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Geologic controls on cave development in Burnsville Cove, Bath and Highland Counties, Virginia

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ABSTRACT

Burnsville Cove in Bath and Highland Counties (Virginia, USA) is a karst region in the Valley and Ridge Province of the Appalachian Mountains. The region contains many caves in Silurian to Devonian limestone, and is well suited for examining geologic controls on cave location and cave passage morphology. In Burnsville Cove, many caves are located preferentially near the axes of synclines and anticlines. For example, Butler Cave is an elongate cave where the trunk channel follows the axis of Sinking Creek syncline and most of the side passages follow joints at right angles to the syncline axis. In contrast, the Water Sinks Subway Cave, Owl Cave, and Helictite Cave have abundant maze patterns, and are located near the axis of Chestnut Ridge anticline. The maze patterns may be related to fact that the anticline axis is the site of the greatest amount of flexure, leading to more joints and (or) greater enlargement of joints. Many of the larger caves of Burnsville Cove (e.g., Breathing Cave, Butler Cave–Sinking Creek Cave System, lower parts of the Water Sinks Cave System) are developed in the Silurian Tonoloway Limestone, the stratigraphic unit with the greatest surface exposure in the area. Other caves are developed in the Silurian to Devonian Keyser Limestone of the Helderberg Group (e.g., Owl Cave, upper parts of the Water Sinks Cave System) and in the Devonian Shriver Chert and (or) Licking Creek Limestone of the Helderberg Group (e.g., Helictite Cave). Within the Tonoloway Limestone, the larger caves are developed in the lower member of the Tonoloway Limestone immediately below a bed of silica-cemented sandstone. In contrast, the larger caves in the Keyser Limestone are located preferentially in limestone beds containing stromatoporoid reefs, and some of the larger caves in the Licking

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Creek Limestone are located in beds of cherty limestone below the Devonian Oriskany Sandstone. Geologic controls on cave passage morphology include joints, bedding planes, and folds. The influence of joints results in tall and narrow cave passages, whereas the influence of bedding planes results in cave passages with flat ceilings and (or) floors. The influence of folds is less common, but a few cave passages follow fold axes and have distinctive arched ceilings.

INTRODUCTION AND BACKGROUND

This field trip provides an overview of the geologic controls on cave development in Burnsville Cove, an ~10-km-long and 3–5-km-wide karst region in Bath and Highland Counties, Virginia (Figs. 1 and 2). Burnsville Cove lies within the Valley and Ridge Province of the eastern United States (Fenneman and Johnson, 1946; Fenneman, 1975). The strata in Burnsville Cove are primarily Paleozoic limestone, sandstone, and shale (Figs. 3 and 4). Structural features include prominent anticlines, synclines, and thrust faults (Figs. 4 and 5). Most of the anticlines and synclines trend northeast, and are asymmetric with beds dipping more steeply on the western flanks. Thrust faults that dip to the southeast are present on the northwestern flanks of many of the major anticlines. Most of the structural features formed during the Alleghanian orogeny. During this event, which occurred ~325–270 million years ago, rocks along the eastern margin of North America were thrust from east to west and folded into the anticlines and synclines that are visible today both in outcrops and in the subsurface (e.g., Hatcher, 1989; Ryder et al., 2008).

Burnsville Cove forms the southwestern extension of the valley occupied by the Bullpasture River in Bath and Highland Counties, Virginia (Fig. 2). Burnsville Cove is bounded on the west by Jack Mountain, on the east by Tower Hill Mountain, on the south by Warm Springs Mountain, and on the north by the Bullpasture River. Chestnut Ridge is a north-trending ridge wholly within Burnsville Cove. With respect to structural setting, Burnsville Cove is located on the east flank of the Bolar anticline, a major northeast-trending anticline with an axis that lies between Little Mountain and Jack Mountain (Fig. 4). This anticline is part of the greater Hot Springs anticline of Butts (1933, 1940). Several parasitic anticlines and synclines are superimposed on the east flank of this major anticline (Fig. 5), including: (1) Sinking Creek syncline or Burnsville Cove syncline; (2) Chestnut Ridge anticline or Bullpasture Mountain anticline; (3) White Oak syncline; and (4) Tower Hill anticline. In the vicinity of Burnsville Cove, these parasitic anticlines and synclines plunge northeast. Most of these anticlines are asymmetrical, with steeper western limbs (nearly vertical to slightly overturned, compared to 20–35° dips for the shallow eastern limbs). Tower Hill anticline, however, is a more symmetrical box-fold in Bullpasture Gorge.

Joints are prominent structural features throughout Bath and Highland Counties, and most are oriented perpendicular to bedding planes and trends of folds. Joints are well exposed in outcrops of the Devonian Millboro Shale along the west side of State Route 678 ~4.0–4.8 km south of U.S. Route 250. The most prominent joint trend in Burnsville Cove is N50W (Deike, 1960b). Part of Mill Run (Fig. 2) follows this joint trend for ~1.6 km. Other joint trends are N40E and N60E. Joints are well exposed in caves of Burnsville Cove, and many of the cave passages are developed along the N50W joint trend.

The strata exposed in Burnsville Cove are primarily limestone, sandstone, and siliciclastic mudstone (shale) of Silurian and Devonian age. Although the stratigraphic nomenclature applied to these rocks has varied greatly over the years (e.g., Bick, 1962; Haynes, 2014; Swezey and Haynes, 2015), the most recent geologic map shows that the Silurian Tonoloway Limestone is the most widely exposed unit in Burnsville Cove (Fig. 4). Most caves are developed in the Tonoloway Limestone.

The locations of caves in Burnsville Cove are influenced both by structure and stratigraphy. In terms of structural setting, many of the larger caves are located along and near the axes of anticlines or synclines. In terms of stratigraphic setting, most of the caves in Burnsville Cove are located within the Silurian Tonoloway Limestone and (or) the Silurian to Devonian Helderberg Group. Nearly all of the larger caves are located in one of the three following stratigraphic intervals: (1) the lower (unnamed) member of the Silurian Tonoloway Limestone beneath a regionally extensive sandstone bed (Haynes, 2014; Swezey and Haynes, 2015); (2) a heterogeneous carbonate unit that is notable for its prominent stromatoporoid bioherms and biostromes, and is mapped as the Silurian Jersey Shore Limestone Member of the Keyser Limestone of the Helderberg Group (Cole et al., 2015; Swezey et al., 2015); and (3) chert-bearing strata of the Devonian Shriver Chert and Licking Creek Limestone of the Helderberg Group beneath the Devonian Oriskany Sandstone. In many of the caves, various sandstone beds (e.g., the Silurian Williamsport Sandstone, sandstone beds within the Tonoloway Limestone, and the Oriskany Sandstone) act as the lower and upper confining units that restrict caves to the intervening limestone.

In addition to influences on cave location, the structural and stratigraphic settings control significant aspects of cave passage morphology. Joints and folds influence passage direction, shape, and size, and in many instances, the overall cave pattern as well. Bedding planes also influence passage morphology at many locations. In general, "joint controlled passages" have greater height to width ratios than "bedding-plane controlled passages."

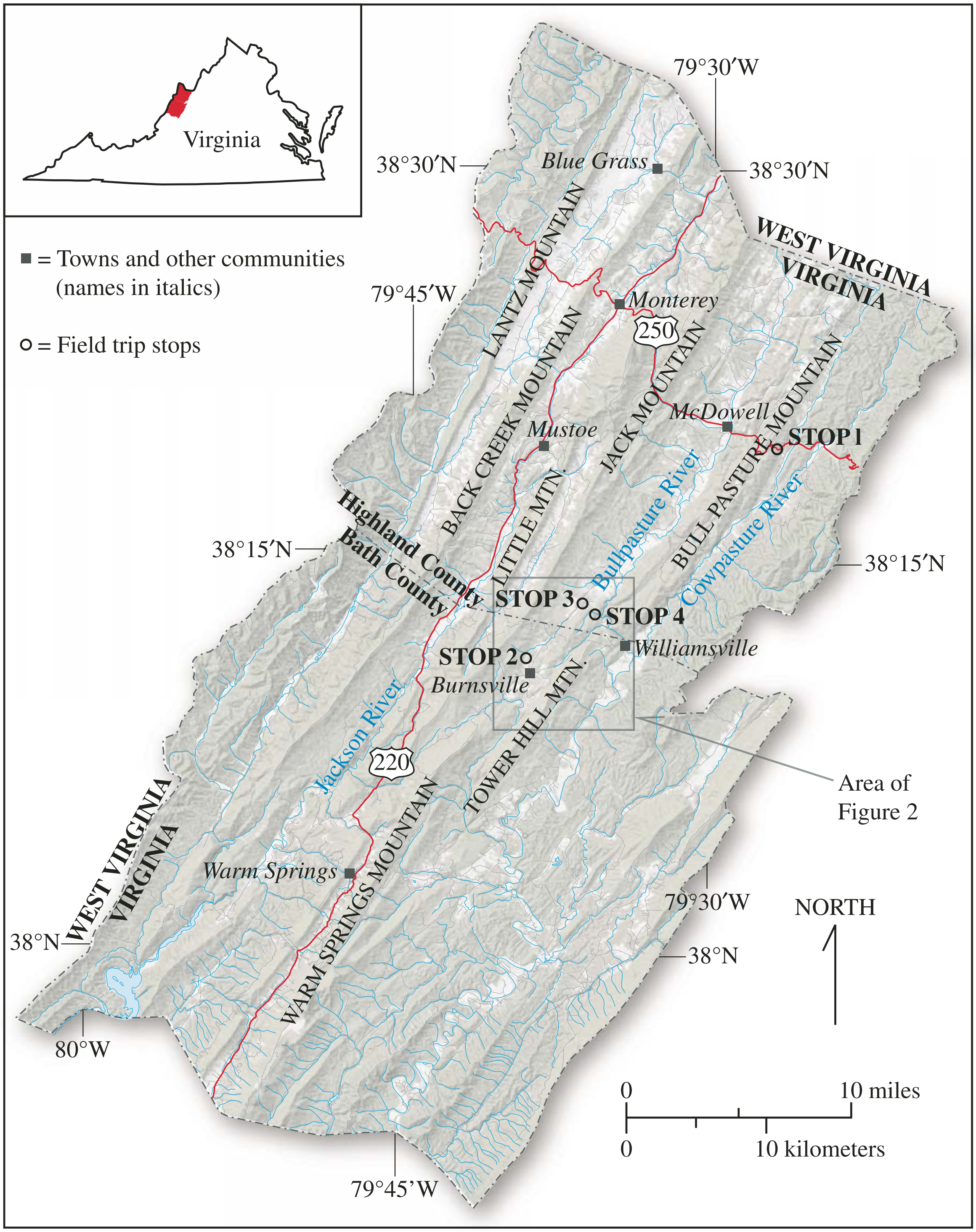


Figure 1. Map of Highland County and Bath County, Virginia (modified from Swezey et al., 2015). Black squares and names in italics denote towns and other communities. Blue denotes hydrological features. MTN.—mountain.

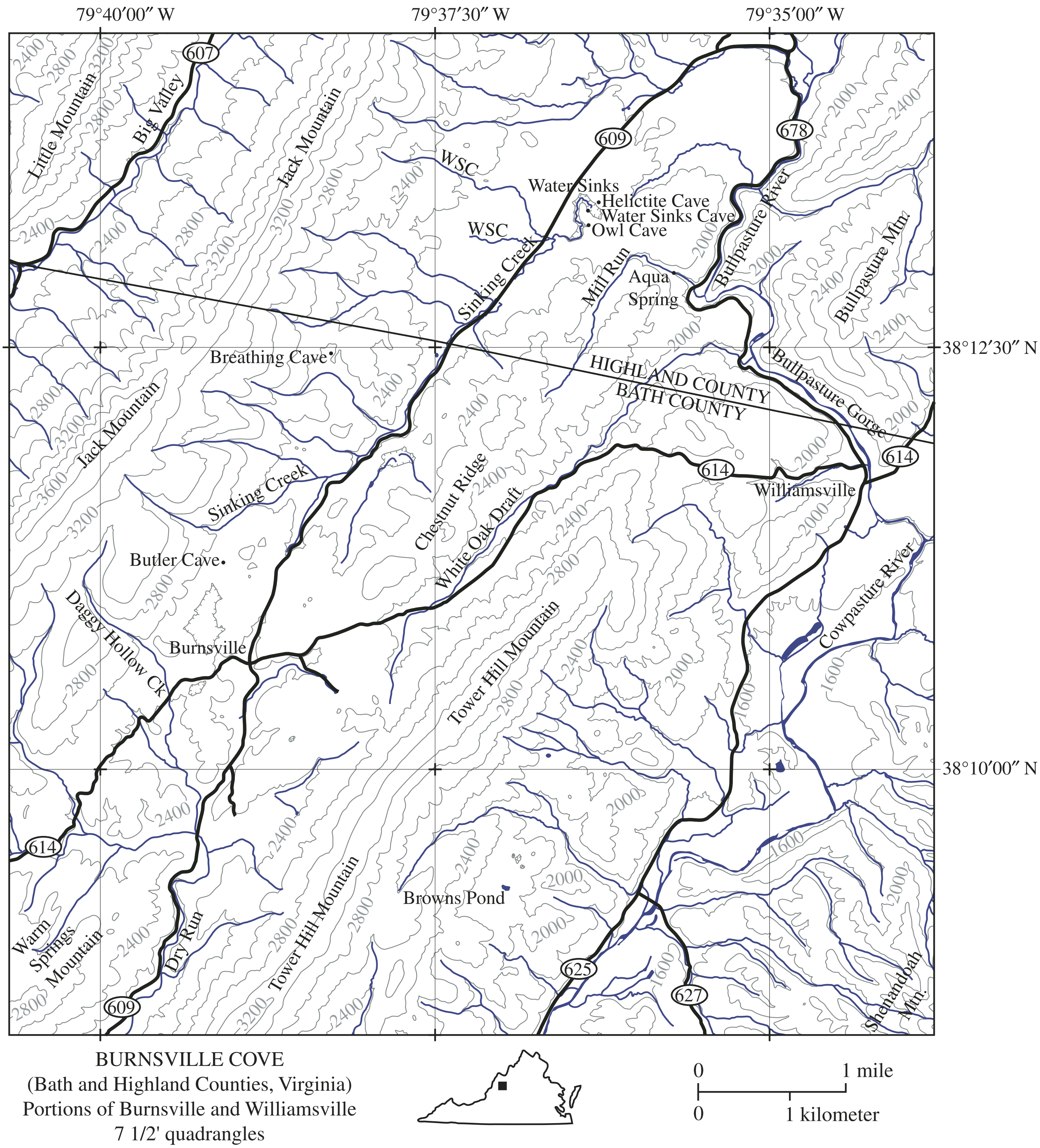


Figure 2. Topographic map of the Burnsville Cove area, Bath and Highland Counties, Virginia (modified from Swezey et al., 2015). Blue denotes hydrological features. WSC—Water Sinks Creek. Contour elevations are given in feet above mean sea level.

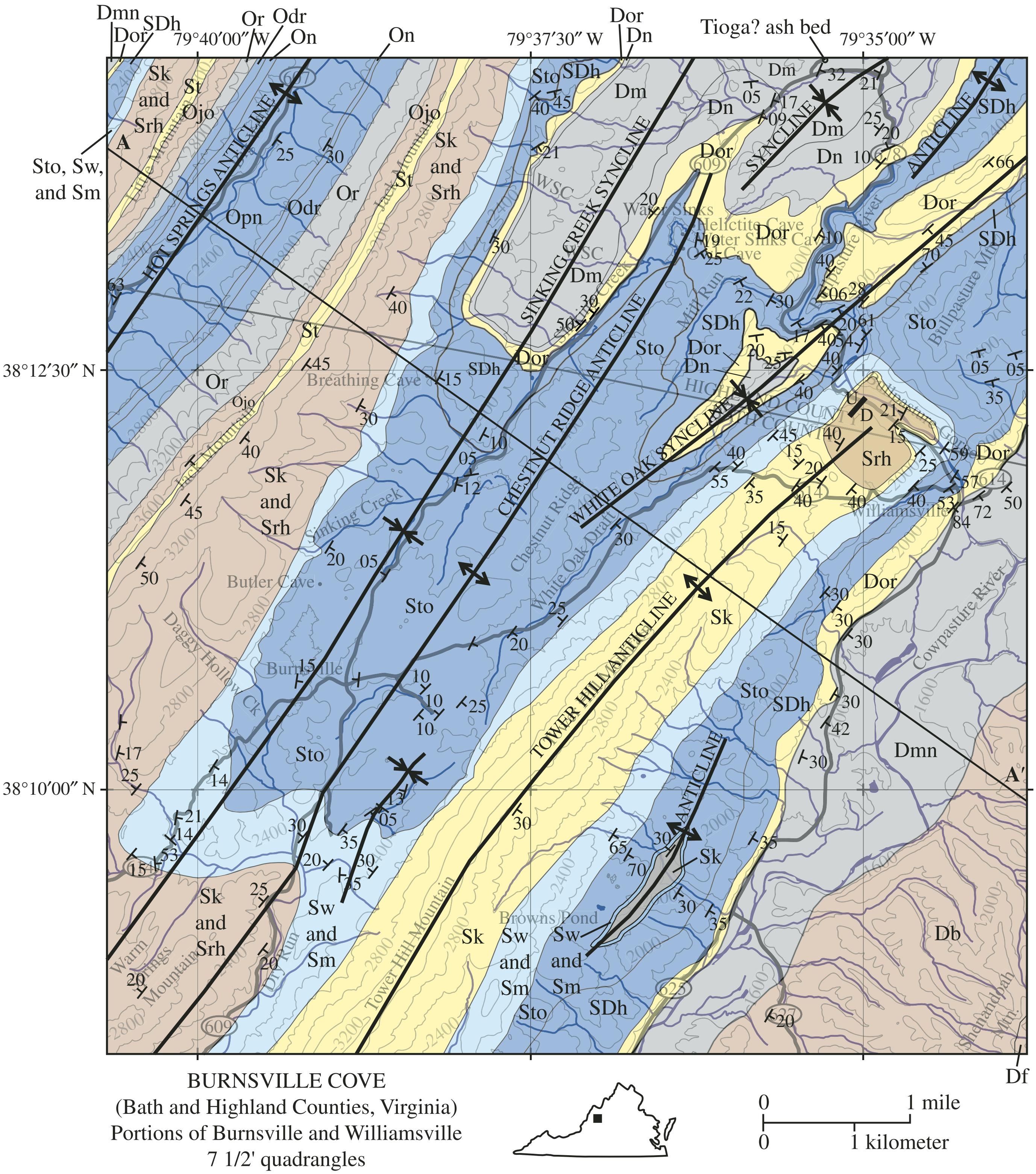


Figure 3. Geologic map of the Burnsville Cove area, Bath and Highland Counties, Virginia (modified and improved from a figure in Swezey et al., 2015). The map base is the same as Figure 2. Geologic map units are shown in Figure 4. Cross section A–A′ is shown in Figure 5.

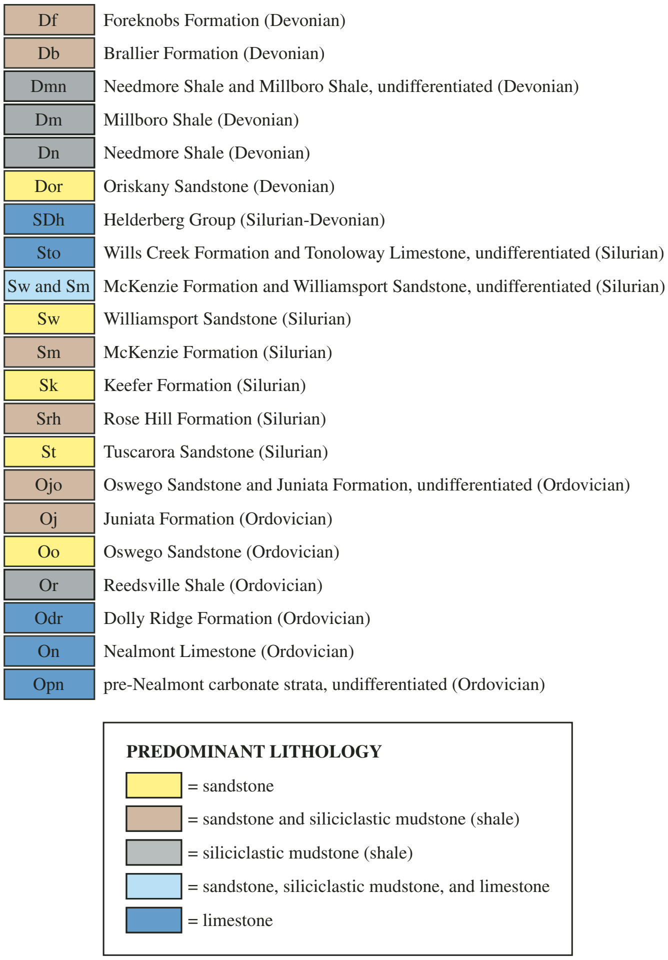


Figure 4. Geologic map units for Figure 3.

GEOLOGIC SETTING OF THE CAVES OF BURNSVILLE COVE

As of October 2013, there were at least 97 known caves in Burnsville Cove (White, 2015a). Prior to 1958, Breathing Cave was thought to be the largest cave in the area, and only a few other caves were known nearby (e.g., Owl Cave and Old Water Sinks Cave, which were known respectively as Siphon No. 1 Cave and Siphon No. 2 Cave). In 1956, diving at Aqua Spring (Stop 4) resulted in the discovery of Aqua Cave (Hewitt, 1956). Butler Cave (Stop 2) was discovered in 1958 (Nicholson, 1958). In 1961, dye traces demonstrated that Aqua Spring (the entrance to Aqua Cave) is the resurgence of water from Butler Cave (Holsinger, 1961). During the subsequent decades, many new caves were discovered in Burnsville Cove (Wefer and Nicholson, 1982; White, 2015a, 2015b, 2015c). Both Breathing Cave and Butler Cave are located on the west side of the Sinking Creek syncline (Fig. 4), but during the mid-1980s several new significant caves were discovered on and around the Chestnut Ridge anticline. Many of these caves were eventually connected to form the greater Chestnut Ridge Cave System (Clemmer, 2015), which has three entrances (Bobcat Cave, Blarney Stone Cave, Burns Cave). Since the mid-1990s, much exploration has focused around the northern end of Chestnut Ridge anticline in the Water Sinks area (Lucas, 2015a, 2015b), leading to the discovery of major additions to the Water Sinks Cave System (Stop 3), and the discovery of Helictite Cave (Stop 3).

There is a long history of study in Burnsville Cove on relations between geologic parameters, cave locations, and cave passage morphologies. Initial work on this topic by Deike (1959, 1960a, 1960b) was focused on Breathing Cave, and that seminal

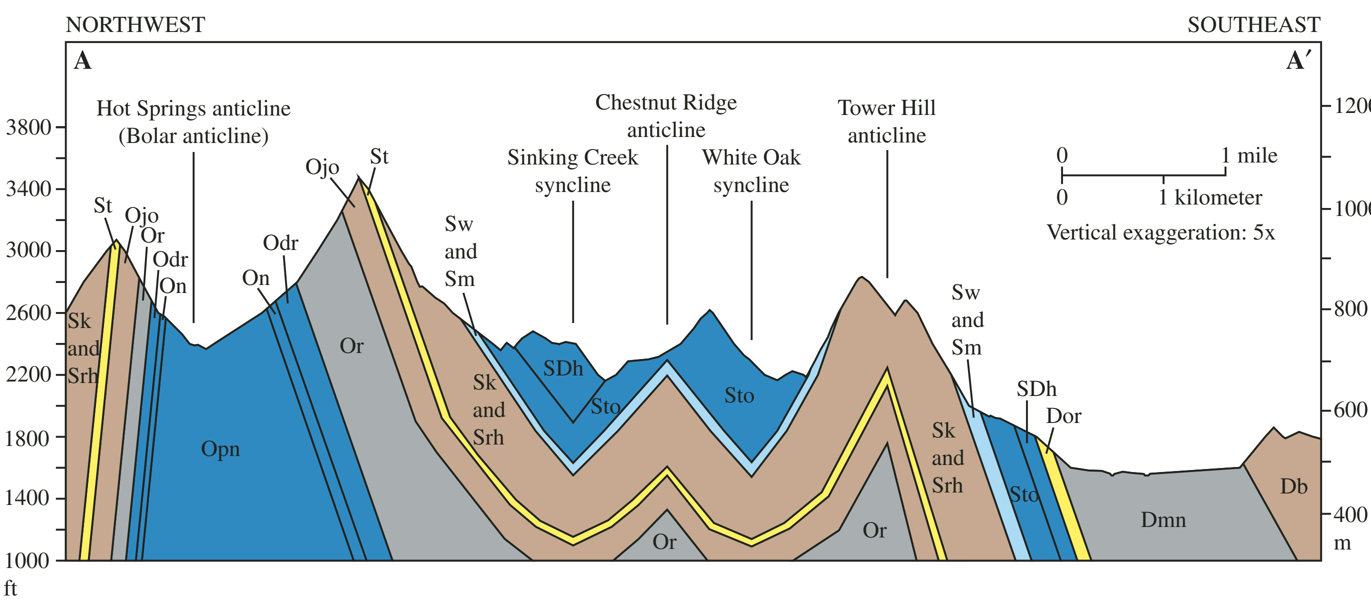


Figure 5. Geologic cross section A–A′ across Burnsville Cove (modified from Swezey et al., 2015). Cross section location is shown in Figure 3. Color code for lithologies, and formation names are shown in Figure 4. Elevation is relative to sea level.

work has been used for many later interpretations of the geologic setting of the caves (e.g., Hess and Davis 1969; White and Hess, 1982). More recently, the geological understanding of Breathing Cave has been updated and revised by Haynes (2014), and a book devoted exclusively to Burnsville Cove contains a revised understanding of the geology of Burnsville Cave as well as specific chapters on the geology of Breathing Cave, Butler Cave, the Chestnut Ridge Cave System, and certain caves of the northern part of Burnsville Cove (Deike, 2015; Swezey et al., 2015; White, 2015d, 2015e, 2015f).

Many of the caves in Burnsville Cove are located preferentially along the axes of synclines and (or) anticlines. For example, much of Butler Cave (Stop 2) is located on the west flank of Sinking Creek syncline, and the main trunk channel of the cave follows the syncline axis. Breathing Cave is also located on the west flank Sinking Creek syncline. Other caves follow the axis of White Oak syncline (e.g., Burns Cave, Blarney Stone Cave). Yet other caves are located on the crest of Chestnut Ridge anticline (e.g., the caves at Stop 3: Water Sinks Cave System, Owl Cave, Helictite Cave).

With respect to stratigraphic setting, many caves are located preferentially beneath sandstone beds. Deike (1960a, 1960b) noted that the ceiling and floor of Breathing Cave are formed by two 3.7-m (12-ft) thick beds of sandstone, which he informally named the "lower Breathing Cave sandstone" and the "upper Breathing Cave sandstone." Likewise, a sandstone bed forms the ceiling of much of Butler Cave (Stop 2). Petrographic studies have determined that the "upper Breathing Cave sandstone" and the Oriskany Sandstone (as well as several of the other Upper Silurian and Lower Devonian sandstone beds of this area) are calcarenaceous quartzarenite (Haynes, 2014), with "calcarenaceous" being defined as siliciclastic sediment in which 10% to 50% of the total framework grains are carbonate grains such as fossil fragments (Pettijohn et al., 1972). This mixture of quartz and carbonate grains is quite different from the Lower Silurian sandstone units (e.g., Tuscarora Sandstone, Rose Hill Formation, Keefer Formation, Williamsport Sandstone), which are typically quartzarenite where the quartz framework grains comprise ≥95% of the total framework grain population (Haynes et al., 2014). This distinction is of particular interest and significance because of its utility in distinguishing different sandstone beds.

Petrographic evidence and stratigraphic correlations (Haynes, 2014) suggest that the sandstone bed that forms the ceiling of many passages in Butler Cave correlates with the "lower Breathing Cave sandstone" of Deike (1960a, 1960b). Butler Cave was for decades thought to be developed in the lower part of the Silurian to Devonian Keyser Limestone of the Helderberg Group (Deike, 1960a; White and Hess, 1982). More recent studies, however, have concluded that both Butler Cave and Breathing Cave are located within the lower member of the Silurian Tonoloway Limestone (Haynes, 2014; Swezey et al., 2015). Furthermore, the work by Haynes (2014) shows that some cave passages that are developed in the lower member of the Tonoloway Limestone have eroded into the top bed of the underlying Williamsport Sandstone. Because of its relatively insoluble nature, the Williamsport Sandstone is effectively the "floor-rock" (versus caprock) for caves in the lower member of the Tonoloway Limestone throughout the area. Exposures of the Williamsport Sandstone are present in some of the lower passages of Butler Cave (Stop 2) and a few other caves in Burnsville Cove.

At least three major caves in Burnsville Cove have formed in the Jersey Shore Limestone Member of the Keyser Limestone, specifically within limestone beds that contain stromatoporoid reefs. These three caves are the Water Sinks Cave System (Stop 3), Owl Cave (Stop 3), and Aqua Cave (Stop 4). These caves are not associated with a sandstone caprock, and thus the upward development of cave passages is not limited by any obvious stratigraphic feature.

Collectively, the structural and stratigraphic setting of Burnsville Cove influences cave passage morphology. For example, many of the cave passages are tall and narrow, and follow joints. Other cave passages are relatively wide, and have flat ceilings and (or) floors. Such passages typically form where the cave ceiling and (or) floor follows a bedding plane. Yet other cave passages have arched ceilings where the passages follow folds in the strata. An excellent example of a cave passage following a fold in the strata is present in Butler Cave (Stop 2).

RELATIONS BETWEEN GEOMORPHOLOGICAL HISTORY AND CAVE BIOLOGY

The geomorphological history of Burnsville Cove may have influenced the faunal diversity of the caves, which are characterized by a relatively impoverished fauna when compared to caves in other river basins of Virginia and West Virginia (Holsinger, 1982; Hubbard, 1994–1995; Holsinger et al., 2013). The Burnsville Cove caves have both low species diversity and a low number of endemic species (species that are restricted to one cave system or a series of caves in a karst region). Studies by Holsinger (1982) have shown that the relative size of a cave system (length, as determined by cave surveys) does not necessarily correlate with species diversity, and that the low species diversity in the Burnsville Cove caves is primarily a function of geomorphologic and hydrologic isolation.

Despite this isolation, white-nose syndrome (WNS) arrived in Burnsville Cove, being detected first in Breathing Cave in 2009 (Dasher, 2009; Lambert, 2009) and later reported from nearby caves in 2011. WNS is a disease caused by the fungus *Pseudogymnoascus destructans*, which first appeared in North America in 2006, and since then has been spreading across North America and killing millions of bats (Gargas et al., 2009; Lorch et al., 2011; Swezey and Garrity, 2011; Minnis and Lindner, 2013; Reeder and Moore, 2013). **In order to minimize the risk of spreading WNS to other karst regions, field trip participants are asked to remove sediment/dirt from their clothes and gear upon exiting a cave. After this field trip, participants are asked to wash their clothes and gear in hot water and to avoid taking the clothing and gear used on this trip into other**

karst regions. Additional details about WNS decontamination protocols are available at www.whitenosesyndrome.org/topics/decontamination.

ROAD LOG AND STOP DESCRIPTIONS

There are four stops scheduled for this field trip (Fig. 1). The first two stops, which are scheduled for Day 1, are: (1) an outcrop along U.S. Route 250 that exposes much of the stratigraphy of the area; and (2) a visit to the property of the Butler Cave Conservation Society and Butler Cave. The third stop, which is scheduled for Day 2, is a visit to private property in order to see a complex blind valley, the Water Sinks Cave System, Owl Cave, and Helictite Cave. The fourth stop (also scheduled for Day 2) is located at Aqua Spring, which is both the entrance to Aqua Cave and the resurgence of water flowing through Butler Cave and the Water Sinks Cave System.

▪ DAY 1 (1 April 2017)

Drive from Richmond west to Highland County, Virginia.

After arriving in Highland County, the field trip begins at Stop 1 on the north side of U.S. Route 250 ~4 km east of the town of McDowell. Stop 1 is a series of outcrops that provide an excellent setting for an overview of the stratigraphy of the area. After Stop 1, the field trip proceeds south to Burnsville Cove and Stop 2 at the Homestead of the Butler Cave Conservation Society (BCCS). At Stop 2, the field trip visits Butler Cave to examine: (1) geologic controls on the cave location, and (2) geologic controls on cave passage morphology.

Stop 1: Outcrops on U.S. Route 250, Highland County, Virginia (N 38° 19′ 18.70″, W 79° 27′ 02.74″)

At Stop 1, which is located on the east flank of Bullpasture Mountain, there are very good outcrops of Silurian to Devonian strata from the top of the Silurian Williamsport Sandstone and going up-section through the Silurian Wills Creek Formation, the Silurian Tonoloway Limestone, the Silurian to Devonian Helderberg Group (including the Keyser Limestone, New Creek Limestone, Corriganville Limestone, Shriver Chert and Licking Creek Limestone), and the Devonian Oriskany Sandstone (Fig. 6). The primary goals at this stop are to examine and discuss key stratigraphic intervals, including: (1) the thin sandstone beds in the Tonoloway Limestone, and (2) the middle member of the Tonoloway Limestone and lithologically similar beds of the Keyser Limestone. This stop will include a discussion of the characteristics of the Keyser Limestone and the Tonoloway Limestone. Some of this discussion will focus on a vexing stratigraphic problem: Butts (1940) measured and described 595 ft (181 m) of Keyser Limestone at this outcrop, whereas Woodward (1941, 1943) measured and described only 155 ft (47 m) of Keyser Limestone here. Recent work in this area suggests that Woodward's measurements are correct (Haynes et al., 2014, 2015), and part of the discussion at this stop will focus on the resulting implications for the cave-forming limestone units to the south in Burnsville Cove.

Silurian Williamsport Sandstone

At Stop 1, the upper 4 m of the Williamsport Sandstone is exposed. This unit consists of beds of yellowish to yellowish-brown to grayish white sandstone (quartzarenite, with minor amounts of glauconite). The unit contains some cross-bedding and prominent ripple structures. Fossils include pyritized ostracod shells and horizontal trace fossils. The lower contact of the Williamsport Sandstone is the top of the Silurian McKenzie Formation (described in Swezey et al., 2015). The upper contact of the Williamsport Sandstone is placed at the top of a unit of brown to yellow-brown sandstone, just below a unit of brown to yellow to green shale, sandstone, conglomerate, sandy limestone, and limestone that is assigned to the Wills Creek Formation. An asymmetric anticline is present in this outcrop of the Williamsport Sandstone.

Silurian Wills Creek Formation

At Stop 1, the Wills Creek Formation is a 10-m-thick unit of brown to yellow to green shale, sandstone, conglomerate, sandy limestone, and limestone. The true thickness of this unit is uncertain because of minor faulting and folding in the outcrop. Some beds display ripple structures, laminations, desiccation cracks, conglomerate lenses composed of flat pebbles of carbonate mudstone, and (or) molds of evaporite crystals. Fossils are not abundant, but where present include leperditian ostracods, stromatolites, and brachiopods. The upper contact of the Wills Creek Formation is placed at the top of a unit of sandstone or shale, just below the lowest bed of gray to black thin-bedded to laminated limestone (carbonate mudstone) that is identified here as the base of the Tonoloway Limestone (Fig. 7).

Silurian Tonoloway Limestone

At Stop 1, the Tonoloway Limestone is a 152-m-thick unit of gray to blue limestone and dolomitic limestone, with some prominent beds of fine-grained to medium-grained sandstone and siliciclastic mudstone (shale). The limestone and dolomitic limestone are primarily thin-bedded to laminated carbonate mudstone, with some beds of bioclastic and oolitic packstone to grainstone. The sandstone beds are composed predominantly of quartz sand with some fossil fragments, and a clay matrix is present at some places. One of these sandstone beds is thought to form the ceiling of much of Butler Cave (Stop 2).

Studies of the Tonoloway Limestone in West Virginia and Maryland have subdivided the formation into three unnamed members (Woodward, 1941; Bell and Smosna, 1999; Haynes, 2014). The following descriptions of the three members from oldest to youngest apply to the field trip area as well as nearby regions:

(1) The lower member is a 65-m-thick unit of gray to black limestone (carbonate mudstone), argillaceous limestone,

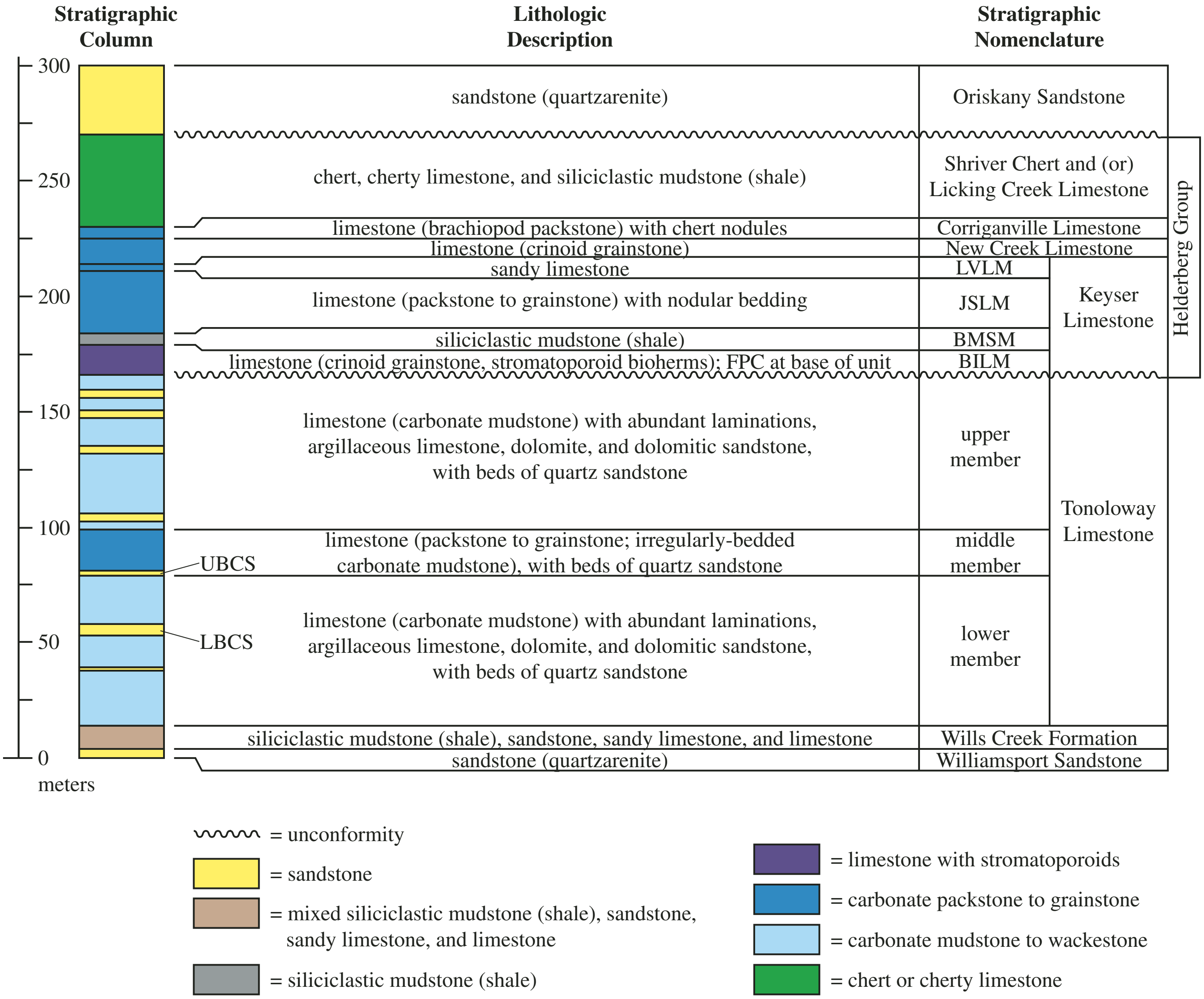


Figure 6. Silurian to Devonian strata exposed along north side of U.S. Route 250 on east side of Bullpasture Mountain, ~4 km (2.5 mi) east of the town of McDowell, Highland County, Virginia (modified from Swezey et al., 2015). BILM—Byers Island Limestone Member; BMSM—Big Mountain Shale Member; FPC—4–6-cm-thick flat-pebble conglomerate (conglomerate composed of flat pebbles of carbonate mudstone); JSLM—Jersey Shore Limestone Member; LBCS—lower Breathing Cave sandstone; LVLM—LaVale Limestone Member; UBCS—upper Breathing Cave sandstone.

dolomite, and dolomitic sandstone, with thin beds of quartz sandstone up to 4 m thick. This member is characterized by abundant thin beds and laminations (some of which are notably pink to red at some locations; Fig. 8), mud cracks, flat pebble conglomerate, and silicified evaporite nodules. Fossils include ostracods, gastropods, stromatolites, and *Tentaculites*.

(2) The middle member is a 20-m-thick unit of thick-bedded gray to black limestone (bioclastic packstone to grainstone), knobby or cobbly to irregularly-bedded blue to gray limestone (carbonate mudstone), and scattered beds of gray calcite-cemented sandstone. Fossils and fossil fragments include corals (including *Favosites* and *Halysites*), stromatoporoids, bryozoans, brachiopods, trilobites, and crinoid stems. In the vicinity of Burnsville Cove, the base of the middle member of the Tonoloway Limestone is the base of a regionally extensive sandstone bed that is informally named the "upper Breathing Cave sandstone."

(3) The upper member is a 67-m-thick unit of limestone (carbonate mudstone), argillaceous limestone, dolomitic sandstone, and dolomite characterized by abundant thin beds and laminations, mud cracks, pseudomorphs of gypsum

Figure 7. Contact of the Silurian Wills Creek Formation (left) and the overlying Silurian Tonoloway Limestone (right) on U.S. Route 250, Highland County, Virginia. Photograph by J.T. Haynes. From left to right, the people in the photograph are James Madison University (JMU) geology student Selina Cole, JMU geology student Kyle Hazelwood, and Richard (Rick) Lambert.

crystals now composed of calcite, and silicified evaporite nodules. Fossils include rare ostracods and *Tentaculites*.

Although the three members of the Tonoloway Limestone are distinct and well-exposed at Stop 1, the lower contact of the Tonoloway Limestone displays some geographic variability and may be difficult to identify at other locations (Fig. 9). At Stop 1, the lowermost bed of the Tonoloway Limestone rests directly on the Wills Creek Formation. However, farther south in Burnsville Cove and in the nearby gorge of the Bullpasture River, the Wills Creek Formation is thin or absent and the Tonoloway Limestone rests directly on the Williamsport Sandstone (Haynes, 2014; Haynes et al., 2015).

Figure 8. Red partings between beds of the lower member of the Tonoloway Limestone on U.S. Route 250, Highland County, Virginia. Photograph by J.T. Haynes.

As with the lower contact, the upper contact of the Tonoloway Limestone also displays some geographic variability. At Stop 1, the upper contact of the Tonoloway Limestone is placed at the top of a unit of gray laminated limestone, just below a unit of thick-bedded to medium-bedded massive to nodular limestone that is assigned to the Byers Island Limestone Member of the Keyser Limestone (Fig. 10). Farther south in Burnsville Cove at the Water Sinks Cave System (Stop 3), however, the uppermost bed of the Tonoloway Limestone is a 25–30-cm-thick carbonate boundstone with prominent "cabbage-head" stromatolites. This bed is capped by an unconformity, above which lies a 1–140-cm-thick conglomerate (composed of flat pebbles of carbonate mudstone), which is overlain by a 10-m-thick cross-bedded sandstone. The conglomerate and overlying sandstone are assigned collectively to the Clifton Forge Sandstone Member of the Keyser Limestone (Fig. 9).

Silurian to Devonian Helderberg Group

At Stop 1, the Helderberg Group is a 104-m-thick unit of predominantly limestone, with beds of siliciclastic mudstone (shale) and cherty limestone. The Helderberg Group is divided into the following four formations (from base to top): (1) Silurian to Devonian Keyser Limestone; (2) Devonian New Creek Limestone; (3) Devonian Corriganville Limestone; and (4) Devonian Shriver Chert and correlative Licking Creek Limestone.

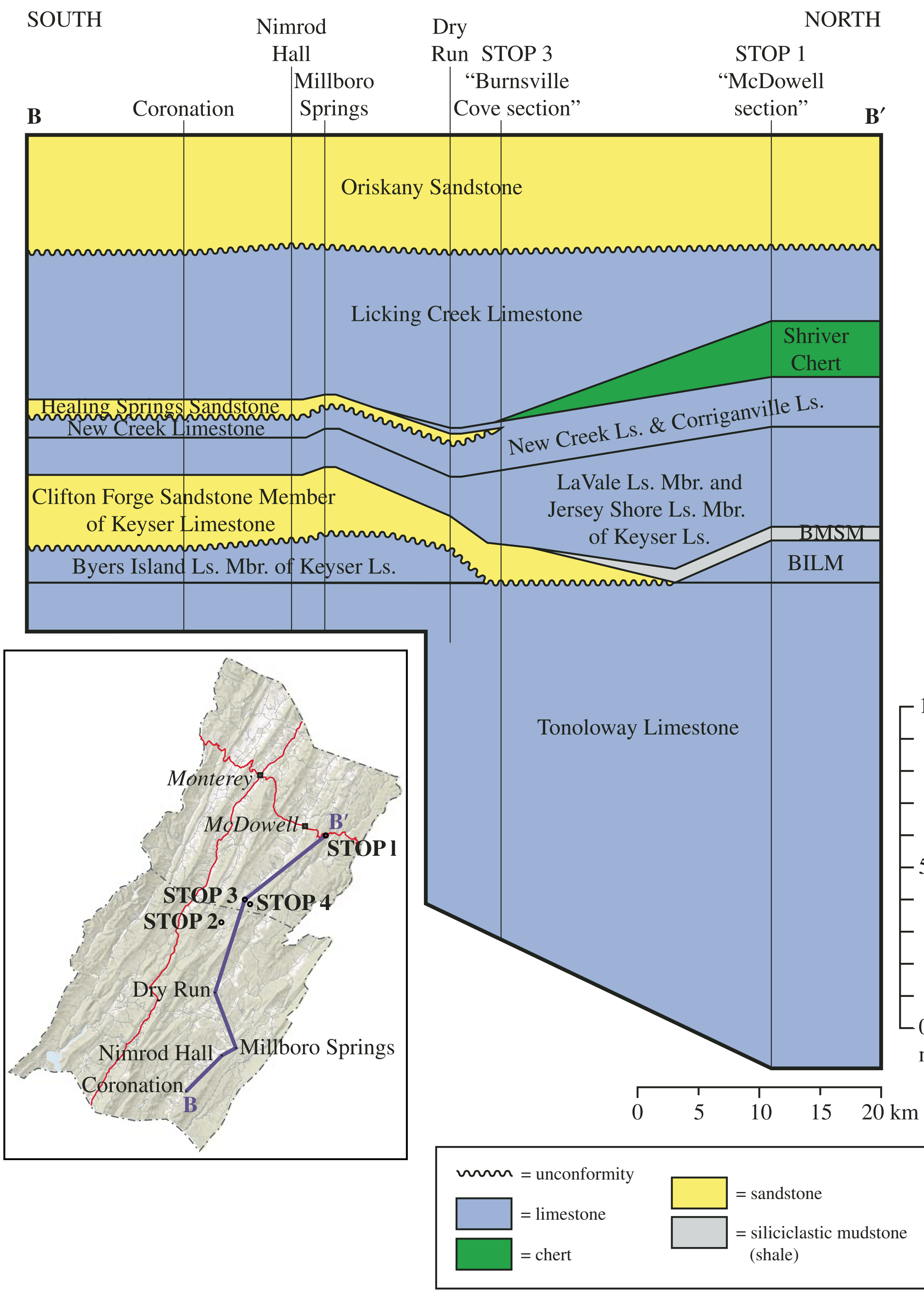


Figure 9. Geologic cross section B–B′ of selected Silurian to Devonian strata in Bath and Highland Counties, Virginia (modified from Swezey et al., 2015). Inset map that shows cross section location is the same area as the map in Figure 1. Section is hung on the top of the Tonoloway Limestone. BILM–Byers Island Limestone Member of Keyser Limestone; BMSM–Big Mountain Shale Member of Keyser Limestone.

Figure 10. Contact of the Silurian Tonoloway Limestone (left) and the overlying Silurian to Devonian Keyser Limestone (right) on U.S. Route 250, Highland County, Virginia. Photograph by J.T. Haynes.

Silurian to Devonian Keyser Limestone

At Stop 1, the Keyser Limestone is a 48-m-thick unit of predominantly limestone with some beds of sandstone and siliciclastic mudstone (shale). Many studies of the Keyser Limestone have divided the formation into a lower carbonate unit, a middle siliciclastic unit with two facies-related members (Big Mountain Shale Member, Clifton Forge Sandstone Member), and an upper carbonate unit (e.g., Swartz, 1930; Butts, 1933, 1940; Woodward, 1943; Bick, 1962; Dorobek and Read, 1986). Head (1972) named the lower carbonate unit the Byers Island Limestone Member, and he also subdivided the upper carbonate unit into the Jersey Shore Limestone Member and the overlying LaVale Limestone Member. The Byers Island Limestone Member, Big Mountain Shale Member, Jersey Shore Limestone Member, and LaVale Limestone Member are the four members of the Keyser Limestone identified and readily visible at Stop 1. The following descriptions of these four members from oldest to youngest apply to the field trip area as well as nearby regions.

(1) The Byers Island Limestone Member (lowermost member) is a 13-m-thick unit of gray to blue-gray crinoid grainstone with some low-angle cross bedding, and patches of stromatoporoid boundstone.
(2) The Big Mountain Shale Member is a 5-m-thick unit of fissile olive green siliciclastic mudstone (shale) and very thin beds of gray sandstone.
(3) The Jersey Shore Limestone Member is a 27-m-thick unit of gray to blue-gray bioclastic limestone (mostly packstone and grainstone), with lesser argillaceous limestone, and some beds that exhibit prominent nodular bedding.
(4) The LaVale Limestone Member (uppermost member) is a 3-m-thick unit of wavy laminated blue-gray to gray sandy limestone to calcite-cemented calcarenaceous sandstone.

The lower contact of the Keyser Limestone displays some striking geographic variability in the field trip area and in nearby areas. At Stop 1, the lower contact of the Keyser Limestone is placed at the base of a 4–6-cm-thick bed of flat-pebble conglomerate that is overlain by a crinoid grainstone with low-angle cross-bedding. This flat-pebble conglomerate and overlying crinoid grainstone, which are described in Figure 6 as "limestone (crinoid grainstone, stromatoporoid bioherms); FPC at base of unit," overlie the uppermost unit of gray laminated limestone of the Tonoloway Limestone. Farther south in the Water Sinks Cave System (Stop 3), however, the lower member of the Keyser Limestone (Byers Island Limestone Member) is absent, and the lower contact of the Keyser Limestone is placed at an unconformity at the base of a 1–140-cm-thick conglomerate (composed of flat pebbles of carbonate mudstone) overlain by a 10-m-thick cross-bedded sandstone that is mapped as the Clifton Forge Sandstone Member (Fig. 10).

The nature of the upper contact of the Keyser Limestone is more consistent across the region. At Stop 1, the upper contact of the Keyser Limestone is placed at the top of a 4-m-thick unit of dark gray, thin- to medium-bedded sandy limestone that has prominent laminations of quartz silt and sand (LaVale Limestone Member), and which underlies a unit of gray crinoid grainstone that is assigned to the New Creek Limestone. As seen at Stop 3, these stratigraphic relations characterize the upper contact of the Keyser Limestone in Burnsville Cove as well.

The Silurian to Devonian boundary is located within the upper Keyser Limestone (Denkler and Harris, 1988a, 1988b; Harris et al., 1994; Baez Rodríguez, 2005). According to Denkler and Harris (1988a, p. B8), "the Silurian-Devonian boundary is between 9 and 3.3 m below the top of the Keyser Limestone" in northeastern West Virginia. This statement suggests that the boundary is either in the uppermost Jersey Shore Limestone Member or in the LaVale Limestone Member of the Keyser Limestone.

Devonian New Creek Limestone

At Stop 1, the New Creek Limestone is an 11-m-thick unit of gray to pink limestone (grainstone) that consists of coarse-grained beds of pelmatozoan debris (primarily crinoids) and lesser amounts of brachiopods, bryozoans, and corals. The unit displays prominent cross-bedding in places. At Stop 1, the upper contact of the New Creek Limestone is placed at the top of a bed of gray limestone (crinoid grainstone), which lies immediately below a bed of gray limestone and cherty limestone (brachiopod packstone to wackestone) that is assigned to the Devonian Corriganville Limestone.

Bick (1962) mapped these strata as the Coeymans Limestone of the Helderberg Group. Bowen (1967), however, demonstrated that the Coeymans Limestone cannot be traced continuously from the type section in New York through intervening areas to Virginia, and so use of the name Coeymans Limestone has been discontinued in areas south of central Pennsylvania. Instead, the name New Creek Limestone is used for these correlative strata in areas south of central Pennsylvania (Bowen, 1967; Dorobek and Read, 1986).

Devonian Corriganville Limestone

At Stop 1, the Corriganville Limestone is not well exposed, but it is an ~5-m-thick unit of gray limestone with prominent light gray to white bedded chert. The limestone lithology is predominantly packstone. Throughout this region, the presence of light gray chert (bedded chert, chert nodules, and/or chert ribbons) is one of the more recognizable and distinguishing characteristics of the Corriganville Limestone. Fossils consist primarily of brachiopods (including *Macropleura* sp.) and corals. At Stop 1, the upper contact of the Corriganville Limestone is placed at the top of a unit of gray limestone and cherty limestone (brachiopod packstone to wackestone), which lies immediately beneath a unit of chert, cherty carbonate mudstone, and calcite-cemented siliciclastic mudstone (shale) that is assigned to the Shriver Chert of the Helderberg Group (or the upper Licking Creek Limestone of the Helderberg Group, according to Dorobek and Read, 1986). South of the Water Sinks depression (Stop 3), however, the upper contact of the Corriganville Limestone is placed at the top of a unit of gray limestone and cherty limestone (brachiopod packstone to wackestone), which lies immediately beneath a unit of gray limestone (with abundant nodules and ribbons of black chert) that is assigned to the Cherry Run Limestone Member of the Licking Creek Limestone of the Helderberg Group (Haynes et al., 2014).

Butts (1940), Woodward (1943), and Bick (1962) identified and mapped these strata as the New Scotland Limestone of the Helderberg Group. Head (1972), however, later demonstrated that the New Scotland Limestone cannot be traced continuously from the type section in New York through intervening areas to Virginia, and so use of the name New Scotland Limestone has been discontinued in areas south of central Pennsylvania. Instead, the name Corriganville Limestone is used for these correlative strata in areas south of central Pennsylvania (Head, 1972; Dorobek and Read, 1986).

In addition to the changes in nomenclature, there has been some confusion in previous literature about the thickness of this unit at this particular outcrop. Specifically, Butts (1940) and Woodward (1943) did not agree on the upper and lower contacts of this unit at this outcrop. Butts (1940) applied the name New Scotland Limestone to an ~33-m (100-ft) thick interval of strata, whereas Woodward (1943) applied the name New Scotland Limestone to an ~5-m (15-ft) thick interval of strata. Subsequent work by Haynes (2014) has shown that Butts (1940) picked the contacts incorrectly, and that the ~5 m (15 ft) thickness described by Woodward (1943) is correct.

Devonian Shriver Chert and Licking Creek Limestone

At Stop 1, there is a 40-m-thick unit of gray to black chert, cherty limestone (some with prominent pinkish partings), and siliciclastic mudstone (shale). This unit has been referred to as the Shriver Chert, the Licking Creek Limestone, and several other stratigraphic names. Beds range from massive to nodular to irregular with thin laminations. The upper contact of the unit is placed at the top of a unit of bedded chert, cherty carbonate mudstone, and calcite-cemented siliciclastic mudstone (shale), just below a unit of brown to yellow sandstone that is assigned to the Devonian Oriskany Sandstone. In some places, the Shriver Chert and Licking Creek Limestone are capped by an unconformity, above which lies the Oriskany Sandstone (Butts, 1940; Dennison, 1985; Dennison et al., 1992).

During the early twentieth century, there was recurring uncertainty and confusion about stratigraphic terms, and many of the early stratigraphic terms applied to these strata are no longer in use. Butts (1933, 1940) mapped these strata in Bath and Highland Counties simply as the Becraft Limestone Member of the Helderberg Limestone. Swartz (1930), however, mapped these strata at various locations in Bath and Highland Counties as the Becraft Limestone of the Helderberg Group, as the Shriver Chert of the Oriskany Group (e.g., near McDowell), and as the Shriver Chert of the Helderberg Group (e.g., Monterey and Back Creek Mountain west of Warm Springs). Woodward (1943) mapped these strata in some parts of Bath and Highland Counties as the Port Jervis Limestone and Chert, which he stated was equivalent in part to the Becraft Limestone and Shriver Chert of previous reports. Woodward (1943) also mapped these strata in other parts of Bath and Highland Counties as the Port Ewen Shale and Chert, which he stated was equivalent in part to the Shriver Chert of previous reports. Furthermore, Woodward (1943) stated that the Port Jervis Limestone at Monterey was previously mapped by F.M. Swartz as the Licking Creek Limestone (which is a unit that was named and defined at a type section in Pennsylvania by Swartz, 1939). Bick (1962) likewise mapped these strata in Bath and Highland Counties as the Licking Creek Limestone of the Helderberg Group.

Because of this confusion, Head (1974) redefined the Licking Creek Limestone as comprising all of the limestone and cherty limestone above the Corriganville Limestone and below the Oriskany Sandstone, and he stated that the Licking Creek Limestone passes laterally into cherty strata that are assigned to the Shriver Chert where chert is the predominant lithology. Head (1974) also designated the following two members of the Licking Creek Limestone: (1) a lower Cherry Run Limestone Member, which is described as dark gray to black fine- to medium-grained calcareous silty shale and shaly limestone with nodules and irregular beds of chert, interbedded with nodular and bedded fine-grained carbonate siltstone and sandstone; and (2) an upper Little Cove Limestone Member, which is described as gray medium- to thick-bedded fine to coarse calcarenite in which chert is generally a minor constituent.

Subsequent work by Dorobek and Read (1986) followed the formation nomenclature of Head (1974), but did not use the member nomenclature. On a regional cross section, they showed a predominantly limestone lithology and used the term Licking Creek Limestone in Bath County (Virginia), and they showed a predominantly chert lithology and used the term Shriver Chert in Pendleton County (West Virginia). In Highland County, their cross section B–B′ shows "Lower Licking Creek Limestone" at Strait Creek near the town of Monterey and "Shriver Chert" at the town of McDowell.

In summary, the Licking Creek Limestone and Shriver Chert are facies-related, correlative units of limestone and black chert that lie immediately above the Corriganville Limestone and below the Oriskany Sandstone. For the purpose of this field trip, the emphasis is placed on mapping at the scale of formations rather than members. Therefore, for all of the strata between the Corriganville Limestone and the Oriskany Sandstone, the term Licking Creek Limestone is used where limestone is more abundant than chert, and the term Shriver Chert is used where chert is more abundant than limestone.

Devonian Oriskany Sandstone

At Stop 1, the Oriskany Sandstone is a 30-m-thick unit of white to yellow to red-brown sandstone (quartzarenite) composed of coarse to very fine quartz sand. Fossils include common molds of the large brachiopod *Spirifer arenosus*. In most places the sand is cemented by calcite. The upper contact of the Oriskany Sandstone is the base of the Devonian Needmore Shale (described in Swezey et al., 2015).

In some publications, the Oriskany Sandstone has been designated as the Monterey Sandstone (e.g., Darton, 1899) or the Ridgely Sandstone (e.g., Schuchert et al., 1913). Butts (1940), however, stated that the Oriskany Sandstone is the same as the Monterey Sandstone and the Ridgely Sandstone. Because the name Oriskany Sandstone has precedence (Vanuxem, 1839), the rules of stratigraphic nomenclature (North American Commission on Stratigraphic Nomenclature, 1983; Salvador, 1994) dictate that that name Oriskany Sandstone should be given to these strata in Bath and Highland Counties.

Drive from U.S. Route 250 to the town of McDowell, and then south on U.S. Route 678 toward the community of Burnsville and the Homestead of the Butler Cave Conservation Society, Bath County, Virginia.

Stop 2: Homestead of the Butler Cave Conservation Society (N 38° 11′ 14.48″, W 79° 39′ 06.70″)

This is private property. Please obtain consent from the owners before proceeding onto the property.

Stop 2 is located on the property of the Butler Cave Conservation Society (BCCS). The BCCS is a non-profit society created in 1968 that is dedicated to the conservation, exploration, survey, preservation, and scientific study of caves in and around Burnsville Cove (Wefer, 1993; Wefer and Wheeland, 2015). The BCCS owns and (or) manages numerous caves in Bath and Highland Counties, including Butler Cave. The entrances to these caves are gated and locked, and coordinates of cave entrances are not given in this field guide in order to protect the caves from possible vandalism. Access to the caves owned and (or) managed by the BCCS may be obtained only via the BCCS (http://www.butlercave.org/). The BCCS also provides small grants for research on cave and karst science in and around Burnsville Cove (http://butlercave.org/science/sinkingcreekgrant.html).

Butler Cave (also known as the Butler Cave–Sinking Creek Cave System) is characteristic of many caves in Burnsville Cove because the cave is quite large but has a very small natural entrance. As of September 2016, Butler Cave was known to have a total passage length of 29.06 km (18.06 mi), and was the fifth longest cave in Virginia (www.caverbob.com/usalong.htm; accessed 28 September 2016). The cave was discovered in 1958 when Tommy Burns and Jimmy Puffenbarger showed Oscar Estes a small hole from which wind was blowing beneath a ledge in the hillside. Estes crawled into the hole a short distance and realized that it led to a much larger cave (Wefer and Nicholson, 1982; White, 2015b). Estes later recounted this discovery to Ike Nicholson, who then organized an exploration party. As the exploration party discovered, this small hole in the hillside leads to a narrow crawlway, which eventually gives access to a vast cave system. In 1998, a 1.5-m-diameter culvert was installed at another location to create a second entrance, which is known as the SOFA (Stubborn Old Farts Access) Entrance of Butler Cave. This field trip uses the SOFA entrance of the cave.

There are several levels to Butler Cave (Fig. 11). The uppermost level is restricted to the area around the original cave entrance. The next upper level is restricted to the southwest side of the cave system and includes the area known as Mbagintao Land. The middle level is the most extensive level of the cave system. The lower level includes an area near the cave entrances and an area on the northeast side of the cave system known as Marlboro Country.

Most of Butler Cave is located on the west side of Sinking Creek syncline, which trends northeast (Fig. 11). According to Deike (1960b), this syncline plunges northeast at ~4.6°. The trunk channel of Butler Cave generally follows the syncline axis, and many side passages follow northwest-trending joints that are approximately perpendicular to the syncline axis. Most of the known side passages, however, are restricted to the west side of the syncline axis. At a distance of ~2.6 km northeast of the Butler Cave entrance, Breathing Cave is also located on the west side of the syncline axis, and many of the Breathing Cave passages follow northwest-trending joints. This characteristic suggests that Breathing Cave might be simply another set of side passages of the greater Butler Cave–Sinking Creek Cave System. As of October 2016, however, a connection had yet to be found between the two caves, and the distal end of Breathing Cave was ~152 m from known passages of Butler Cave (White, 2015b).

With respect to stratigraphic setting, Butler Cave is located in the lower member of the Silurian Tonoloway Limestone. Previously Butler Cave was reported to be located in the Silurian to Devonian Keyser Limestone of the Helderberg Group (White and Hess, 1982), but subsequent studies have concluded that both Butler Cave and Breathing Cave are located in the lower

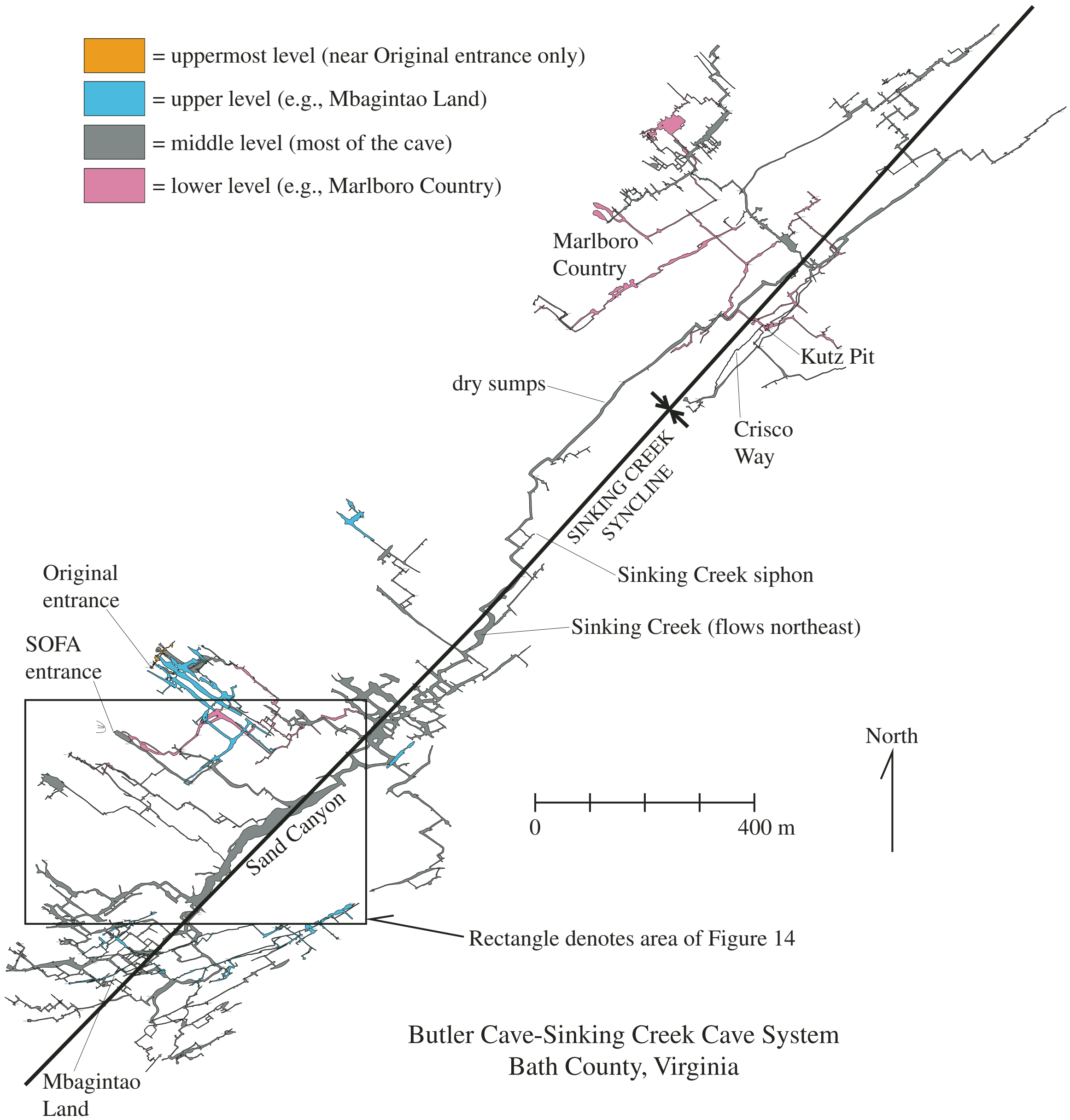


Figure 11. Map of Butler Cave (modified from White, 2015d) and selected geological data.

member of the Tonoloway Limestone (Haynes, 2014; Swezey et al., 2015). Within the lower member of the Tonoloway Limestone, two ~3.7-m-thick beds of sandstone are visible in many parts of Butler Cave. Much of Butler Cave is located immediately below the lower sandstone bed, although this sandstone bed is breached at several locations within the cave (Fig. 12). At the southern (upstream) end of the cave, the sandstone bed is breached at a location named Mbagintao Land, and beyond this breach the cave passages are developed in limestone above the lower sandstone bed. Likewise, to the north (downstream of Sand Canyon), the sandstone bed is breached by the trunk channel at a location named the “dry sumps” where the cave is restricted to one passage. This restriction of the cave passages is quite apparent in a plan view map of the cave (Fig. 11), and suggests that the sandstone bed has been a barrier to fluid flow. To the north of this breach, the cave again enters limestone above the lower sandstone bed and the cave map shows many intersecting passages. In this northern (downstream) portion of the cave, the sandstone bed is breached again at a few other locations (Kutz Pit, Crisco Way) and some cave passages known as Marlboro Country are located in limestone below the lower sandstone bed.

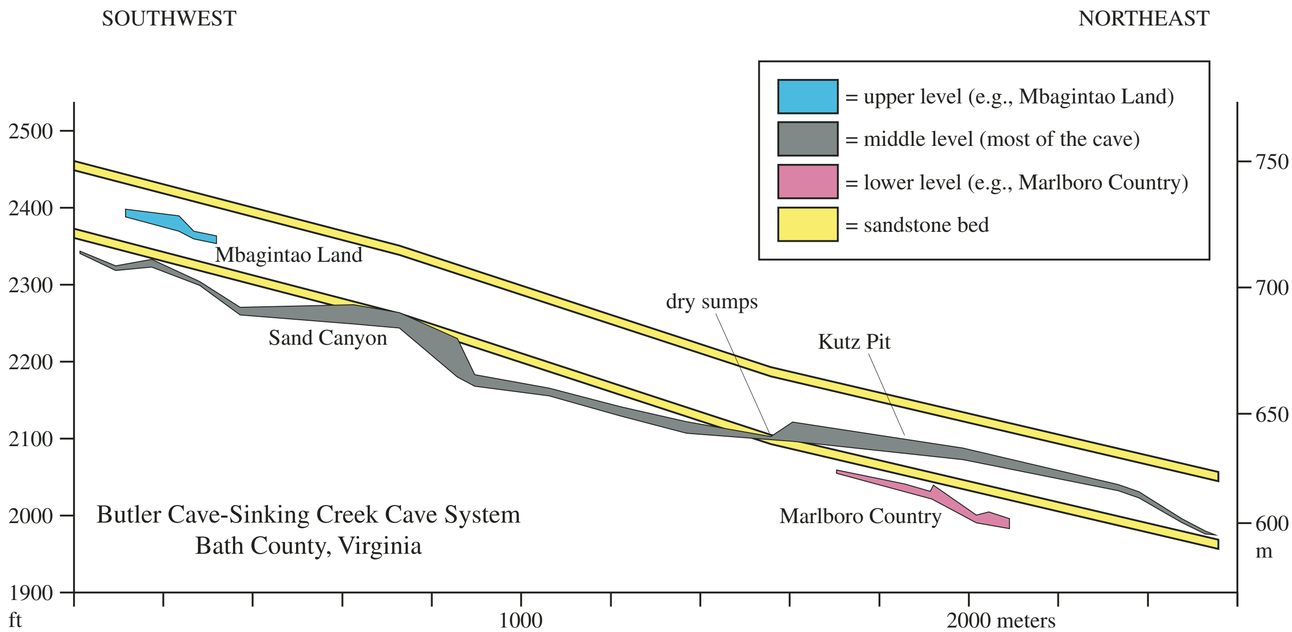


Figure 12. Geologic cross section of Butler Cave. The section approximately follows the axis of Sinking Creek syncline from southwest (left) to northeast (right). Elevation is relative to sea level.

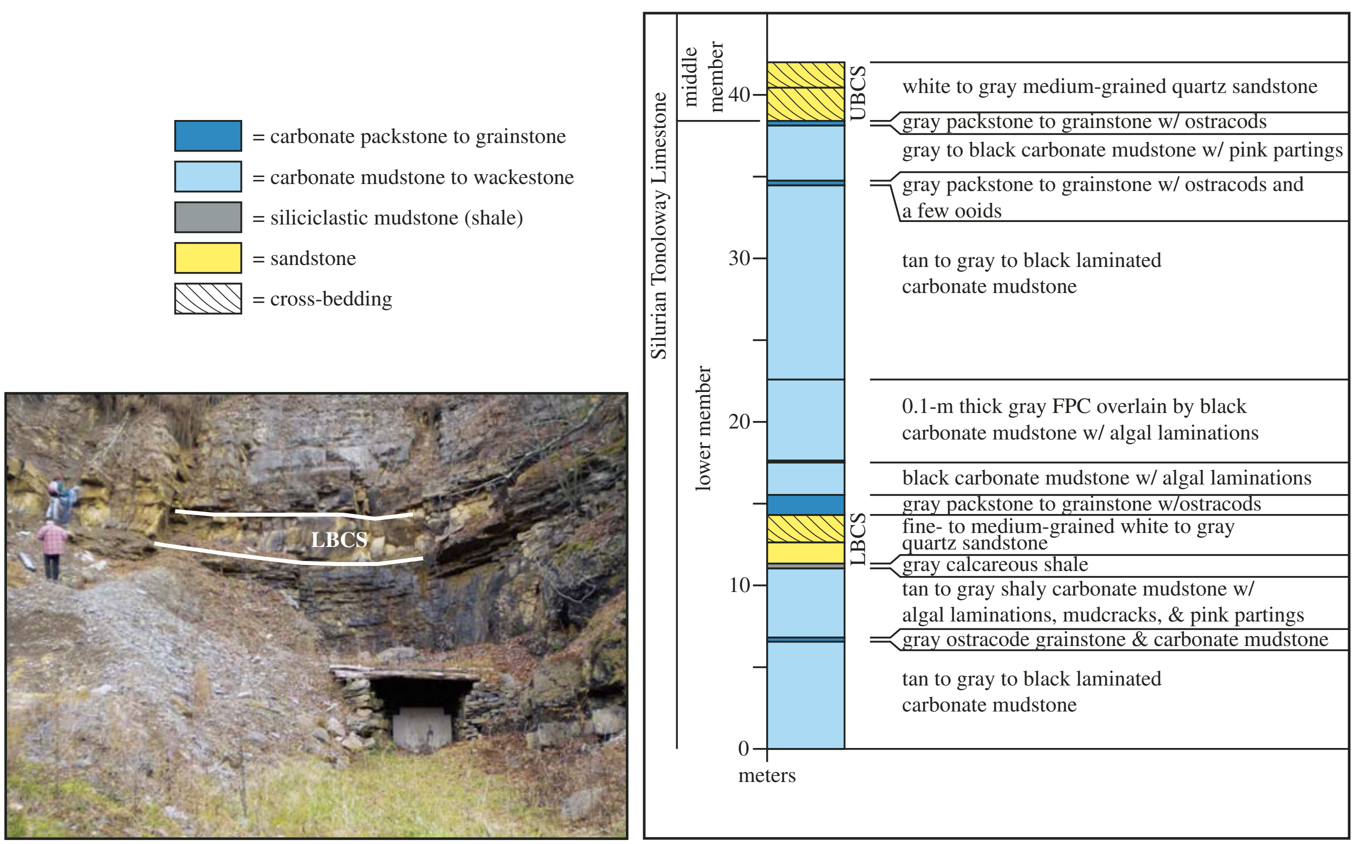


Figure 13. Outcrop above the SOFA entrance to Butler Cave (photograph by R.A. Lambert), and corresponding stratigraphic column at approximately the same scale. FPC—flat-pebble conglomerate; LBCS—lower Breathing Cave sandstone; UBCS—upper Breathing Cave sandstone. At this outcrop, a bed of fine-grained to medium-grained sandstone (LBCS) is present within the predominantly carbonate strata. This bed of sandstone forms the ceiling of Butler Cave at many locations, and is correlated with the "lower Breathing Cave sandstone" that forms the floor of Breathing Cave ~2.6 km north of the entrance of Butler Cave.

The Stop 2 tour of Butler Cave begins at the SOFA entrance (Fig. 13), then proceeds along a southeast-trending passage (Dave's Gallery) that follows the dip of the bedding, then goes through a northeast-trending passage that follows a fold axis, and then arrives at a set of rimstone dams (Fig. 14). From these rimstone dams, the tour continues along a southeast-trending passage and stops where the passage turns again to the northeast at the start of a narrow passage named the 90 Ugh Crawl.

As an optional extension of the tour, if conditions permit and the participants are willing, the tour will continue through the 90 Ugh Crawl to where the passage becomes walkable again, and then go southeast to intersect Sand Canyon, which is the major trunk channel of the cave system. This trunk channel trends northeast, along the axis of Sinking Creek syncline. Upon arriving at Sand Canyon, the tour will proceed northeast (downstream) along Sand Canyon to examine spectacular sand and gravel accumulations within the trunk channel. The tour will then return and proceed to the southwest (upstream) portion of Sand Canyon to examine a site where potholes are carved into the lower member of the Tonoloway Limestone. After examining the potholes, the tour will return via the 90 Ugh Crawl to the SOFA entrance of the cave.

SOFA Entrance to Butler Cave

The SOFA entrance to Butler Cave is located in the lower member of the Tonoloway Limestone (Fig. 13). The outcrop above the SOFA entrance displays mostly thin beds of gray to blue limestone and dolomitic lime mudstone, with a few beds of bioclastic and oolitic packstone to grainstone. At this outcrop, a bed of fine-grained to medium-grained sandstone is also visible within the predominantly carbonate strata. This bed of sandstone is situated ~8 m above the SOFA entrance, and it forms the ceiling of Butler Cave at many locations. This bed of sandstone at the SOFA entrance of Butler Cave is correlated with the "lower Breathing Cave sandstone" that forms the floor of Breathing Cave, which is located ~2.6 km north of the entrance of Butler Cave (Haynes, 2014).

In the petrographic terminology of Pettijohn et al. (1972), the sandstone at the SOFA entrance is a quartz wacke (Fig. 15). On the basis of 300 point counts of one thin section of a sample from this location, the composition of the sandstone is $F_{63}M_{29}C_{08}P_{tr}$ (F—framework grains, M—matrix, C—cement, P—porosity, "tr"—trace amount). The framework grains are predominantly monocrystalline quartz, with minor to trace amounts of feldspar and carbonate fossil fragments. Most of the quartz

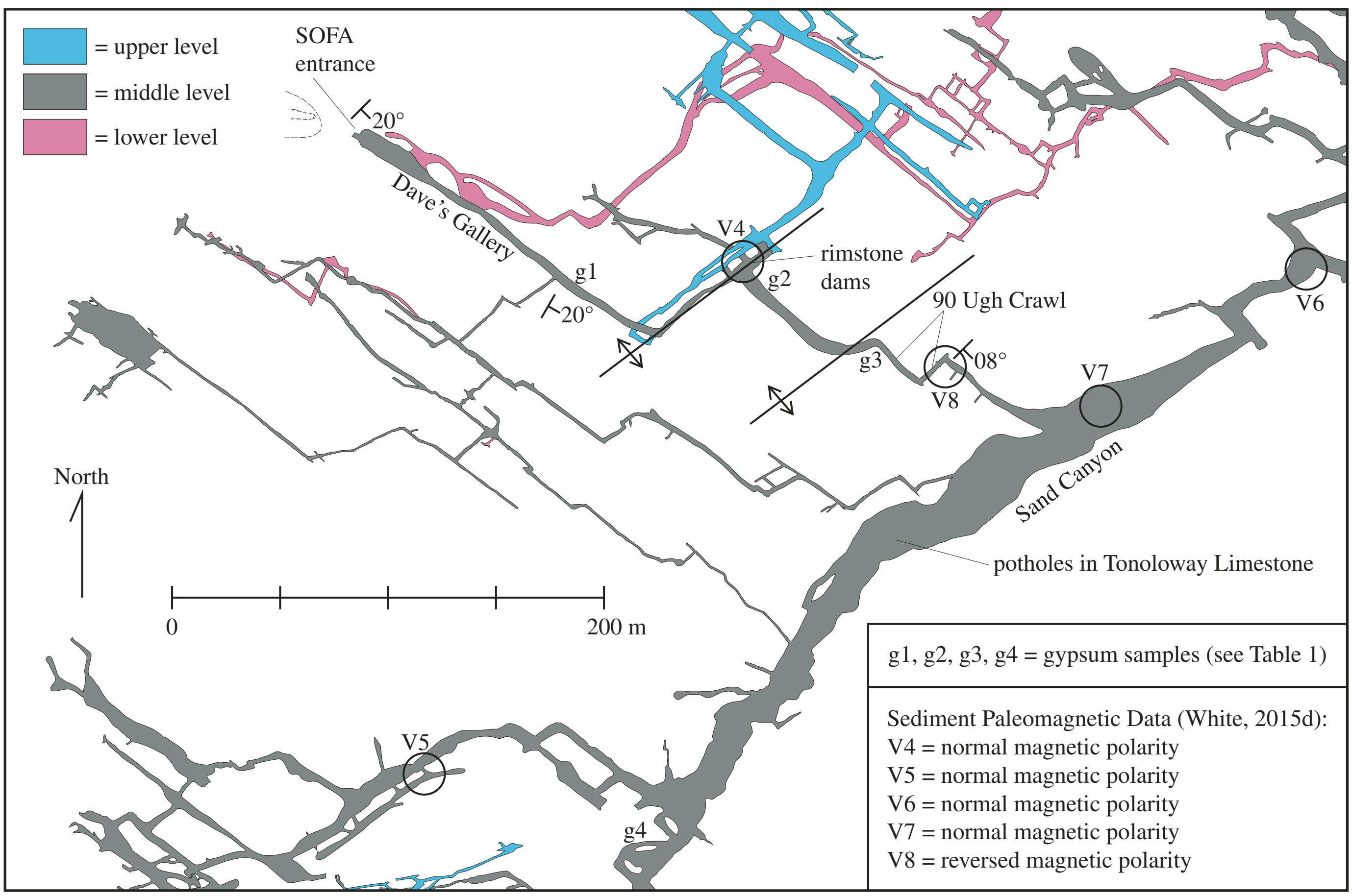


Figure 14. Detail of Butler Cave map showing field guide tour route and geology data.

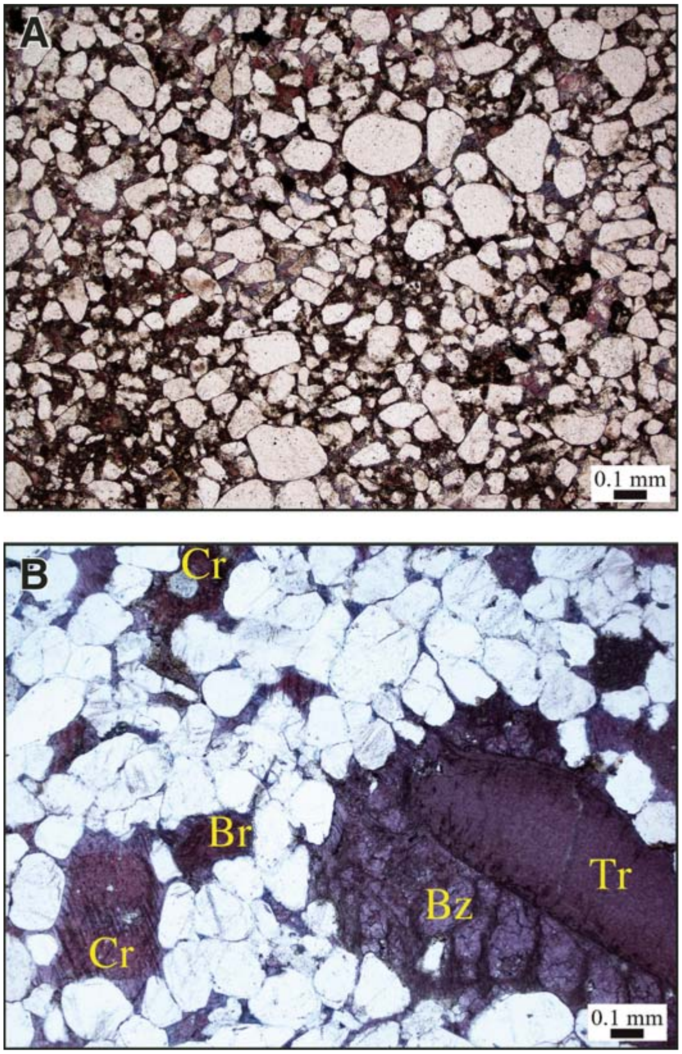


Figure 15. (A) Photomicrograph of sandstone within the lower member of Tonoloway Limestone at outcrop above the SOFA entrance to Butler Cave. The sample has been stained using the Dickson (1965) method, which allows for differentiation of carbonate minerals. This photomicrograph shows a quartz wacke with grains of rounded, subrounded, and subangular quartz (white) set in a muddy fine-grained matrix of a mixture of carbonate minerals (lilac and red colors) and siliciclastic (clay) minerals (dark yellowish brown). Photomicrograph by J.T. Haynes. (B) Photomicrograph of the basal sandstone unit (upper Breathing Cave sandstone) of the middle member of Tonoloway Limestone. This sample is from Chestnut Ridge 1.9 km (1.2 mi) east of the town of Burnsville. The sample has been stained using the Dickson (1965) method, which allows for differentiation of carbonate minerals. This photomicrograph shows a calcarenaceous quartzarenite with grains of quartz (white) and fragments of fossil debris including crinoids (Cr), bryozoans (Bz), brachiopods (Br), and trilobites (Tr). The fossil fragments are cemented by overgrowths of ferroan calcite (lilac color), and many of the quartz grains are cemented by overgrowths of quartz. Photomicrograph by J.T. Haynes.

grains are subrounded to subangular grains of fine to very fine sand (0.2–0.1 mm), although grain sizes range from medium sand (0.3 mm) to silt (<0.062 mm). The carbonate grains include a few very small brachiopod and crinoid fragments, but the origin of most of the carbonate material is not recognizable because the particles are too small. The matrix is a silty clay that varies in color from pale green to dark green to dark yellowish brown to brownish black. The composition of the silty clay is mostly illite, mixed-layer illite/smectite (I/S), chlorite, and minor carbonate mud. The non-matrix cement consists of ferroan dolomite (56% of the non-matrix cement) and ferroan calcite (44% of the non-matrix cement).

The petrographic data suggest several interpretations. Because of the striking petrographic similarity to mixed-layer illite/smectite in Paleozoic potassium-rich bentonites (Haynes, 1992, 1994), the illite/smectite in the sandstone bed at the SOFA entrance may be derived at least in part from the alteration of airborne volcanic ash (tephra) that mixed with the original sediments. The presence of appreciable matrix in the sandstone probably prevented the development and growth of quartz cement overgrowths on the quartz framework grains.

The composition of the sandstone at the SOFA entrance (lower Breathing Cave sandstone) is quite different from that of the upper Breathing Cave sandstone. In the petrographic terminology of Pettijohn et al. (1972), the upper Breathing Cave sandstone is a calcarenaceous quartzarenite (Fig. 15). On the basis of 300 point counts each of two thin sections of samples from an outcrop 1.9 km (1.2 mi) east of the road intersection at the town of Burnsville, the composition of the upper Breathing Cave sandstone is $F_{77}M_{tr}C_{21}P_{02}$. The framework grains are predominantly medium- to coarse-grained (0.5–1.0 mm) monocrystalline quartz, with the remainder being various bioclasts and a trace amount of detrital zircon and tourmaline. Matrix is nearly absent. The most common cement is optically continuous quartz overgrowths (81% of total cement), occurring around many of the quartz grains and recognized by euhedral crystal edges beneath "dust lines" between the detrital quartz framework grain and the quartz overgrowth. Less common cements are ferroan and non-ferroan calcite (19% of total cement) around the carbonate framework grains, including many optically continuous syntaxial overgrowths on echinoderm fragments, as well as minor ferroan dolomite cement (<1% of total cement) that occurs primarily in remnant interparticle pore spaces between quartz grains. Porosity is almost entirely secondary intraparticle porosity, and it appears to have developed as a result of the dissolution of both carbonate grains and carbonate cements.

In summary, there are some notable differences between the two sandstone beds. The lower sandstone displays a greater range in sizes of the framework grains (<0.062–0.3 mm), and very few of the framework grains are bioclasts. In contrast, the upper sandstone displays a smaller range in the sizes of framework grains (0.5–1.0 mm), and many of the framework grains are recognizable bioclasts (whole and fragmented fossils). In addition, the lower sandstone has a fine-grained sedimentary matrix, whereas the upper sandstone lacks such a matrix. Furthermore, the lower sandstone lacks quartz overgrowth cement and patches of silica (chert) cement, whereas the upper sandstone has abundant quartz overgrowth cement and abundant patches of silica (chert) cement.

Because of these differences, the lower sandstone is mechanically much weaker than the upper sandstone. In turn, these observations give rise to questions about why and how the sandstone beds preferentially formed the ceiling and (or) floor of the caves. A sandstone bed might have been a conduit for fluid flow, allowing fluids to move through the sandstone and preferentially dissolving the underlying limestone. Alternatively, a sandstone bed might have been a barrier to fluid flow, inhibiting fluids from reaching underlying or overlying limestone. A look at the geologic cross section through Butler Cave suggests that the lower sandstone was a barrier to fluid flow during the time of cave formation. As stated above, the restriction of the cave passages at the "dry sumps" (Figs. 11 and 12) suggests that the sandstone bed inhibited the movement of water and the development of cave passages. It appears likely that water flowing through phreatic passages reached the water table at the location of the dry sumps, and the development of additional cave passages was inhibited until water was able to breach the sandstone and enter the overlying limestone beds.

Dave's Gallery

The SOFA entrance of Butler Cave leads to Dave's Gallery, which is a passage that extends southeast, following the dip of the bedding in the lower member of the Tonoloway Limestone. In this portion of the cave, the strata dip 20° to the southeast toward the axis of Sinking Creek syncline (Fig. 16), and most of the passages trend northwest (Fig. 14). Some portions of these northwest-trending passages have relatively broad and flat ceilings defined by bedding planes so that passage profiles are approximately rectangular, square, or trapezoid. Other portions of these northwest-trending passages have relatively narrow ceilings defined by joints so that passage profiles are approximately triangular. The passages with joint-controlled morphologies tend to have greater height to width ratios. In Dave's Gallery, for example, bedding-plane controlled passages are generally 3–6 m high and 1.6–3 m wide, whereas joint controlled passages are generally 6–12 m high, 3 m wide at the base of the passage, and <0.2 m wide at the top of the passage where the joint is visible. In some instances, both bedding planes and joints exert strong controls on passage morphology, leading to a trapezoid passage profile with a narrow inverted V-shaped cavity at the top of the trapezoid shape (e.g., profile C–C′ in Fig. 17).

Fold Axis at Southern End of Dave's Gallery

The southern end of Dave's Gallery terminates at a passage that extends northeast and has an arched ceiling (profile D–D′ in Fig. 17). This passage follows the axis of a northeast-trending anticline within laminated carbonate mudstone of the lower member of the Tonoloway Limestone. There are many minor folds and faults within this member of the Tonoloway Limestone, and most of these folds and faults do not extend beyond individual beds. The minor faults show displacements ranging from a few centimeters to ~1.5 m.

Rimstone Dams

The northeast-trending passage that follows the fold axis terminates at some rimstone dams where shallow, slow-moving water flows out of a small passage from the northwest (Fig. 18). Several small dams are present, separating the passage into a series of small pools. Rimstone dams are usually composed of

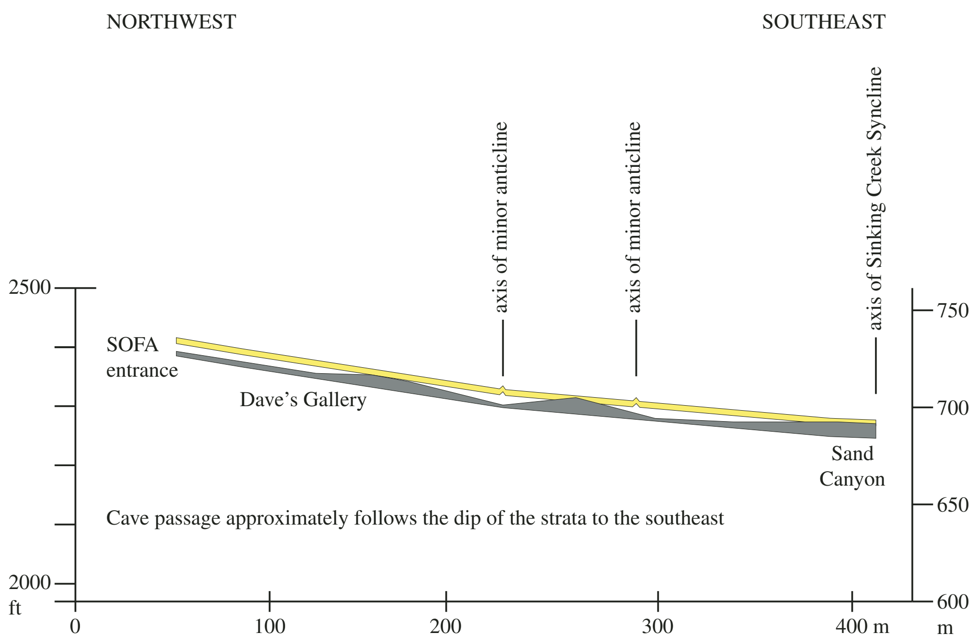
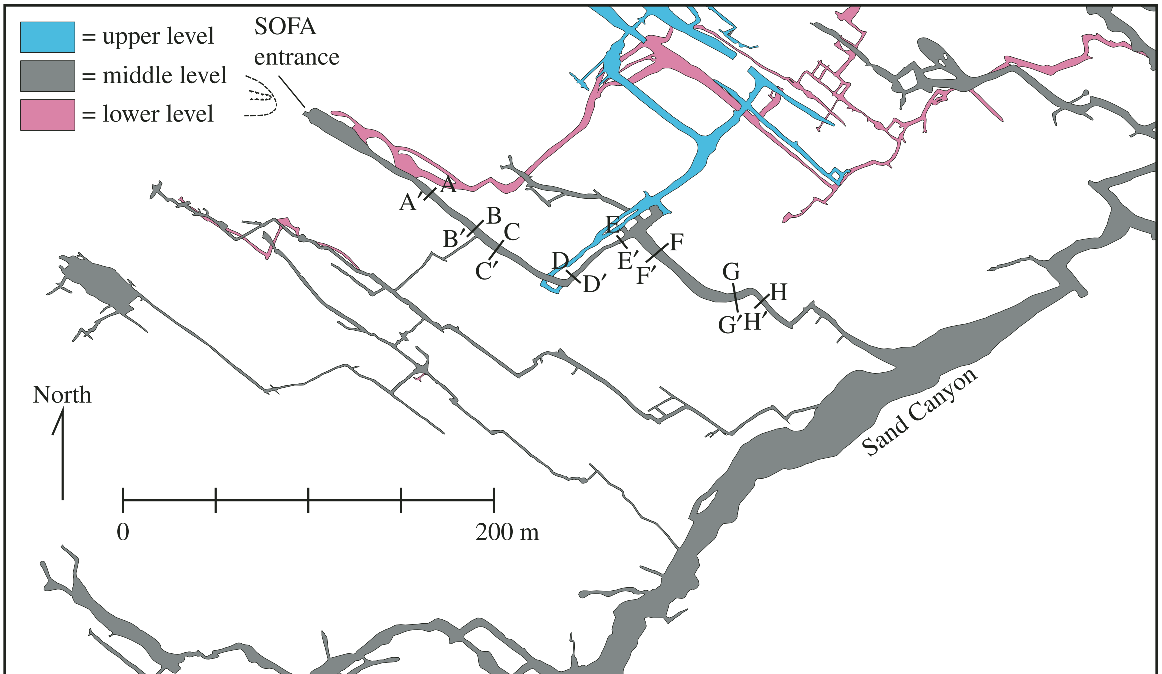
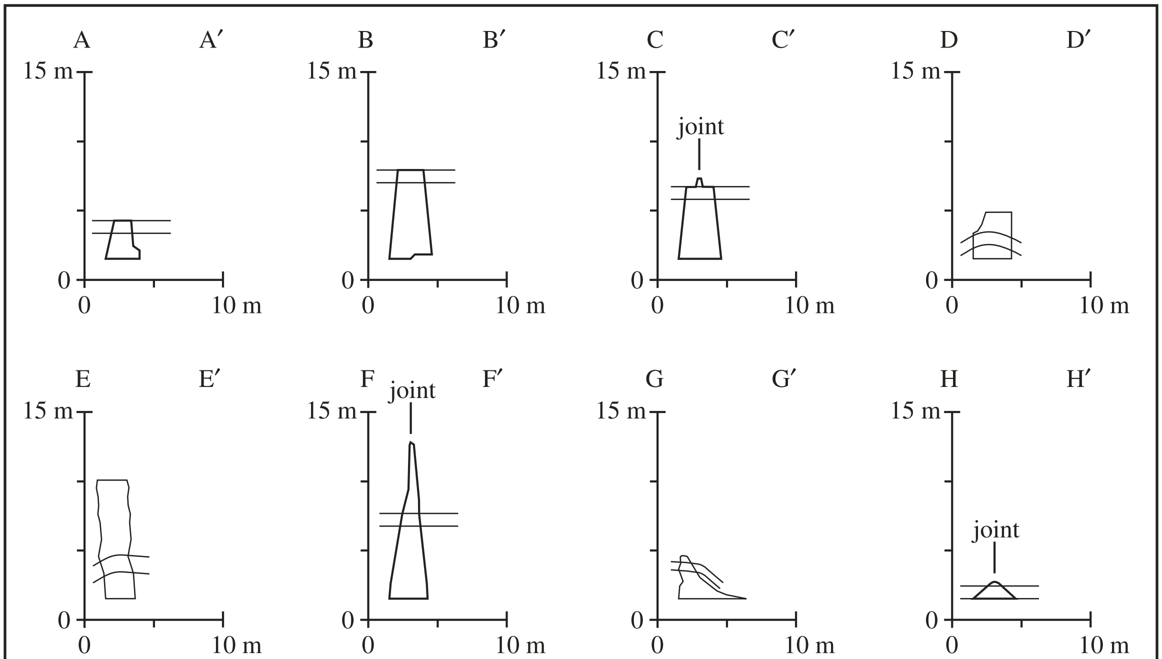


Figure 16. Geologic cross section along Dave's Gallery in Butler Cave from SOFA entrance to Sand Canyon. Yellow denotes sandstone bed (lower Breathing Cave sandstone). Elevation is relative to sea level.



Locations of passage profiles in Butler Cave (shown in box below).



Passages profiles in Butler Cave. Light-weight lines show approximate dip of bedding.

Figure 17. Passage profiles along Dave's Gallery in Butler Cave from SOFA entrance to Sand Canyon.

Figure 18. John Sweet beside rimstone dams in Butler Cave. Photograph by C.S. Swezey.

Figure 19. White gypsum crust along joints and laminations in the lower member of the Tonoloway Limestone, Butler Cave. An 8.9-cm-long pocket knife provides a sense of scale. Photograph by C.S. Swezey.

calcite, and they form in areas of ponded water that is supersaturated with calcium carbonate. Rimstone dams are not common in Butler Cave.

Gypsum Crusts along Passage Southeast of Rimstone Dams

From the rimstone dams, the cave tour follows the dip of the strata along another joint-controlled passage that extends to the southeast (profile F–F′ in Fig. 17). Needles and crusts of gypsum are present along walls of this passage (Fig. 19). Gypsum samples from this location and other locations in Butler Cave have yielded $\delta^{34}S$ sulfur isotope values ranging from –5.2 to –1.0 parts per thousand (Table 1). As discussed in Swezey et al. (2002) and Swezey and Piatak (2003), this range of isotope values provides some information about the possible source of the sulfur in the gypsum. Specifically, these isotope values indicate that the sulfur is not derived from the simple dissolution and re-precipitation of primary gypsum beds in the stratigraphic section. Gypsum that formed via simple precipitation from marine water would have positive sulfur isotope values. Instead, the negative sulfur isotope values suggest that there has been some biologic fractionation either (1) during recent formation of the gypsum, or (2) during the distant geologic past, with the sulfur being derived from a source of biologically fractionated sulfur (e.g., from organically-bound sulfur, or from the oxidation of sulfide minerals such as pyrite in Devonian shale). In addition, the concentration of sulfate (SO_4^{2-}) in stream water in Butler Cave is ~16 mg per liter (mg/L), which is too low to precipitate gypsum, and thus the sulfur in the cave gypsum is probably being transported by seepage groundwater. Furthermore, studies of Butler Cave meteorology have revealed that the relative humidity at Sand Canyon is ~99% (Wefer, 1991; Wefer and Lucas, 2015). A relative humidity of 99% indicates that the air is nearly saturated with water, and thus the process of gypsum formation is not likely to have occurred via precipitation with evaporation. Instead, the process of gypsum formation is likely to have occurred via a chemical reaction (not evaporation) as sulfate in the groundwater reacted with the limestone, when the groundwater came into contact with the cave atmosphere.

The passage that extends southeast from the rimstone dams terminates at a passage that extends northeast and has an asymmetric profile (profile G–G′ in Fig. 17). Like the previous northeast-trending passage, this passage follows the axis of a fold within laminated carbonate mudstone of the lower member of the Tonoloway Limestone. The core of the fold is deformed by several small thrust faults, with displacement of less than a few meters (Fig. 20). This area of deformation marks the beginning of a constricted area named the 90 Ugh Crawl, which is ~1 m high and extends for ~15 m (Fig. 14). Paleomagnetic studies have revealed the presence of sediment with normal geomagnetic polarity and sediment with reversed geomagnetic polarity in Butler Cave (White, 2015d). The 90 Ugh Crawl is one site with sediment of reversed geomagnetic polarity (Fig. 14), indicating that

TABLE 1. SULFUR ISOTOPE VALUES FROM GYPSUM IN BUTLER CAVE

Location on Figure 14	USGS laboratory sample ID	Sulfur isotope values	Gypsum morphology
g1	BUT-1-CS	–5.2	Crust
g2	BUT-2-CS	–1.0	Crust
g2	BUT-2-CS duplicate*	–1.0	Crust
g3	BUT-3-CS	–4.1	Crust
g4	BUT-4-CS	–4.1	Needles and flowers

Note: From Swezey and Piatak (2003). USGS—U.S. Geological Survey.

*Duplicate—duplicate analysis of sample g2.

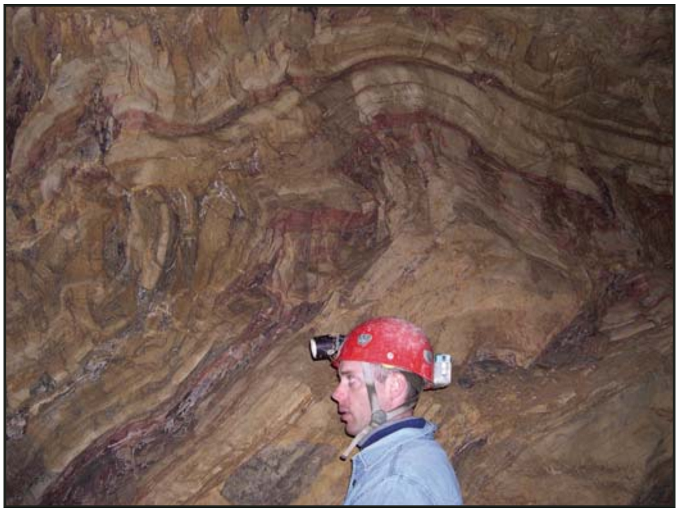


Figure 20. Benjamin Schwartz beside faults and folds in the lower member of the Tonoloway Limestone at the northwest entrance to the 90 Ugh Crawl in Butler Cave. Photograph by C.S. Swezey.

Figure 21. Cobbles of various lithologies on the floor of Sand Canyon in Butler Cave. Photograph by C.S. Swezey.

this sediment is at least 781,000 years old (the age of the most recent geomagnetic polarity reversal, according to Gradstein et al., 2004).

Optional Portion of Stop 2

At this point in the tour, there are a few additional sites that might be visited but these sites are optional. The tour will go to these additional sites only if ALL participants are interested and willing to go through the 90 Ugh Crawl and reach the walkable passages beyond. The additional sites are located along Sand Canyon (the main trunk channel of the cave) and include the axis of Sinking Creek syncline, extensive sand and gravel accumulations within the cave trunk channel, and potholes within laminated carbonate mudstone of the lower member of the Tonoloway Limestone.

Sand Canyon Camp

After exiting the 90 Ugh Crawl, the passage turns southeast. The dip of the beds is very low in this area, and the passage profile is blocky and follows bedding planes. The passage extends to the southeast and intersects a much larger cave passage named Sand Canyon, which is ~9–25 m wide, 6–9 m high, trends northeast, and is the main trunk channel of Butler Cave. The site where the passage from the 90 Ugh Crawl intersects with Sand Canyon is named Sand Canyon Camp. The passage from the 90 Ugh Crawl comes out on top of a vast accumulation of sand and gravel that is a bank of sediment along the inside curve of a meander of an underground river. At Sand Canyon Camp, a prominent fold in the strata is visible in the ceiling on the southeast wall of Sand Canyon. The axis of Sinking Creek syncline is located near the southeast side of Sand Canyon, such that the main trunk channel of Butler Cave essentially follows the syncline axis (Fig. 11).

Main Trunk Channel Northeast (Downstream) of Sand Canyon Camp

The floor of Sand Canyon is covered with gravel of various sizes (ranging from granule to boulder) and various lithologies (e.g., limestone, sandstone). Some gravel is angular and has probably not been transported very far, whereas other gravel is well rounded and has probably had a much longer transport history. Some gravel lithologies are quite distinctive, such as the laminated limestone of the Silurian Tonoloway Limestone and the red sandstone of the Silurian Rose Hill Formation (Fig. 21). The cobbles of Tonoloway Limestone probably came from within the cave, whereas the cobbles of Rose Hill Formation are "exotic" and must have come from outside the cave. The diversity of exotic lithologies reflects the diversity of outcrops within the drainage basin during the time that the cobbles entered the cave.

Near the northern end of Sand Canyon, prominent banks of gravel and sand are present against the southeast wall of the cave passage (Fig. 22). These banks display a sedimentary sequence that consists of a 1.5-m-thick unit of gravel and sand, overlain by 0.5-m-thick unit of coarse to medium sand. The sand displays parallel laminations (upper plane bed laminations) and cross-bedding. Most of the cross-bedding dips to the northeast, indicating flow direction to the northeast. These exposures of sediment are interpreted as the deposits of a high energy debris flow, with the overlying sand having accumulated as flow velocities waned to the point where gravel could no longer be transported.

At the very northern terminus of Sand Canyon, a 7.3-m-high accumulation of gravel, sand, and mud is visible against the northwest wall of the cave passage. These sediments extend the

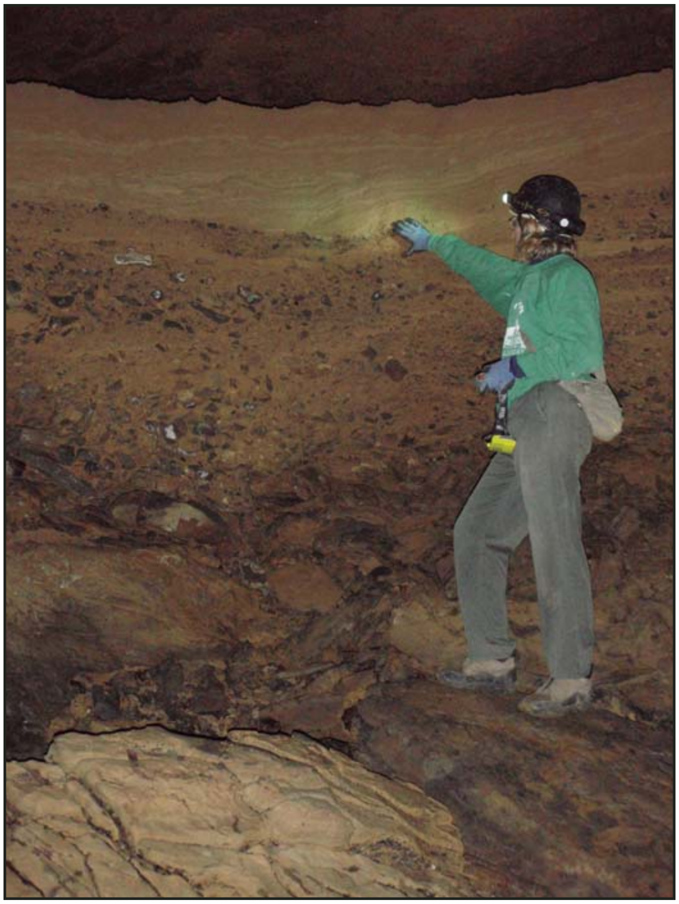


Figure 22. Selene Deike in front of bank of gravel and sand along the southeast side of Sand Canyon, the main trunk channel of Butler Cave. Photograph by C.S. Swezey.

Figure 23. Potholes in the lower member of the Tonoloway Limestone at the southwest (upstream) end of Sand Canyon in Butler Cave. An 8.9-cm-long pocket knife provides a sense of scale. Photograph by C.S. Swezey.

entire distance from the passage floor to the ceiling, suggesting that the passage was filled completely with sediment at least once during its history. Most of this exposure reveals multiple sequences, each of which consists of a unit of gravel and sand overlain by a unit of sand. Each of these sequences is interpreted as a debris flow deposit, similar to the sequence of gravel and sand exposed along the southeast wall of Sand Canyon. Near the ceiling of Sand Canyon, however, there appears to be a unit of mud, which is thought to have accumulated during very low energy ("slackwater") conditions when most of the entire cave passage was filled with gravel and sand.

In contrast with Sand Canyon, many of the northwest-trending side passages to the northeast (downstream) of Sand Canyon are filled primarily with mud that in places appears to be 3–6 m thick, but has been incised and is currently undergoing some erosion. Such thick mud accumulations are interpreted as low energy deposits ("slackwater" conditions). In summary, the various sequences of gravel, sand, and mud suggest that the cave has experienced multiple episodes of sediment accumulation and sediment erosion, and that the modern setting is primarily a setting of sediment erosion and removal.

Main Trunk Channel Southwest (Upstream) of Sand Canyon Camp

A short distance southwest (upstream) from Sand Canyon Camp, the major trunk channel of Butler Cave has a smaller diameter, and there are notably fewer sand and gravel deposits along the trunk channel. There are several possible explanations for this relatively abrupt absence of sand and gravel: (1) sand and gravel may have once accumulated at this site, but have since been removed by erosion; (2) sand and gravel may have been transported from the southwest along the trunk channel, but only accumulated where the channel became wider (where flow velocity would have decreased); and (3) the sand and gravel may have entered the cave via Dave's Gallery/90 Ugh Crawl, and were never present farther upstream along the main trunk channel.

Potholes in the Tonoloway Limestone are well exposed in this portion of Sand Canyon, where the size of the trunk channel decreases (Fig. 23). Such potholes are thought to form where pebbles and (or) sand grains are swirled around by turbulent water, eroding into the relatively soft substrate of the Tonoloway Limestone. Erosion by cavitation (implosion of air bubbles

Figure 24. The Pancake Field, Highland County, Virginia (Stop 3-1). The flatness of the field is quite striking in this karst landscape. Photograph by P.C. Lucas.

in water) may also play a role in the formation of potholes. Similar potholes are present in a similar setting in the Water Sinks Cave System along the Sweet Dreams Passage, which is a cave passage in thin-bedded and laminated carbonate mudstone of the lower member of the Tonoloway Limestone.

Drive from the Homestead of the BCCS to Monterey, Virginia. Spend the night in Warm Springs.

▪ DAY 2 (2 April 2017)

Drive from Warm Springs to the Water Sinks depression, Highland County, Virginia.

The second day of the field trip begins at the Water Sinks depression (Stop 3), which is a very large compound sink located on private property. After Stop 3, the field trip proceeds to Aqua Spring (Stop 4), which is the resurgence site of water flowing through Butler Cave and the Water Sinks Cave System.

Figure 25. Limestone cliff in the largest sinkhole of the Water Sinks depression and some of the entrances to Water Sinks Cave System (Stop 3-2), Highland County, Virginia (modified from Swezey et al., 2015). Photograph by P.C. Lucas.

Stop 3: Water Sinks Depression, Highland County, Virginia (N 38° 13′ 20″, W 79° 36′ 19″)

This is private property. Please obtain consent from the owners before proceeding onto the property.

At this stop, the field trip visits the following four sites within the Water Sinks depression: (1) Stop 3-1: the Pancake Field and the Sink of Sinking Creek; (2) Stop 3-2: the Water Sinks Cave System, which consists of the Old Water Sinks Cave and the Water Sinks Subway Cave; (3) Stop 3-3: Owl Cave; and (4) Stop 3-4: Helictite Cave. Access to these caves may be obtained only via permission of the property owners, and coordinates of the cave entrances are not given in this field guide in order to protect the caves from possible vandalism.

Stop 3-1: The Pancake Field and the Sink of Sinking Creek, Highland County, Virginia

The Water Sinks depression is a very large and deeply incised blind valley near the northern end of Burnsville Cove (Fig. 2). This blind valley is a closed topographic depression that is the terminal sink point of several surface streams, most of which are intermittent streams (Lucas, 2015a). The southern part of the Water Sinks depression is a 1.6-km-long flat field named the Pancake Field on the southeast side of U.S. Route 609 (Fig. 24). Sinking Creek and Water Sinks Creek are two of the intermittent streams that flow into the Water Sinks depression. Sinking Creek enters from the southwest and flows into a swallet in the bottom of a steep-walled sinkhole at the southern end of the Pancake Field. Water Sinks Creek enters the blind valley from the slopes of Jack Mountain to the west, and flows under U.S. Route 609 as two separate channels. These separate channels become a single channel, and then the stream flows east across the Pancake Field to the flank of Chestnut Ridge. From there, the stream turns and flows north. One of the sink points is the location of Owl Cave (Stop 3-3), which functions as a high-water overflow for Water Sinks Creek. After flowing around the west and north sides of a knoll in the center of the Water Sinks depression, the streambed reaches its lowest surface elevation at the base of a 34-m-high limestone cliff that forms the northwest side of the knoll (Fig. 25). This limestone cliff provides access to the Water Sinks Cave System (Stop 3-2), including the several entrances to Old Water Sinks Cave and the engineered entrance to Water Sinks Subway Cave.

In 2009, the U.S. Geological Survey (USGS) took a 19.6-m-long core from the Pancake Field (Bernhardt, 2016). This core consists of 20–50-cm-thick beds of clay and silt, and a few 10–20-cm-thick beds of sandy gravel composed of limestone, sandstone, and black shale. The beds of silt and clay are interpreted as lacustrine sediments that accumulated when the sinkhole at the north side of the Pancake Field was blocked and prevented water from draining from the Pancake Field. At 17–12 m depth, the core consists of laminated clay that has yielded palynomorphs having dominant taxa of *Picea* spp. (spruce), *Abies* (fir), *Pinus banksiana* (jack pine), and *Lycopodium* (ground pine or creeping cedar). At 13.24–13.22 m depth, organic matter in the laminated clay yielded a bulk radiocarbon age of 21,710 ± 90 ^{14}C yr B.P. [USGS radiocarbon sample WW9119]. The pollen data and the radiocarbon age indicate that this area was occupied by a boreal-like forest during the last glacial maximum.

Stop 3-2: Water Sinks Cave System

Stop 3-2 is located within the biggest sinkhole of the Water Sinks depression. In this area, there are several levels of interconnected cave passages that comprise the vast Water Sinks Cave System (Fig. 26). There are two upper cave levels that form the Old Water Sinks Cave System. As recounted in Lucas (2015a), the Old Water Sinks Cave System was previously known as "Siphon No. 2 Cave." Deike (1960a) published a map of this cave, which was surveyed in 1959 by Ruth Deike, John Haas, and Bette White. Although the cave has several entrances, the survey of the cave was considered to be complete as of 1990. During the winter of 1997–1998, however, a flood caused the sinkhole at the base of the 34-m-high escarpment to become a lake, after which a collapse occurred at one of the cave entrances and created a 5.5-m-deep hole through an accumulation of old flood sediments and plant debris. Another collapse occurred in the same area in 2006. In 2007, after some digging around the collapse site, several lower cave levels were discovered and were designated as the Subway Section of the Water Sinks Cave System. As of 20 September 2016, the entire Water Sinks Cave System was known to have a total passage length of 2.22 m (11,767 ft) and was not connected to nearby Owl Cave.

The Water Sinks Cave System is located along the axis of the northeast-trending Chestnut Ridge anticline (Fig. 26). The bedding dips away from the anticline axis at relatively low angles, and many of the cave passages have flat ceilings defined by bedding planes. Maze patterns are common in parts of the Water Sinks Cave System and many of the passages are developed along systematic joint systems that are orthogonal to bedding.

Unlike Butler Cave (Stop 2), which is located in the lower member of the Tonoloway Limestone, the Water Sinks Cave System is located in higher stratigraphic intervals. The Water Sinks Cave System is developed in the upper member of the Tonoloway Limestone, the Clifton Forge Sandstone Member of the Keyser Limestone, and the Jersey Shore Limestone Member of the Keyser Limestone (Fig. 27). The 34-m-high escarpment in the Water Sinks depression exposes the upper part of the Jersey Shore Limestone Member, the LaVale Limestone Member of the Keyser Limestone, the New Creek Limestone, and part of the Corriganville Limestone (Fig. 27). The cave entrances and many of the cave passages of the Old Water Sinks Cave section of the system occur preferentially in the upper of two beds with bioherms and biostromes of stromatoporoids and corals within the Jersey Shore Limestone Member of the Keyser Limestone (Fig. 28). A detailed study of the lower of the two stromatoporoid-coral buildups in the Water Sinks Cave System revealed that >95% of the stromatoporoids were disturbed (not in situ), and that the ratio

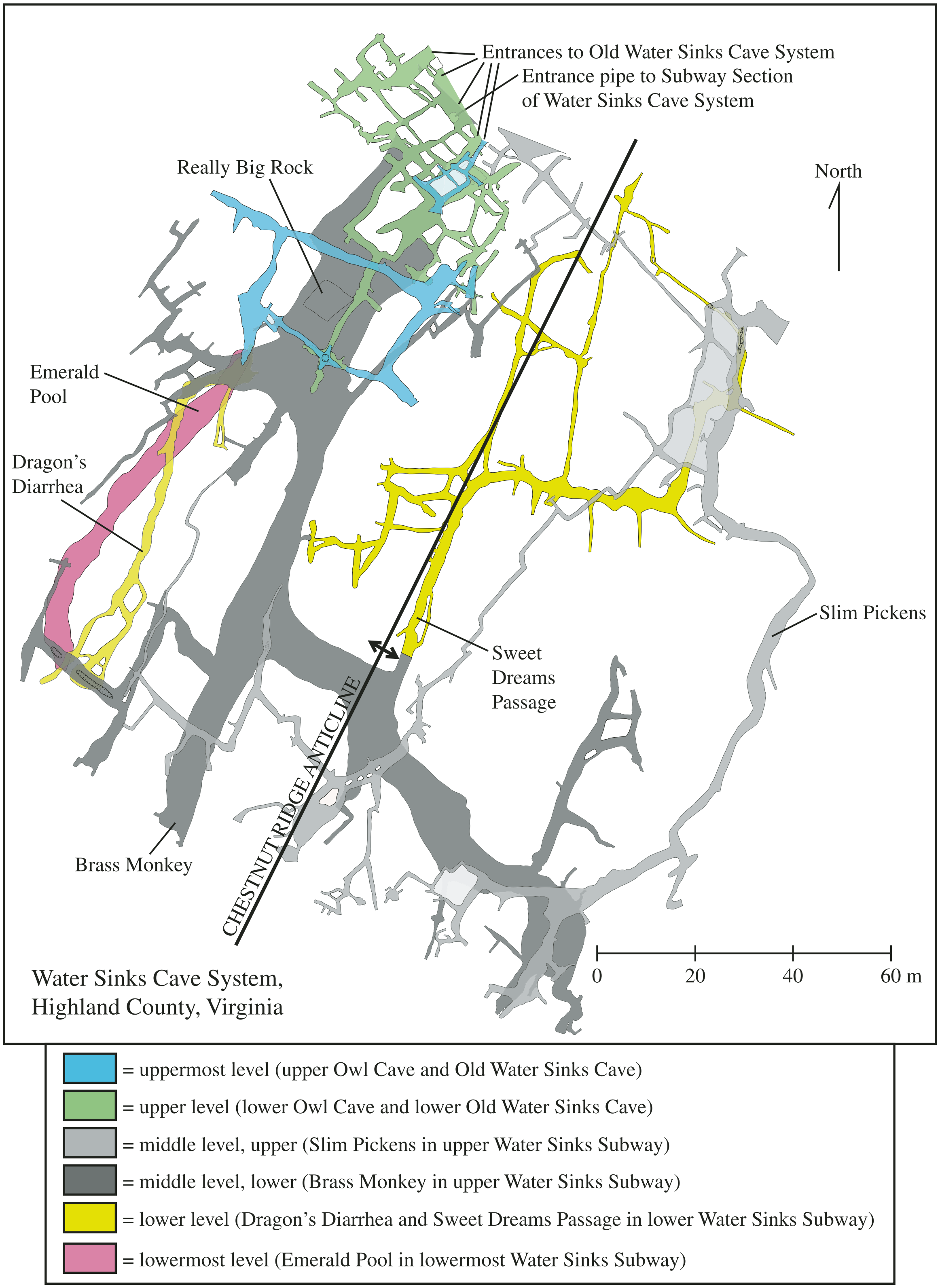


Figure 26. Map of Water Sinks Cave System (modified from maps associated with Lucas, 2015a) and approximate location of the Chestnut Ridge anticline.

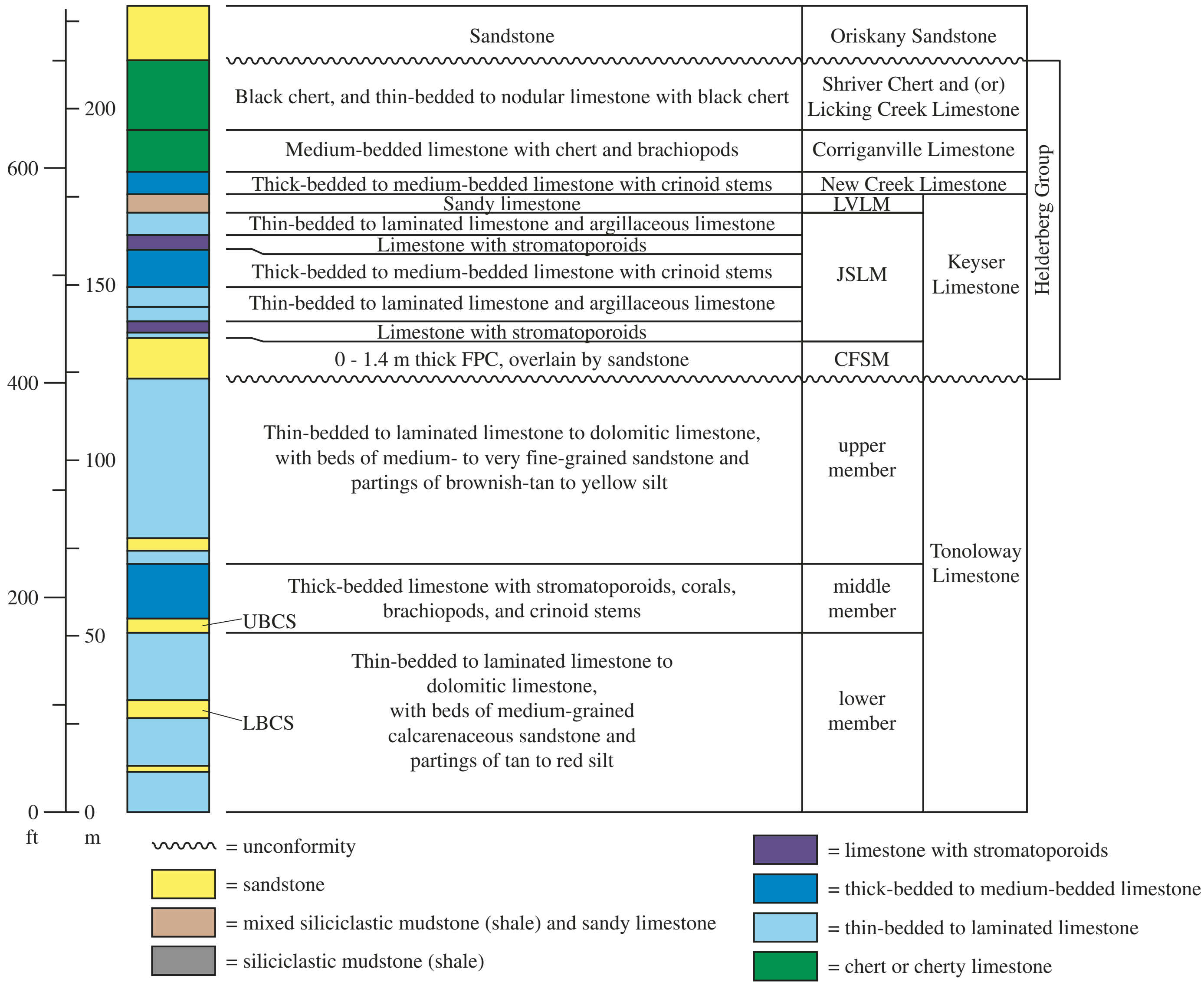


Figure 27. Composite stratigraphic section of the Water Sinks area (modified from Swezey et al., 2015). CFSM—Clifton Forge Sandstone Member; FPC—flat-pebble conglomerate; JSLM—Jersey Shore Limestone Member; LBCS—lower Breathing Cave sandstone; LVLM—LaVale Limestone Member; UBCS—upper Breathing Cave sandstone.

of stromatoporoids to tabulate corals decreases vertically upward through the buildups (Cole et al., 2015).

The preferential development of caves in reef-bearing strata is common in this part of Burnsville Cove (e.g., Water Sinks Cave System, Owl Cave), and is quite different from the setting of Butler Cave and other caves on the west side of Burnsville Cove. It is not clear whether the preferential location of caves in the reef-bearing strata is a function of: (1) location near the axis of the Chestnut Ridge Anticline; or (2) lithologic properties of the reef-bearing strata. The observations that >95% of the stromatoporoids are disturbed (not in situ) and that the ratio of stromatoporoids to tabulate corals decreases vertically upward through the buildups prompts speculation that the basal parts of the stromatoporoid buildups were areas of greater porosity and permeability. If this is true, then initial fluid flow may have been concentrated initially near the base of the stromatoporoid buildups, resulting in preferential locations for cave development.

The presence of stromatoporoid reefs in this area may be related to some aspects of Silurian tectonics or paleogeography. Dennison (1985) and Dennison et al. (1997) suggested that the presence of a stromatoporoid reef near the town of Mustoe (Fig. 1) was related to a tectonic hinge-line that trends across the area. A change in depositional environment in this area is suggested by the facies changes between Stop 3-2 in southern Highland County and Stop 1 in north-central Highland County whereby the Clifton Forge Sandstone Member of the Keyser Limestone

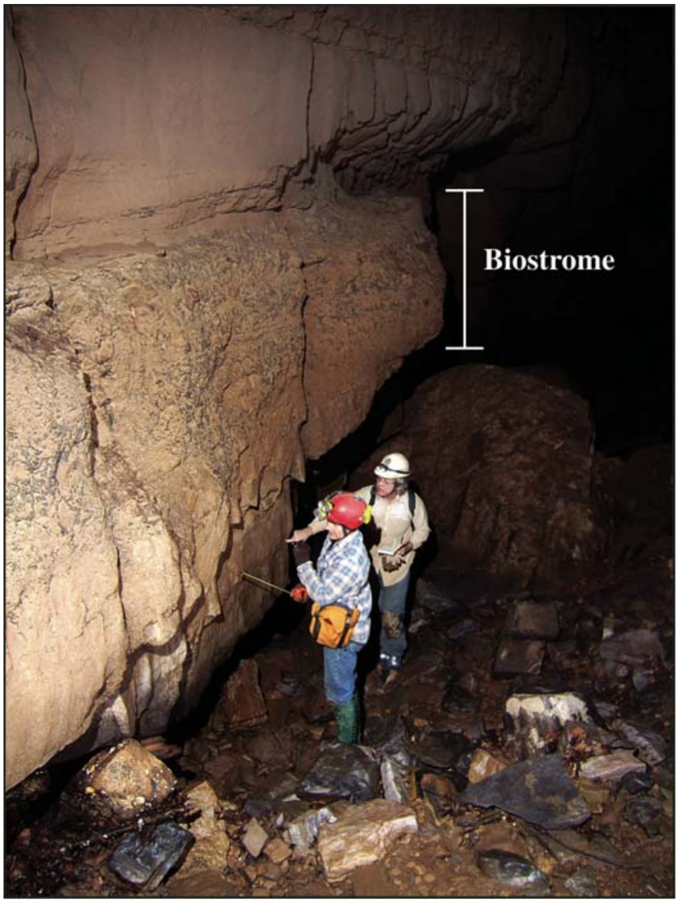


Figure 28. Richard (Rick) Lambert and John Haynes in front of stromatoporoid and coral biostrome in the Jersey Shore Limestone Member of the Keyser Limestone, Subway Section of the Water Sinks Cave System, Highland County, Virginia (modified from Haynes, 2014). Photograph by P.C. Lucas.

grades north into the Big Mountain Shale Member (Fig. 9), and also by the thinning and (or) pinching out to the north of sandstone beds in the Tonoloway Limestone. Thus, it appears that sand in relatively shallow water at Burnsville Cove and farther south changed to siliciclastic mud in relatively deeper water farther north. In the vicinity of Burnsville Cove, stromatoporoid reefs may have grown preferentially along a break in topographic slope that separated sand (to the south) from siliciclastic mud (to the north). Stromatoporoid reefs have been reported from correlative strata in Pennsylvania (Coffey and Taylor, 1989; Wertz and Schiappa, 2009), and it is possible that additional mapping of the distribution of stromatoporoid reefs throughout the Appalachian Basin might reveal that Burnsville Cove is located on the southern margin of a Silurian depositional basin that is rimmed by stromatoporoid reefs.

Stop 3-3: Owl Cave

Owl Cave is located on the southeast side of the Water Sinks depression, and has several entrances (Fig. 29). The cave was known for many years as "Siphon No. 1 Cave" (Lucas, 2015a), and was first mapped by Deike (1960a). The cave was later resurveyed by Phil Lucas and others, and the new map includes a previously unsurveyed upper level (shown in blue in Fig. 30) of the cave. As of October 2015, Owl Cave was known to have a total passage length of 959 m. Although the passages of Owl Cave overlie many of the passages of the Subway Section of the Water Sinks Cave System, Owl Cave is not yet connected to the greater Water Sinks Cave System (Fig. 31). As of October 2015, the closest known passage in Owl Cave is only 7 m from known passage in the Water Sinks Cave System.

Few geologic details are available for Owl Cave. Part of the cave is located near the axis of the northeast-trending Chestnut Ridge anticline, although most of the cave lies to the east of the

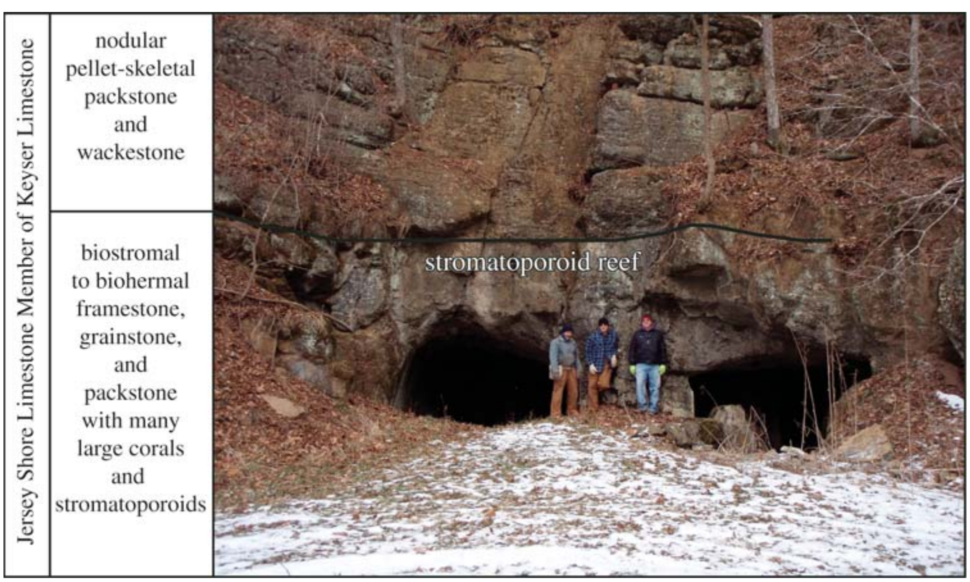


Figure 29. Richard (Rick) Lambert, Phil Lucas, and John Haynes in front of the main entrances to Owl Cave, Highland County, Virginia (modified from Swezey et al., 2015).

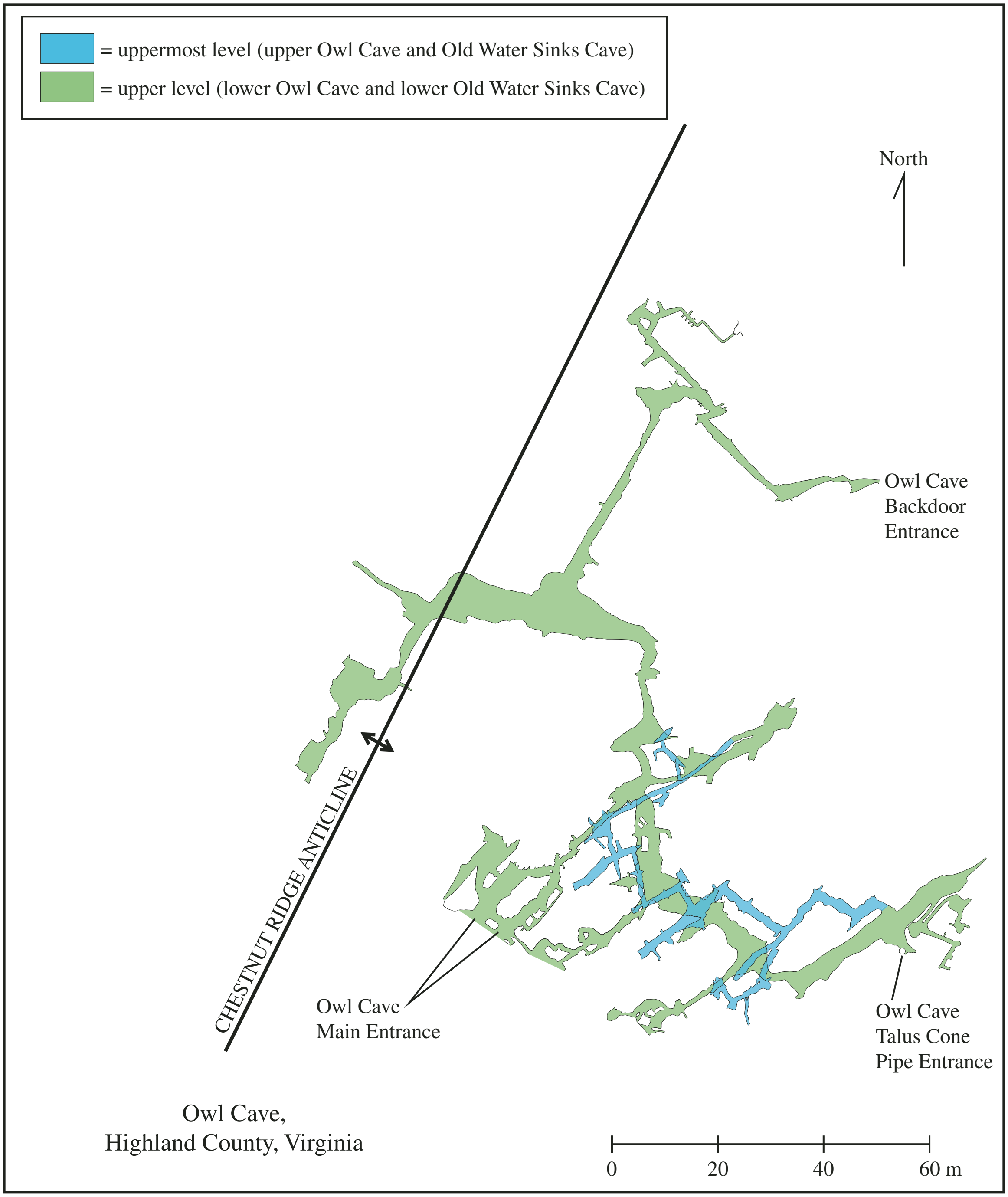


Figure 30. Map of Owl Cave (modified from maps of Lucas, 2015a) and approximate location of the Chestnut Ridge anticline.

anticline axis (Fig. 26). As with the Water Sinks Cave System, the bedding in Owl Cave dips away from the anticline axis at relatively low angles. With respect to stratigraphic setting, Owl Cave is situated within the Keyser Limestone (Lucas, 2015a). The twin main entrances to the cave and the more extensive lower level of the cave are developed in stromatoporoid- and coral-bearing beds of the Jersey Shore Limestone Member of the Keyser Limestone (Fig. 32). The smaller (and engineered) Talus Cone entrance is located in the lower part of the New Creek Limestone, which is 7.5 m thick at this site. The "Talus Cone entrance" provides access to the less extensive upper level of the cave, which is developed in the LaVale Limestone

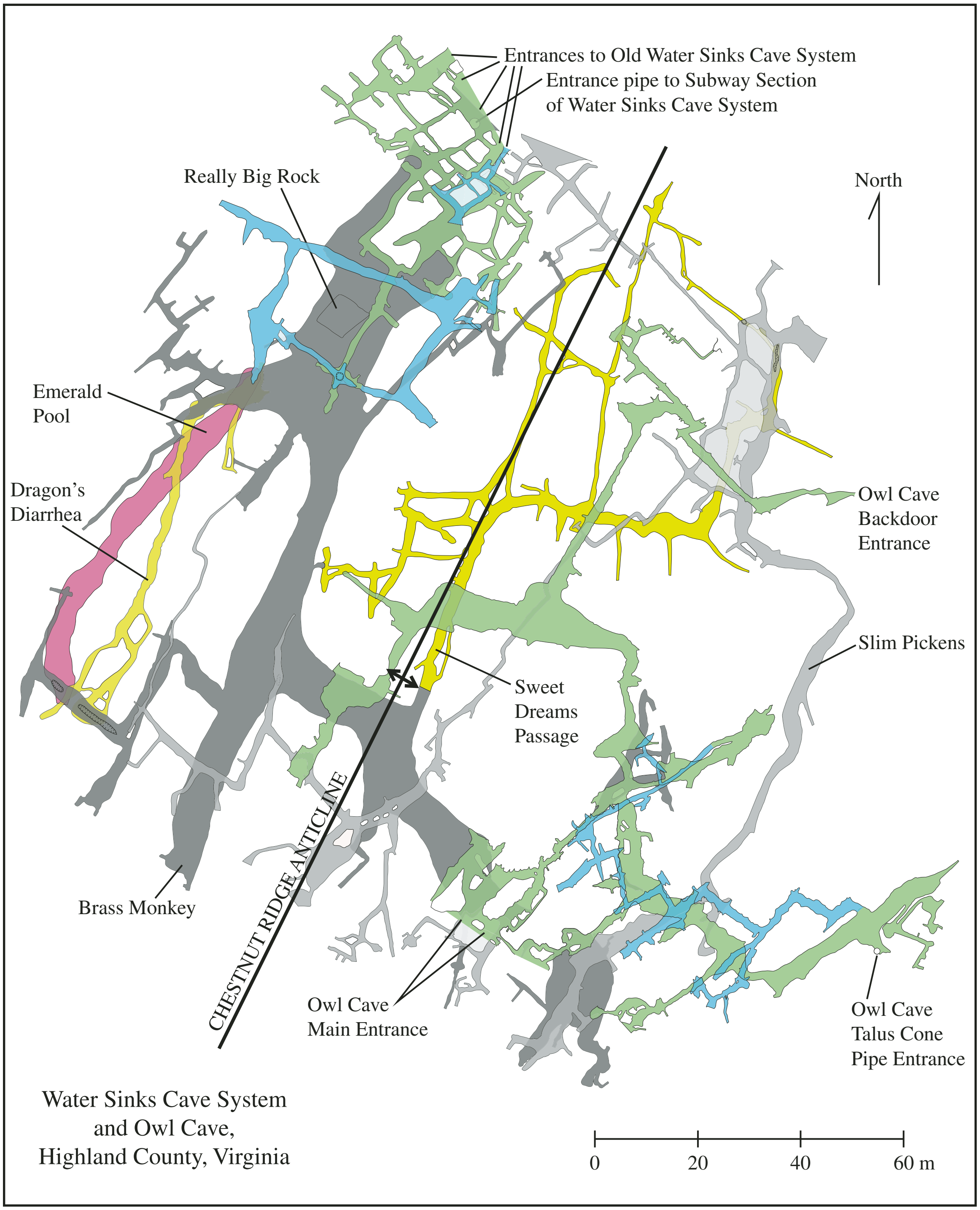


Figure 31. Map of Water Sinks Cave System and Owl Cave (modified from maps associated with Lucas, 2015a) and approximate location of the Chestnut Ridge anticline. Color code is the same as in Figure 26.

Figure 32. Corals and stromatoporoids in bioherm within the Jersey Shore Limestone Member of the Keyser Limestone, just inside the main entrances to Owl Cave, Highland County, Virginia. Photograph by C.S. Swezey.

Member of the Keyser Limestone and in the overlying New Creek Limestone.

Stop 3-4: Helictite Cave

Helictite Cave is located on the north-northeast side of the Water Sinks depression (Fig. 2). The cave was discovered by digging at the bottom of a fissure from which cool air was blowing, an endeavor that was begun in 1988 by Phil Lucas and many others, although discovery of the cave did not occur until 1996 (Lucas, 2015b). As of 2015, Helictite Cave had a total known passage length of 11,394 m (7.08 mi).

As with the Water Sinks Cave System, Helictite Cave is located along the axis of the northeast-trending Chestnut Ridge anticline (Fig. 33). The beds dip away from the anticline axis at relatively low angles, and many of the cave passages have flat ceilings defined by bedding planes. Maze patterns are also common in parts of Helictite Cave. Such maze patterns in caves are often located along or near the crests of anticlines (e.g., Palmer, 1975; Swezey, 2015).

Helictite Cave has two distinct levels that are located in separate stratigraphic units (Lucas, 2015b). A lower level passage (the Streamway) trends northeast and is located within limestone beds of the Jersey Shore Limestone Member and LaVale Limestone Member of the Silurian to Devonian Keyser Limestone, the Devonian New Creek Limestone, and the Devonian Corriganville Limestone. In contrast, much of the rest of the cave is an upper level set of maze passages that are located within the Devonian Shriver Chert and (or) Licking Creek Limestone. The cave entrance is also located within the Shriver Chert and (or) Licking Creek Limestone (Fig. 34).

Stop 4: Aqua Spring, Highland County, Virginia (N 38° 12′ 55″, W 79° 35′ 40″)

This is property of the Virginia Department of Game and Inland Fisheries (DGIF). Please obtain consent from the DGIF before proceeding onto the property.

There are several distinct karst drainage basins within Burnsville Cove, and much work has been done using dye traces to delineate the extent and boundaries of these drainage basins (e.g., Davis and Hess, 1982; Harmon and Hess, 1982; Davis, 2015). Many of these drainage basins send water to outlet springs in and around the Bullpasture Gorge (Fig. 2). In 1956, dye traces by Holsinger (1961) demonstrated that Aqua Spring (Fig. 35), which is the entrance to Aqua Cave, is the resurgence of water from Sinking Creek in Butler Cave (Fig. 11). As described by Holsinger (1961, p. 13), his study

> ascertained a connection between Sinking Creek in Butler Cave and the resurgent stream in Aqua Cave. … Several conclusions may be reached regarding these test results. It appears that Sinking Creek, from its siphon point in Butler Cave, must flow for at least three miles northeast and under Chestnut Ridge in order to reach its point of resurgence in Aqua Cave. It seems rather obvious the Sinking Creek is only one of several feeder streams for the Aqua Cave water when one takes into account the fact that approximately eight gallons of water flow from Aqua Cave per second while only between one and two gallons flow into the Sinking Creek siphon per second. The change in elevation between these two points is between 300 and 400 feet. The time that it takes for water to flow between these two points falls somewhere between seven and fifteen days as indicated by the test data.

Subsequent dye traces have confirmed that water from Butler Cave, as well as water from the Water Sinks Cave System and from several other caves in Burnsville Cove, flows to Aqua Spring (Davis, 2015; Lucas, 2015a). Furthermore, at least three other major springs have been identified along the Bullpasture River in Bullpasture Gorge (Davis, 2015). All of these springs drain portions of the limestone of Burnsville Cove.

End of field trip. Begin drive back to Richmond.

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The authors extend sincere thanks to the Butler Cave Conservation Society (BCCS) for encouraging the scientific study of caves and karst in and around Burnsville Cove, and for permitting access to Butler Cave for this field trip. The authors also thank Nevin W. Davis for his generous assistance in the field over many years, and numerous geology students from James Madison University (JMU) for their help since 2009. Christopher S. Swezey thanks Gregg Clemmer (BCCS) and Tony Canike (BCCS) for general encouragement with this project. John T. Haynes acknowledges support from the U.S. Geological Survey (USGS) National Cooperative Geologic Mapping Program under USGS EDMAP award numbers G09AC00122, G11AC20278, and G12AC20312 for bedrock mapping in

Figure 33. Map of Helictite Cave (modified from maps associated with Lucas, 2015b) and approximate location of the Chestnut Ridge anticline.

Figure 34. Phil Lucas beside the Devonian Shriver Chert and (or) Licking Creek Limestone above the entrance to Helictite Cave (from Swezey et al., 2015). Photograph by C.S. Swezey.

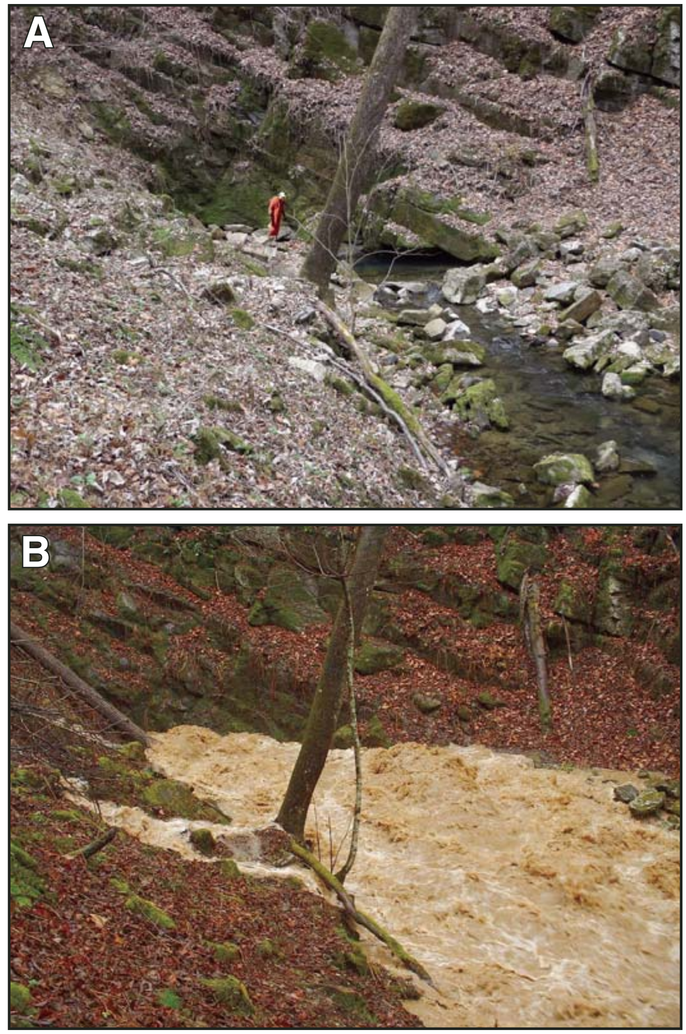

Figure 35. Aqua Spring, Highland County, Virginia. (A) depicts normal flow conditions (discharge rate of ~0.1–0.3 m^3 per second), whereas (B) depicts flood conditions (discharge rate of ~42 m^3 per second). The field of view is approximately the same in both photographs. Photographs by P.C. Lucas.

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