the tourist trail. A large gap along a fault plane, filled with a six inch-thick breccia, can be seen in the northern end of the Graffiti Hallway in the Wind Tunnel (Fig. 34C). In addition, slickensides are well-preserved along the western wall between the Spiral Stairway Shaft and the Sample Room (Fig. 34D). Breccias and slickensides along the bedding planes probably determined the position of other passages in Lewis and Clark Caverns, but the solutional enlargement of the faults removed the evidence.

The steep dip of the beds also favored flow along the bedding planes. Groundwater could follow any path ranging from 54° to nearly 0° , by following the dip, the strike, or an angle between. In flat-lying limestone, the water can follow a steep gradient only by exploiting a large number of joints (Ford and Ewers, 1978).

Joint Control

The 45° plunge of the nose of the Colter anticline encouraged groundwater to exploit joints, rather than the bedding plane, at the fold axis. The 45° angle was much steeper than the 28° average gradient of the individual strike passages. Thus, the water abandoned its course

along the bedding planes, and exploited joints at the axis.

The Wind Tunnel and Brownies Battlefield-Grand Canyon Room area formed along east-west joints near the fold axis, because of increased permeability at the axis. Fracturing along the axis should have produced joints along one or two other directions, but the cave development occurred only on the east-west system. The most direct route to the Jefferson River Canyon, the probable discharge point of the cave, is to the west, favoring the east-west joint system.

The east-west fractures near the anticlinal axis parallel the nearby Cave fault and Jefferson Canyon fault, but are joints produced by the fold, rather than a third, parallel fault. The fractures cut across bedding planes along the southwestern limb of the fold, but show no offset of the beds. In addition, no offset is visible on the cliffs outside of the cave.

Hydrologic Controls

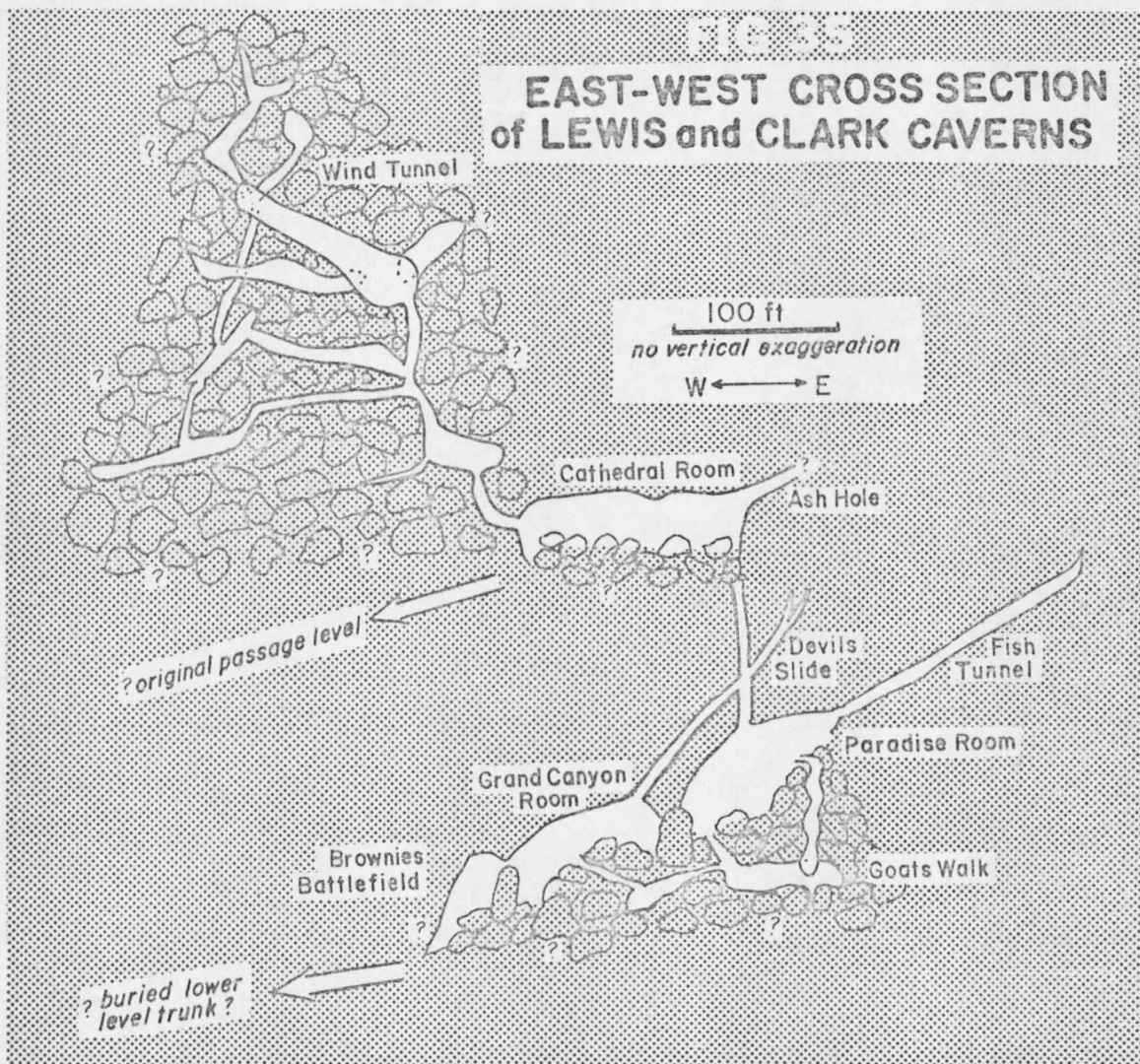
Introduction

Lewis and Clark Caverns formed in a branchwork pattern, with only minor, localized maze sections. Many features indicate that the cave formed in the phreatic zone, but little evidence suggests that solution took place along the water table, or in the vadose zone.

Branchwork Pattern

Palmer (1975) incorrectly classified Lewis and Clark Caverns as a maze cave, based on the confusing plan view of the caverns. New mapping of the caverns (Plate III) and the cross section view of the cave (Plate IV) suggest a branchwork pattern. The strike passages occur as narrow, elongate bands with few connections between, rather than a wide mesh of interconnecting tubes. Tributary passages, such as Little Eagle Alley and Princess Dungeon feed into the main strike passages. The strike passages converge at the fold axis, where a major trunk passage formed on two levels of the cave. The enormous size of the Wind Tunnel area and Grand Canyon Room-Brownies Battlefield indicates that they served as these trunk passages (Fig. 35). In addition, the long strike passage, the Garden of the Gods, is terminated where it intersects the trunk passage in the Brownies Battlefield. The trunk passage probably continued to the west, but has apparently been buried by breakdown. No other passages of significant size leave the area of the fold axis, so only these large rooms could have carried the water away. Finally, the wind circulation through the Wind Tunnel shows that the passage reaches outside of the cave. This lower,

unknown entrance probably also served as the discharge point for the cave water.



The branchwork pattern indicates that water entered the cave at fairly distinct points, rather than dispersed through millions of cracks (Palmer, 1975). The water was probably collected in landforms similar to shallow sinkholes, elongated along the bedding plane. However, the upstream end of the

passages has since been eroded or filled with sediments, so the exact nature of the local catchment area cannot be determined. The Ash Hole apparently retained some of its original surface morphology during Recent times, which allowed the flow of vadose water and sediments through the former phreatic system.

In summary, water entered the cave through many distinct points, flowing along the bedding planes to the fold axis. The combined flow of water exited the known portions of the cave through trunk passages. The exact size and shape of the trunk passages has since been masked by breakdown.

Termination of the Middle Level

Only one of the five major strike passage systems does not reach the trunk passages at the fold axis. The Princess Palace-Uncle Sams Room is nearly 400 feet long, but the passage abruptly ends at the north wall of Uncle Sams Room. The walls and floor appear to be bedrock and not breakdown, suggesting that the water flow changed direction here. The flow was probably pirated by the upper strike passage that connects the entrances to the Cathedral Room.

Piracy may occur between passages early in the development of a cave system when one passage reaches its discharge point before its neighbors (p. 76). Neither of the

strike passages above or below the pirated level could have reached the ultimate discharge point of the cave, the Jefferson River Canyon, but both reached the joint system, and met with other strike passages. This probably served as a local discharge target.

Many lines of evidence suggest that groundwater flowed from Uncle Sams Room, up the Spiral Stairway Shaft, and discharged through the Cathedral Room and Wind Tunnel on the upper level of the cave. Several short passages connected Uncle Sams Room to the bottom of the shaft. Although these were partly filled in the 1950's, wall pockets, carved by flowing water in the passages as well as at the bottom of the shaft, are still visible. Water flow up the 90 foot shaft would require a large hydrostatic head, and these pockets record its presence.

The large diameter of the Spiral Stairway Shaft indicates that it carried a significant flow of water between the two levels of the cave. Above the upper strike passage, the small diameter of the Shaft shows that it carried little water flow there, and that the entire shaft was probably the size of the narrower, upper shaft before the piracy.

The water could not have flowed down the main shaft from the upper strike level, because there is no known outlet in

Uncle Sams Room. The large size of the upper strike passage suggests that water was added to it by the shaft. The diameter of this strike passage is larger downstream of the Spiral Stairway Shaft than upstream. Thus water flowed up the Spiral Stairway Shaft from Uncle Sams Room, joining the flow of the upper strike passage en route to the Wind Tunnel entrance in the Cathedral Room.

Some evidence, however, shows that an outlet path through the Garden of the Gods is possible, though less likely than up the Spiral Stairway Shaft. A downhill path from Uncle Sams Room would be shorter and simpler. Although a hydrostatic lift of 90 feet is less than some discussed by Ford (1971), a source for that head cannot be proven. The highest elevation in the Princess Palace-Uncle Sams Room is only half the needed distance. However, the passage probably continues beyond what can be explored. Finally, the northwest corner of Uncle Sams Room extends downdip at least 25 feet towards the Garden of the Gods. However, this extension pinches out before reaching the Garden of the Gods.

A third optional discharge point for the Princess Palace-Uncle Sams Room is the joint controlled area near the fold axis. Uncle Sams Room extends as far north as the joint-controlled Brown Waterfall Room and almost as far

north as the Wind Tunnel, but misses both of these passages vertically by over 100 feet. Uncle Sams Room possibly entered this joint system; but there is no sign of the presence of such a passage, nor a corresponding strike passage on the other limb of the anticline.

In summary, groundwater probably flowed from the Princess Palace-Uncle Sams Room level to the upper level through the Spiral Stairway Shaft, and exited the caverns through the Wind Tunnel.

Maze Sections

Maze patterns formed at two locations in the caverns. Both mazes occur along the axis of the Colter anticline (Fig. 32). A relatively high density of fractures along the fold axis provided many competing routes for the groundwater flow, forming a maze pattern (Palmer, 1979, written commun.).

Both of these maze sections, as well as the "proto-maze" areas, which display phreatic features such as ceiling tubes, occur in the steepest passages in the caverns. Water flow through the many irregular maze passages was less efficient than flow through the major strike passages, due to lower permeability. Thus, a steeper gradient was needed to power the water through the maze sections.

The maze tendency may have been aided by the buildup of the hydrostatic pressure because of a bend in the passage. The Devils Slide and Fish Tunnel enter large breakdown rooms at the modern ceiling level (Fig. 35). Previous to the falling of the breakdown, the ceiling was much lower (Fig. 25 A&B), and was not in grade with the slope of the maze passage. Thus, the passages probably bent sharply down to meet the larger rooms. This postulated bend in the passage may have constricted the water flow and built up the hydrostatic pressure, further favoring the maze pattern.

In conclusion, the relative abundance of joints along the axis of the Colter anticline caused the formation of the maze passages. However, large rooms, such as the Paradise Room, also formed on the axis. The large rooms were each fed by several major strike passages, but the small, maze passages formed where less water entered the fold axis.

Phreatic and Vadose Solution

The cave clearly formed in the phreatic zone, and was not significantly modified by vadose streams. Several fea-

tures in the caverns record its solution in the phreatic zone. Ceiling and wall pockets are common. The maze sections also show that solution occurred in filled conduits under hydrostatic pressure. Only one feature, the Link Shaft, possibly formed by vadose solution. Its nearly vertical path, with no obvious structural control, suggests vadose gravity flow (Bretz, 1942). However, no signs of vadose solution can be seen above the shaft, where the Ash Hole carried the silt into the Cathedral Room, or below in the breakdown floor of the Paradise Room. Slight scouring from the clastic sediments is visible, but vadose channels are not. Furthermore, deposition occurred during at least part of the vadose flow, evidenced by the abundant popcorn speleothems that are interbedded with the silt on the walls of the Link Shaft. Thus vadose solution played only a small role, if any, in the formation of Lewis and Clark Caverns.

The factors which favored the solution of the cave in the phreatic zone, rather than the vadose zone, are unclear. According to criteria suggested by Ford (1971), the cave should have formed in the vadose zone. The cutting of the Jefferson River Canyon provided high topographic relief. In addition, the permeability of the limestone at Cave Mountain was high, due to fracturing along the Colter anticline. Both of these factors would encourage a low water table, and

thus a thick vadose zone. The cave possibly formed with both vadose and phreatic levels, but the higher, vadose sections have since been removed by the erosion of Cave Mountain. However, there is no evidence for this.

The modern topographic expression of Cave Mountain has discouraged recent vadose modification. Only a small catchment area supplies water to the cave, but the resistant peak does not concentrate the slight amount of water together, but disperses it over many routes. Vadose water reaches the cave today as millions of drips and seeps.

Water Table Control

Lewis and Clark Caverns formed with no apparent water table control. Passage levels of a water table cave, such as the Mammoth-Flint Ridge system, can be correlated with terrace levels of the local surface river (Miotke and Palmer, 1972). However, the three levels of passages in Lewis and Clark Caverns lie over 1,000 feet above the modern Jefferson River channel, but the two terrace levels lie less than 150 feet above the floodplain. In addition, the vertical spacing between the cave levels averages over 100 feet, while the terrace levels are within 50 feet of each other (Reshkin, 1963).

The steep gradient of the passages in Lewis and Clark Caverns also suggests that the water table did not control the formation of the passages. Water table passages in the folded Appalachians average less than 2° (Davies, 1960). Similarly, state 1 cave passages normally have a very low gradient (Ford, 1971). Thus the 28° average gradient of the "strike" passages of Lewis and Clark Caverns does not fit this model.

The cause for the steep gradients of the passages is not clear, but structural control may have played a role. Although the passages clearly follow the bedding plane, the gradient of the passages may have been caused by the angle of intersection of the beds with joints. The junction of the bedding planes with these joints parallels the gradient of the ceiling in the lower Garden of the Gods. In addition, the prominent bedrock shelves above the stairs of the Spiral Stairway Shaft suggest this effect as well. Thus, the horizontal component of the water flow was determined by the strike of the bedding planes, but the vertical component may have been caused by the angle of intersection of joints with the beds.

Two other factors contributed to the steep gradient of the passages, by encouraging the great depth of the cave.

The steep dip of the beds allowed groundwater to descend well over 100 feet while following a single bedding plane. In addition, the great topographic relief caused by the cutting of the adjacent Jefferson River Canyon allowed deep flow through the subsurface routes.

The different levels of passages in Lewis and Clark Caverns probably formed at the same time. Individual levels of water table caves, such as the Mammoth Cave-Flint Ridge system, formed during different periods of time, determined by pauses in the entrenchment of the river. In Lewis and Clark Caverns, the large quantity of groundwater that flowed through the Spiral Stairway Shaft between the two strike levels indicates that both passages were active at the same time. In addition, no terrace levels have been found in the Jefferson Basin that would correlate with the three cave levels (Reshkin, 1963).

Mixing Corrosion

The Cathedral Room and Paradise Room appear anomalously large, compared to the strike passages that carried water into them. Mixing corrosion likely caused greater solution in these rooms than was possible in the strike passages. The mixing effect also probably continued into the trunk passages, enlarging them greatly as well.

Cave Sedimentation

Introduction

Breakdown was encouraged by the steep dip of the Mission Canyon Limestone, and the high degree of fracturing from the folding of the rock. Most of the breakdown probably fell during the drainage of the cave. The smectite muds were deposited after, and possibly during the episode of breakdown, perhaps by flooding. The silts are the youngest clastic sediments, and washed through the caverns during the Recent by a small, vadose stream.

The speleothems are the major attraction of Lewis and Clark Caverns. The great quantity of fractures through the rock, as well as the convex shape of Cave Mountain, disperses modern groundwater through millions of small cracks, encouraging super-saturated solutions of calcite. The flow of winds throughout the cave greatly retards the accumulation of carbon dioxide, and allows the continued precipitation of calcite, forming speleothems.

Breakdown

The steep dip of limestone favors the breakdown process, because triangular pieces of the ceiling can easily slip out

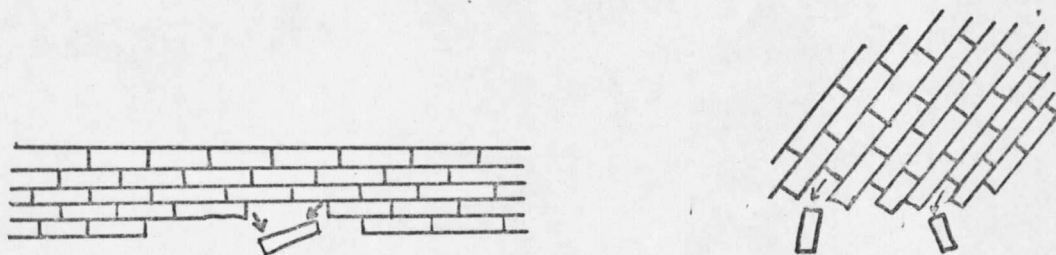


Fig. 36 Breakdown in horizontal limestone requires two weak joints for release of the block, but only one joint is needed in steeply-dipping beds. (Modified from White and White, 1969).

(Fig. 36) (White and White, 1969). Thus the steep dip of the Mission Canyon Limestone was important in the breakdown process at Lewis and Clark Caverns. The greatest volume of breakdown is seen in the joint-controlled areas, showing that fracturing of the limestone was important as well. The breakdown process did not create any known sinkholes, because the solution of the cave occurred deep inside of Cave Mountain. The breakdown process was terminated in most rooms of the cave by the formation of an arch-shaped ceiling, most notably in the Cathedral and Paradise Rooms (Fig. 25C1). The collapse in the Wind Tunnel and Bacon Room was more extensive, apparently due to greater fracturing at those sites. These

areas were stabilized when the breakdown blocks filled the room to the ceiling (Fig. 25C2).

The episode of breakdown probably took place during the drainage of the caverns. In the Wind Tunnel area, some breakdown blocks are coated on all surfaces with dogtooth spar, indicating that they fell into a supersaturated water. However, most blocks have no crystal coating, so moved later or did not fall into the water. In the Cathedral and Paradise Rooms, massive speleothems coat most of the breakdown blocks, thus the breakdown fell long ago. However, at least one block has fallen more recently, evidenced by stalactites hanging from its base at an angle of about 30 degrees from vertical.

Smectite Muds

The mechanism of deposition and the source for the smectite muds are unclear. The Mission Canyon Limestone has very small amounts of illite clay, but no smectite, so it is clearly not the source. Pockets of Tertiary Bozeman deposits were possibly preserved as recently as late Quaternary above Cave Mountain, supplying the smectite. However, no such deposits exist now. In addition, Laramide igneous dikes are common in the region, but none have been located above the

cave, to provide smectite as weathering products. A final possible source is the tephra found inside of the cave. The color and laminations of the smectite muds resembles the silt deposits. However, the distribution of the muds is very different from that of the silts. In addition, many of the ash shards found in the silt have not been devitrified, so complete weathering of this material to smectite is unlikely.

The muds coat many blocks of breakdown, and their laminae parallel the present position of the blocks, so the muds were deposited after the blocks assumed their present positions. This contrasts with the findings of Kuhns (in press), who believed that the muds were deposited before the breakdown. The very patchy distribution of the muds, rather than a blanket cover of the muds, is problematic (Fig. 33D). Flooding of the lower portion of the cave probably deposited the muds, but this does not explain its discontinuous exposures in breakdown areas.

The dessication of the muds has been a continuing process since the time of their deposition. Some of the muds remain damp and unlithified, thus the profound period of dessication in the caverns history proposed by Kuhns (in press) is not justified.

Vadose Silt

The vadose silt was partly supplied by Recent ash deposits. Several different horizons of ash can be seen in the pits dug in the back of the Cathedral Room, and possibly represent several different eruptions. The ash has not been positively identified, but preliminary work shows that it is probably not from the Rainier or Glacier Peak eruptions (Wilcox, 1979, written commun.). Additional work on the identity of the tephra in the caverns is needed. The shards are quite fresh, and appear to be Recent. This age is confirmed by the presence of the Recent deer legbone found mixed with the sediments. Phlogopite flakes found in the silt probably came from a local dike. This postulated dike may have also supplied the smectite muds to the cave.

The silt appears to be younger than the smectite muds, because the silt coats the lithified muds in the floor of the Paradise Room. The silts are also younger than the breakdown. Large columns and stalagmites dammed the silt where it entered the Cathedral Room, causing the four foot depth of sediments there. These and other large speleothems coat the breakdown blocks. Large speleothems also coat the smectite muds (Fig. 33D), but only minor speleothem growth has covered the silts. Thus, the breakdown and muds are

probably much older than the silts.

Speleothems

Vadose dripstone and flowstone cover many of the walls of the caverns, thus burying any earlier phreatic deposits. However, dogtooth spar can be found in many locations, especially in the Wind Tunnel. This shows that the groundwater lost its solutional aggressiveness, and deposited calcite instead of dissolving it.

Lewis and Clark Caverns has an unusual wealth of vadose speleothems, especially compared to other western caves (Campbell, 1978). Only two other caves in Montana have helictites (Campbell, 1978). Large stalagmites (Fig. 33F) testify to fast, perhaps sporadic water flow. The millions of helictites (Fig. 33E) and shields record very slow flow through tiny cracks and pores.

Relatively slow flow through the limestone above the cave is probably an important reason for the abundance of speleothems. The convex shape of Cave Mountain contributes to this slow seepage, because the topography disperses the water through millions of routes. Many famous caves that are also heavily decorated share this same topographic expression (Palmer, 1975). Other caves, such as Mammoth Cave

in Kentucky, are poorly decorated, because they are fed by sinkholes (Miotke and Palmer, 1972). Rapid recharge of water to the cave, through the large, efficient sinkholes, keeps the water acidic, thus preventing the deposition of speleothems. Folding of Cave Mountain during the Laramide provided millions of joints that dispersed the flow into the caverns. Finally, the thick cover of Cave Mountain encouraged the saturation of incoming water. Passages nearest the surface, such as the entrance, the Fish Tunnel, and the Nasal Passages have relatively few speleothems.

Most caves, however, have speleothems near their entrance, where carbon dioxide can escape from the cave and very few deep inside (Moore and Nicholas, 1964). The winds that flow through Lewis and Clark Caverns are apparently strong enough to remove the carbon dioxide that accumulates during the deposition process. Hence, the most decorated portions occur in the deepest portions of Lewis and Clark Caverns, most notably, the Paradise Room area.

Geomorphic Controls

Climate

Karst solution requires at least 10-12 inches of precipitation per year (Sweeting, 1973). No measurements have

been taken at the caverns site, but regional data suggests that the rainfall averages over 15 inches per year. The amount of rainfall in the Three Forks Basin depends on geographic location, varying from 11 inches near the center of the valley to 22 inches against the front of the Bridger Mountains in one year (Hackett and others, 1960). Bozeman, near the mountain front, and 50 miles east of the caverns, averages about 18 inches per year. Although not as high as the Bridger Range, the London Hills also act as a topographic barrier to weather systems. The temperature and precipitation is noticeably different between the caverns area and the town of Three Forks, in the valley.

Thus, precipitation in the caverns area is currently enough for karst development, although the modern topography and precipitation could not account for the quantity of water required to fill and dissolve the caverns. Slightly greater precipitation, and a larger catchment area during the earlier Quaternary is reasonable, and probably required.

Relation to the Jefferson Canyon

Abundant evidence suggests that the caverns discharged into the Jefferson River Canyon. Three major Laramide faults surround Cave Mountain. Water could flow through the Mission

Canyon Limestone only to the west, towards the canyon, because three major Laramide faults surround Cave Mountain (Plate II). The Morrison fault to the east and the Cave fault to the north, displaced the Precambrian LaHood arkose against the Mission Canyon Limestone. The Jefferson Canyon fault to the south separates the Mission Canyon Limestone from the Jefferson River with a small anticline cored by the LaHood.

A layer of solution breccia probably limited cavern growth up-section, preventing the cave from reaching to the north. This same breccia forms the walls of the Jefferson Canyon above the Jefferson Canyon fault. The biosparite seen in the cave possibly crops out on these walls, but is well exposed in the Limespur Quarry area, suggesting that the cave probably discharged near Limespur. In addition, the strike passages of the cave converge towards the west, and the Wind Tunnel extends further to the west. The cave winds, probably matching those in the quarry, also supports the argument.

Age of the Caverns

The cave formed during the cutting of the Jefferson River Canyon, but fairly early in the episode. A more exact

date could be proven by finding the relative elevation of the unknown entrance to the cave in the canyon.

Campbell (1977) hinted that Lewis and Clark Caverns formed during the late Mississippian stage of karst development. He believed that the cave passages would be horizontal if the cave had formed during the Quaternary, forming along a horizontal water table. However, the passages are steep, thus he reasoned that they originally formed along horizontal water routes before the cave was tilted by forces of the Laramide orogeny. Ford (1971) found that this generalization is incorrect because cave passages form at steep as well as low angles, often due to structural control. In addition, Lewis and Clark Caverns formed at the base of the Mission Canyon Limestone. This requires a depth of solution of well over 1,000 feet, which is quite unreasonable for karst development on the low relief plain during the late Mississippian. The maximum reported depth of solution by the Mississippian episode is 200 feet (Sloss and Hamblin, 1942).

The Mission Canyon Limestone at Cave Mountain might have been exposed during part of the Tertiary by post-Laramide erosion. However, data suggests that the area was covered to an elevation of at least 5750 feet by the Ballard

gravels, as well as to 6,000-6,600 feet by Tertiary fill (Peterson, 1974; Fields, oral commun., 1979). The highest elevation of solutional features in the cave is at an elevation of 5725, thus below the level of these sediments. The deep erosion during the cutting of the Jefferson Canyon exposed the limestone at Cave Mountain, and encouraged underground water circulation.

Kuhns (in press) believed that the cave formed just before the canyon began to be cut. The requirement of an entrenching surface drainage makes this unreasonable (Thornbury, 1969), especially in light of the great depth of the cave. The cave probably did not form late in the episode of the canyon cutting, when the drainages began to assume the appearance we see today. Sheep Gulch separated the cave area from the high level to the north during late Quaternary (Fig. 14C, 7B), thus cutting off the supply of water to the cave. Early in the Quaternary, water could gather in this area and flow to the canyon, around or through Cave Mountain.

Thus the caverns formed during the early Quaternary, during the early stages of the entrenchment of the Jefferson River. Subsequent downcutting separated the cave from its recharge source, draining the caverns. Subsequent erosion produced the modern landscape and discouraged any significant water flow through the caverns.

CHAPTER 7

SUMMARY OF THE CENOZOIC HISTORY RELATING TO LEWIS AND CLARK CAVERNS, MONTANA

The Cenozoic Era opened with the Laramide orogeny. Broad, open folds imparted a steep dip to the strata of Cave Mountain. The layers were later folded into two adjacent, plunging folds, the Colter anticline and Potts syncline, by plastic deformation just before the brittle rupture of two major Laramide faults, the Cave Fault and Jefferson Canyon fault. Both fractures trend east-west, and appear to be related to the Precambrian Willow Creek fault.

Following the Laramide orogeny, streams reduced the mountains and filled the intervening basins. Many streams flowed from the Jefferson Basin to the Three Forks Basin across the London Hills area, covering most of the region with a thick blanket of sediments of the Tertiary Bozeman Group. The Starretts Ditch fault uplifted the western flank of the London Hills area during the late Cenozoic, with at least 1,000 feet of offset. Deposition filled the resulting half-graben and buried the uplift. A major river flowed across the area during the early Quaternary, depositing the Ballard gravels.

Shortly thereafter, the entire Missouri River system began to downcut its channels, producing nearly 20 major canyons in the northern Rocky Mountains. The 1400 foot deep Jefferson River Canyon was cut at this time, primarily by superposition, but by some antecedent arching as well. This deep entrenchment exposed the limestone at Cave Mountain to erosion, and allowed the rapid movement of groundwater through the limestone. Lewis and Clark Caverns formed during the early Quaternary, in the early stages of the cutting of the Jefferson River Canyon.

The massive bedding and great amount of soluble material of the Mission Canyon Limestone favored the solution of many caves throughout the western United States. The relatively insoluble Lodgepole Limestone stopped the downward flow of groundwater through the Mission Canyon Limestone, concentrating the flow and solution at the base of the Mission Canyon Limestone, where Lewis and Clark Caverns formed. Most passages followed the bedding planes, and met at the fold axis. Trunk passages formed at the axis along joints produced by the folding of the limestone. The caverns formed in the phreatic zone, with no apparent water table control. Significant vadose solution did not occur, but fine clastic sediments, including volcanic ash, were carried through a part of the cave by a small vadose stream. Vadose speleo-

thems formed in great abundance in response to dispersed water flow from the convex surface of Cave Mountain through many small fractures, and the removal of carbon dioxide by cave winds.

REFERENCES CITED

- Alexander, Roger G. Jr., 1955, Geology of the Whitehall area, Montana: Yellowstone-Bighorn Research Project, Contribution 195, 111 p.
- Becker, H.F., 1973, The York Ranch flora of the upper Ruby River Basin, southwestern Montana: *Paleontographica*, Band 141. Abt. B.
- Berry, G.W., 1943, Stratigraphy and structure at Three Forks, Montana: *Geol. Soc. America Bull.*, v. 54, p. 1-30.
- Blackstone, D.L., 1977, The overthrust belt salient of the Cordilleran fold belt western Wyoming-southeastern Idaho-northeastern Utah: 29th Annual Field Conference, Wyoming Geological Association Guidebook, p. 367-384.
- Bretz, J. Harlen, 1942, Vadose and phreatic features of limestone caverns: *Jour. Geology*, v. 50, no. 6, p. 675-811.
- Burfeind, Walter John, 1967, A gravity investigation of the Tobacco Root Mountains, Jefferson Basin, Boulder batholith, and adjacent areas of southwestern Montana (Doctoral thesis): Indiana University, Bloomington, Indiana, 146 p.
- Campbell, Newell P., 1977, Possible exhumed fossil caverns in the Madison Group (Mississippian) of the Northern Rocky Mountains: a discussion: *Bull. Nat. Spel. Soc.*, v. 39, no. 2, p. 43-54.
- , 1978, Caves of Montana, *Mont. Bur. Mines and Geol. Bull.* 105, 169 p.
- Chesnutt, Vernon K., 1908, written correspondence describing Morrison Cave to Frank Bond, Chief Clerk, General Land Office, 4 p.
- Conn, Herbert W., 1966, Barometric wind in Wind and Jewel Caves, South Dakota: *Bull. Nat. Spel. Soc.*, v. 28, no. 2, p. 55-69.
- Conn, Herb and Conn, Jan, 1977, The Jewel Cave adventure: Teaneck, New Jersey, Zephyrus Press, Inc., 239 p.
- Davies, William E., 1960, Origin of caves in folded limestone: *Bull. Nat. Spel. Soc.*, v. 22, p. 5-18.

- Davis, William M., 1930, Origin of limestone caverns: Bull. Geol. Soc. America, v. 41, p. 475-628.
- ✓ Davis, W.E., Kinoshita, W.T., and Robinson, G.D., 1965, Bouguer gravity, aeromagnetic, and generalized geologic map of the western part of the Three Forks Basin, Jefferson, Broadwater, Madison, and Gallatin counties, Montana: U.S. Geologic Survey Geophysical Investigations, Map GP-197.
- Deal, Dwight E., 1962, Geology of Jewel Cave National Monument, Custer County, South Dakota, with special reference to cavern formation in the Black Hills (Masters thesis): University of Wyoming, Laramie, Wyoming, 183 p.
- Ewers, Ralph O., 1972, A model for the development of subsurface drainage routes along bedding planes (Masters thesis): University of Cincinnati, Cincinnati, Ohio, 84 p.
- Ford, Derek, 1971, Geologic structure and a new explanation of limestone cavern genesis: Trans. Cave Research Group of Gr. Brit., v. 13, no. 2, p. 81-94.
- Ford, Derek, and Ewers, R.O., 1978, The development of limestone cave systems in the dimensions of length and depth: Can. Jour. Earth Sciences, v. 15, p. 1783-1798.
- Ford, T.D., 1976, The geology of caves in Ford and Cullingford, ed., The science of speleology: New York, Academic Press, p. 11-60.
- Ford, T.D., and Cullingford, C.H.D., ed., The science of speleology: New York, Academic Press, 593 p.
- Hackett, O.M., Visher, F.N., McMurtrey, R.G., and Steinhilber, W.L., 1960, Geology and groundwater resources of the Gallatin Valley, Gallatin County, Montana: U.S. Geol. Survey Water-Supply Paper 1482, 282p.
- Harrison, J.E., Griggs, A.B., Wells, J.D., 1974, Tectonic features of the Precambrian Belt Basin and their influence on post-Belt structures: U.S. Geol. Survey Prof. Paper 866, p. 1-15.
- Hill, Carol A., 1976, Cave minerals: Austin, Texas, Speleo Press, 137 p.

Horberg, Leland, 1949, Geomorphic history of the Carlsbad Caverns area, New Mexico: Jour. Geol., v. 57, p. 464-476.

Jagnow, David H., 1978, Geology and speleogenesis of Ogle Cave: Bull. Nat. Spel. Soc., v. 40, p. 7-18.

Jennings, J.N., 1971, Karst: Cambridge, Massachusetts, M.I.T. Press, 252 p.

✓ Kuenzi, Wilbur David, 1966, Tertiary stratigraphy in the Jefferson River Basin, Montana (Doctoral thesis): University of Montana, Missoula, Montana, 293 p.

—1974, Fluvial depositional systems in the Tertiary Boze-
man Group (Montana) and their paleogeomorphic implica-
tions: Geol. Soc. America, Abs. with Programs (Rocky
Mountain Sec.), v. 6, no. 5, p. 451.

✓ Kuenzi, W.D. and Fields, Robert W., 1971, Tertiary stratig-
raphy, structure, and geologic history, Jefferson Basin,
Montana: Geol. Soc. America Bull., v. 82, p. 3373-3394.

Kuenzi, Wilbur David and Richard, Benjamin H., 1969, Middle
Tertiary unconformity, North Boulder and Jefferson Ba-
sins, southwestern Montana: Geol. Soc. America Bull.,
v. 80, p. 315-320.

✓ Kuhns, Roger J., in press, The speleology of Lewis and Clark
Caverns State Park: Montana Bur. Mines and Geol.

✓ Lowell, W.R., 1957, Tertiary geologic history of the Rocky
Mountains in Montana, U.S.A.: Rept., 20th Int. Geol.
Congress, Proc. Section 5, p. 455-467.

Malott, C.A., 1937, The invasion theory of cavern develop-
ment, Abs., Geol. Soc. America, Proc., p. 323.

Martel, E.A., 1905, La Spéléologie au XX^e siècle (revue et
bibliographie des recherches souterraines de 1901 à
1905): Spelunca, v. VI, 81 p.

McMannis, W.J., 1963, LaHood Formation--a coarse facies of
the Belt Series in southwestern Montana: Geol. Soc.
America Bull., v. 74, p. 407-436.

- McMannis, William J., 1965, Resume of depositional and structural history of western Montana: Bulletin of the American Assoc. of Petroleum Geologists, v. 49, no. 11, p. 1801-1823.
- Miotke, Franz-Dieter, and Palmer, Arthur N., 1972, Genetic relationship between caves and landforms in the Mammoth Cave National Park area: Privately published, 69 p.
- Moore, George W., and Nicholas, Brother G., 1964, Speleology, the study of caves: Boston, D.C. Heath and Company, 120 p.
- Nelson, Warren L., 1962, A seismic study of North Boulder Valley and other selected areas, Jefferson and Madison counties, Montana (Masters thesis): Indiana University, Bloomington, Indiana, 32 p.
- Palmer, Arthur Nicholas, 1969, A hydrologic study of the Indiana karst (Doctoral thesis): Indiana University, Bloomington, Indiana, 181 p.
- 1975, The origin of maze caves: Bull. Nat. Spel. Soc., v. 37, p. 56-76.
- Pardee, J.T., 1950, Late Cenozoic block faulting in western Montana: Geol. Soc. America Bull., v. 61, p. 356-406.
- Parker, John S., 1961, A preliminary seismic investigation of Tertiary basin fill in the Jefferson Island Quadrangle, Montana (Masters thesis): Indiana University, Bloomington, Indiana, 35 p.
- Perry, E.S., 1946, Morrison Cave, Lewis and Clark Caverns State Park: Butte, Montana, T. Greenfield, Inc.
- 1950, Highlights of geology between Three Forks and Whitehall: Billings Geol. Soc., 1st Annual Field Conf., Guidebook, p. 56-59.
- Peterson, James Carl, 1974, Geology of the Sweetwater Canyon area and origin of the interbasinal canyons, southwestern Montana (Masters thesis): Western Michigan University, Kalamazoo, Michigan, 89 p.

- Petsch, Bruno, 1936, Geologic report of Morrison Cave State Park SP-3, Jefferson County, Montana: National Park Service, region II report, 19 p.
- Picknett, R.G., Bray, L.G., and Stenner, R.D., 1976, The chemistry of cave waters in Ford and Cullingford, ed., The science of speleology: New York, Academic Press, p. 213-266.
- Reshkin, Mark, 1963, Geomorphic history of the Jefferson Basin, Jefferson, Madison, and Silverbow counties, Montana (Doctoral thesis): Indiana University, Bloomington, Indiana, 146 p.
- Richard, Benjamin H., 1966, Geologic history of the intermontane basins of the Jefferson Island Quadrangle, Montana (Doctoral thesis): Indiana University, Bloomington, Indiana, 73 p.
- Roberts, A.E., 1966, Stratigraphy of the Madison Group near Livingston, Montana, and discussion of karst and solution breccia features: U.S. Geol. survey Prof. Paper 526B, 23 p.
- Robinson, G.D., 1959, The Disturbed Belt in the Sixteenmile Creek area, Montana: Billings Geol. Soc., 10th Ann. Field Conf., Guidebook, p. 34-40.
- ✓ — 1963, Geology of the Three Forks Quadrangle, Montana: U.S. Geol. Survey Prof. Paper 370, 143 P.
- Roy, William R., and Hall, Robert D., 1979, Glacial geology of the South Boulder River Valley, Tobacco Root Mountains, Montana: Geol. Soc. of America, Abs. with Programs, (Rocky Mountain Section) p. 300.
- Sando, William J., and Dutro, J.T., 1974, Type sections of the Madison Group (Mississippian) and its subdivisions in Montana: U.S. Geol. Survey Prof. Paper 842, 22 p.
- Schmidt, Christopher J., 1975, An analysis of folding and faulting in the Northern Tobacco Root Mountains, southwest Montana (Doctoral thesis): Indiana University, Bloomington, Indiana, 480 p.

- Schmidt, Christopher J., 1976, Structural development of the Lewis and Clark Cavern State Park area, southwest, Montana: Tobacco Root Geological Society Guidebook, p. 140-150.
- Sloss, L.L. and Hamblin, R.H., 1942, Stratigraphy and insoluble residues of the Madison Group (Mississippian) of Montana: American Assoc. Pet. Geol. Bull., v. 26, p. 305-335.
- Smith, D.I., Atkinson, T.C., and Drew, D.P., 1976, The hydrology of limestone terrains in Ford and Cullingford, ed., The science of speleology: New York, Academic Press, p. 179-212.
- Smith, D.L., 1972, Depositional cycles of the Lodgepole Formation in central Montana: Mont. Geol. Society, 21st Annual Geol. Conf., v. 1, p. 29-35.
- Sprouse, Peter S., 1978, Longest, deepest caves: Nat. Speleological Society News, p. 99-101.
- Sweeting, Marjorie M., 1973, Karst landforms: New York, Columbia University Press, 362 p.
- Swinnerton, A.C., 1932, Origin of limestone caverns: Geol. Soc. America Bull., v. 43, p. 622-693.
- Thornbury, William D., 1965, Regional geomorphology of the United States: New York, John Wiley and Sons, 609 p.
- 1969, Principles of geomorphology: New York, John Wiley and Sons, 594 p.
- Thrailkill, John, 1968, Chemical and hydrologic factors in the excavation of limestone caves: Geol. Soc. America Bull., v. 79, p. 19-46.
- Tuttle, Merlin D. and Stevenson, Diane E., 1977, Variations in the cave environment and its biological implications: National Cave Management Symposium, Proc., p. 108-121.
- Wegemann, Carroll H., 1936, Report on Morrison Cave State Park, SP-3, Montana: National Park Service Region II Report, 8 p.

White, William B., 1960, Termination of passages in Appalachian caves as evidence for a shallow phreatic origin: Bull. Nat. Spel. Soc., v. 22, p. 43-53.

———1976, Cave minerals and speleothems in Ford and Cullingford, ed., The science of speleology: New York, Academic Press, p. 267-328.

White, William B., and Van Gundy, James J., 1974, Reconnaissance geology of Timpanogos Cave, Wasatch County, Utah: Bull. Nat. Spel. Soc., v. 36, p. 5-18.

White, William B., and White, Elizabeth L., 1968, Dynamics of sediment transport in limestone caves: Bull. Nat. Spel. Soc., v. 30, p. 115-129.

———1969, Processes of cavern breakdown: Bull. Nat. Spel. Soc., v. 31, p. 83-96.

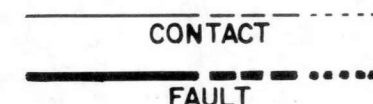
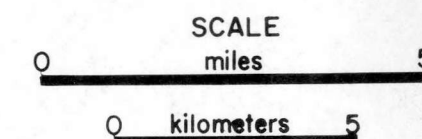
Wigley, T.M.L., and Brown, M.C., 1976, The physics of caves in Ford and Cullingford, The sciences of speleology: New York, Academic Press, p. 329-358.

Wilson, Daniel A., 1962, A seismic and gravity investigation of the North Boulder River and Jefferson River valleys, Madison and Jefferson counties, Montana (Masters thesis): Indiana University, Bloomington, Indiana, 43 p.



EXPLANATION

	QUATERNARY	Alluvium
	TERTIARY QUATERNARY	undifferentiated
	TERTIARY	Bozeman Group
	CRETACEOUS	volcanics intrusives
	CRETACEOUS JURASSIC	undifferentiated
	PERMIAN PENNSYLVANIAN MISSISSIPPIAN	undifferentiated
	DEVONIAN CAMBRIAN	undifferentiated
	PRECAMBRIAN	Belt Supergroup
	PRECAMBRIAN	metamorphics



dashed = approximately located
dotted = buried or inferred

② = location of Starretts Ditch fault (see text)



GEOLOGIC MAP of the area between the JEFFERSON and THREE FORKS BASINS

compiled from:
Alexander, 1955; Berry, 1943;
Davis and others, 1965; Kuenzi
and Fields, 1971; Perry, 1950;
Robinson, 1963; Schmidt, 1975

PLATE I

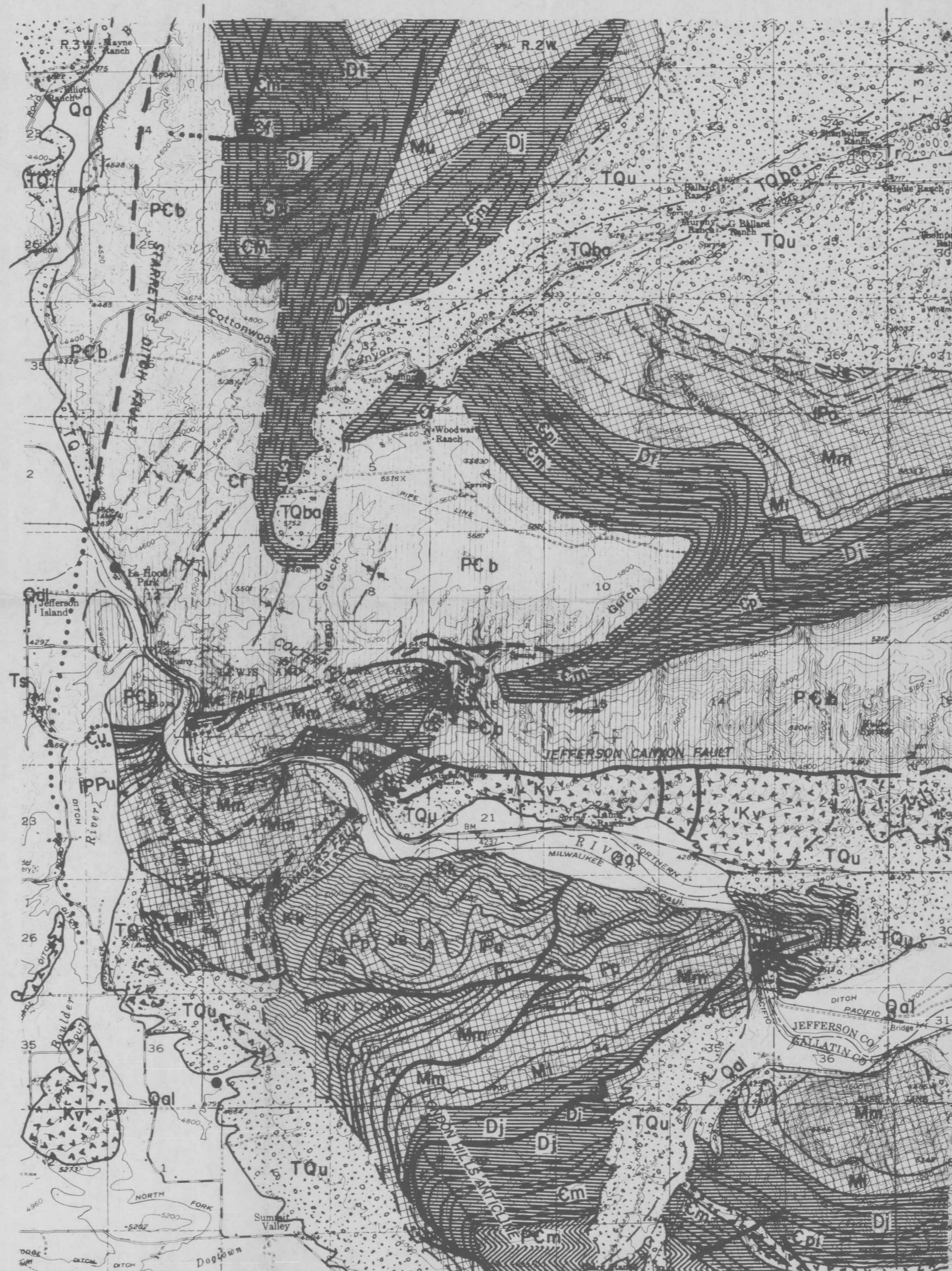
PLATE I

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GEOLOGIC MAP OF THE LONDON HILLS AREA

PLATE II



EXPLANATION

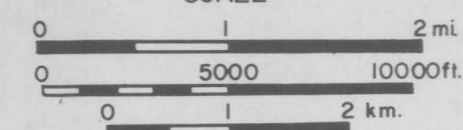
●	QUATERNARY	Holocene ash
□	QUATERNARY	Qa1 Holocene alluvium
□	TERTIARY QUATERNARY	TQu undifferentiated
□		TQu pediment gravels, loess, minor Bozeman Group outcrops
□		TQba Ballard gravels
□	TERTIARY	Tb Bozeman Group
□		Ts Six Mile Creek Fm.
□		Tr Renova Fm.
□	CRETACEOUS	Kv Elkhorn Mtn. Volcanics misc. intrusives
□	CRETACEOUS JURASSIC	Kk Kootenai Fm.
□		Jm Morrison Fm.
□		Je Ellis Fm.
□	PERMIAN PENNSYLVANIAN	Pp Phosphoria Fm.
□		IPa Quadrant Fm.
□		IPa Amsden Fm.
□	MISSISSIPPIAN	Mm Mission Canyon Fm.
□		MI Lodgepole Fm.
□	DEVONIAN	Dt Three Forks Fm.
□		Dj Jefferson Fm.
□		Dm Maywood Fm.
□	CAMBRIAN	Cpi Pilgrim Fm.
□		Cp Park Fm.
□		Em Meagher Fm.
□		Ew Wolsey Fm.
□		Cf Flathead Fm.
□	PRECAMBRIAN	PCb Belt Supergroup
□		Empire Fm.
□		LaHood Fm.
□	PRECAMBRIAN	PCm metamorphics

CONTACT

FAULT

dashed- approximately located
dotted- buried or inferred

SCALE



STUDY AREA



Jefferson Island Quadrangle (USGS)
MONTANA

geology compiled from:
Schmidt, 1975; Berry, 1943



rba 1979

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PLATE II

SCHEMATIC DRAWING OF LEWIS AND CLARK CAVERNS

by Richard B. Aram

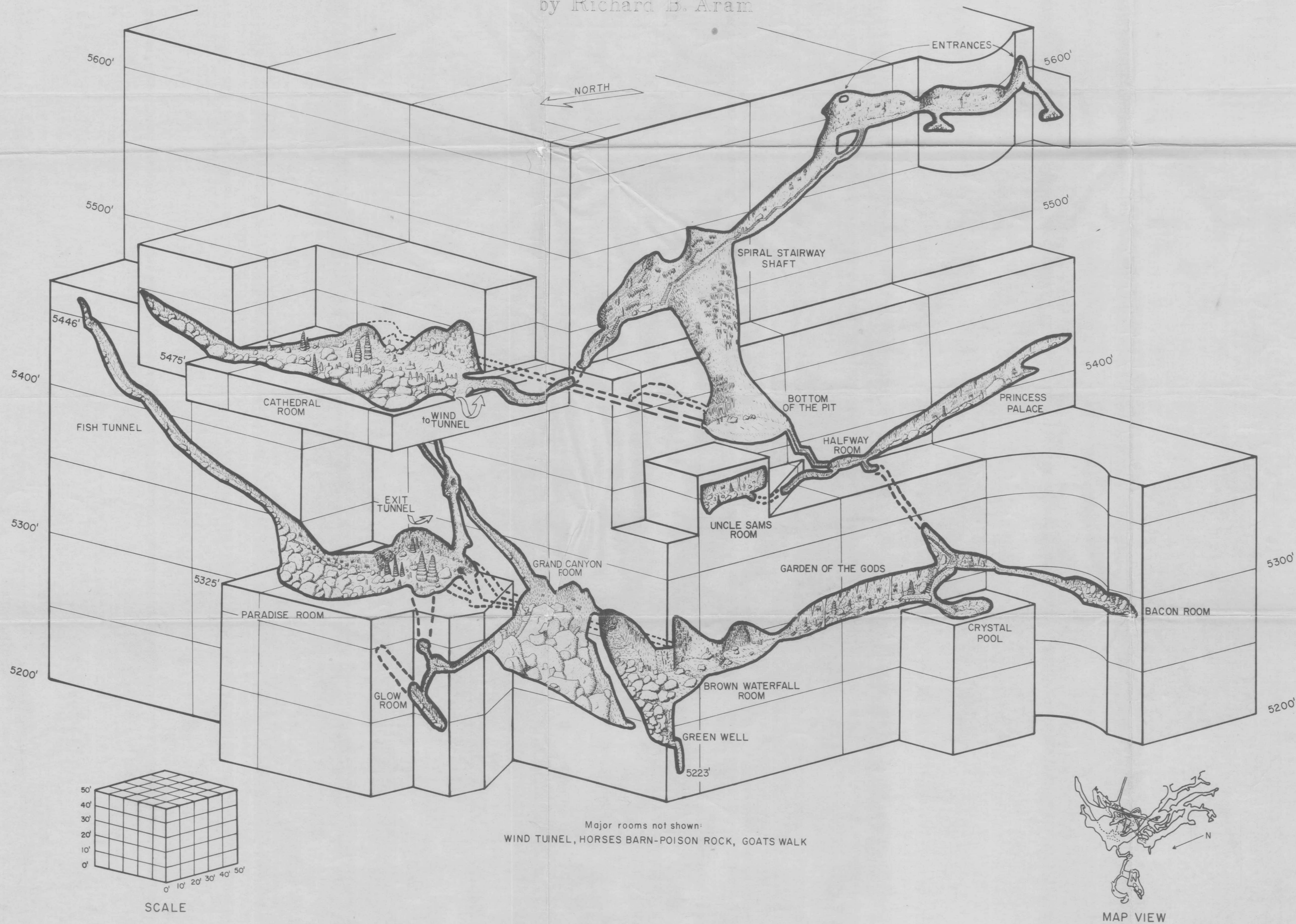


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