A review of the karst resources of the Antietam National Battlefield, the Harpers Ferry National Historical Park, and the Chesapeake and Ohio National Historical Park

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Executive Summary

Antietam National Battlefield (ANTI), Chesapeake and Ohio National Historical Park (CHOH), and Harpers Ferry National Historical Park (HAFE) share similar landforms and karst processes. These parks are within the Great Valley physiographic province and are at least partially underlain by Cambro-Ordovician carbonate (limestone and dolomite) formations. Lithological characteristics of these formations include high density of fractures, limited pore space, high percentage of dolomite, and flat to vertical dip of the bedrock. Regional strike is northeast-southwest and many mapped and unmapped faults and folds are present. The carbonate rocks contain karst landforms, particularly springs and caves and to a lesser extent sinkholes. Where soil is thin or absent and dip is steep, bands of exposed bedrock, termed 'striped epikarst' are present. This is particularly evident at ANTI.

The transmission of water through the carbonate aquifer is complex, dispersed, and sometimes radial or half-radial in nature. Half-radial flow is incompletely understood but is driven by a combination of base elevation, bedrock, structure and fracture density. Repeated tracer tests indicate that multiple springs can be influenced by a single source within an aquifer. Springs tend to be numerous and small. Springs arise not only at formation boundaries, but within formations, probably due to intra-formational lithological variation. Many streams within the area begin as springs; considerably less surface drainage occurs on carbonate bedrock.

The movement of water through the carbonate aquifer has enlarged a portion of the bedding planes and fracture networks to the size of traversable caves. The development of caves is not uniform across the Great Valley. The longest caves tend to develop in the Chambersburg Formation, Tomstown Formation and the St. Paul Group. Caves formed within the Great Valley tend to be of two varieties: single passages or network mazes. Maze caves are frequently developed under dispersed, low gradient conditions, whereas single passages often form from concentrated alloigenic recharge. Recently, a hypogenic origin (aggressive water rising from depth) has been suggested for these caves as well.
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Table 1: GIS Data Sources
Acronyms and Abbreviations

ANTI - Antietam National Battlefield
CHOH - Chesapeake and Ohio National Historical Park
C&O - Chesapeake and Ohio
GIS - Geographical Information System
HAFE - Harpers Ferry National Historical Park
MM - Mile marker (along the C&O Canal)
MYA - Millions of years ago
NTU - Nephelometric Turbidity Units (turbidity measurement unit)
USGS - United States Geological Survey
UTM - Universal Transverse Mercator
WVACS - West Virginia Association of Cave Studies
WVASS - West Virginia Speleological Survey
WVGES - West Virginia Geological and Economic Survey
1 Introduction and purpose

The karst of the Appalachian Great Valley is complex and varied along its entire length from New York to Alabama (Figure 1). Karst is a term applied to both a specific landscape and the processes required to create that landscape. Karst settings are most often found in areas underlain by carbonate rocks (limestone and dolomite). In eastern North America the karst landscape is dominated by springs, sinkholes and caves.

The Maryland - West Virginia subregion of the Great Valley, in which the study area is located, is comprised of Washington County in Maryland and Berkeley and Jefferson Counties in West Virginia. It is characterized by relatively flat terrain underlain by horizontal to vertical geologic units of Cambrian and Ordovician age. These formations are arranged in parallel north-northeast to south-southwest beds and consist of limestones and dolostones to form a karst landscape.

Appendix A contains a detailed description of lithological formations in context of the geologic time period.

The purpose of this project was to gather and combine geological information into a Geographical Information System (GIS) to create a better regional-scale understanding of the geology. The secondary goal was, based on the compiled maps and data, to develop a conceptual model for the karst system regionally and at three parks: Harpers Ferry National Historical Park (HAFE), Antietam National Battlefield (ANTI), and the Chesapeake and Ohio Canal National Historical Park (CHOH).

2 Karst systems: A brief review

2.1 Karst system components for a conceptual model

Generalized conceptual models for karst systems include both processes and surface-subsurface features. Processes include recharge, transmission, storage and discharge. The processes are linked to
specific surface features such as sinkholes, gentle subsidence features, springs, caves, and other subsurface dissolution passages. While no single model applies to all karst systems, there are several components that must be incorporated into any model.

Recharge. The water in karst aquifers is recharged from the surface by either rapid injection into the system (via sinkholes, along bedding planes, or via sinking streams) or more slowly via dispersed infiltration. A further distinction can be made between allogenic recharge (from outside the karst terrain) or autogenic recharge (from direct precipitation onto the karst terrain). Autogenic recharge tends to be dispersed (from precipitation) or locally concentrated (through sinkholes). The primary source of allogenic recharge is concentrated stream water reaching the karst landscape from neighboring elastic (non-karst) rocks. In practice, the recharge to most karst systems is a mixture of sources and input types (Ford and Williams, 2007).

Subsurface Transmission of Water. Groundwater is stored in voids in the bedrock. In karst systems, these voids have been classified into three categories by size (a) microscopic voids between grains (b) fractures (c) solution-developed conduits within the rock. This transmission of water through a karst aquifer has been referred to as the “triple permeability model”. The scale of the triple permeability model encompasses all voids from large open caves (a traversable conduit) to pores between rock grains (Figure 2). Conduits are the largest openings and allow for quick transport of large volumes of water at high velocities through the system to the spring. The minimum size for a conduit has been suggested to be 1 cm (White, 1988). It is at this aperture that the water flow becomes turbulent and is capable of moving fine sediment.

Fractures are smaller than conduits and range from 0.1 to 1 mm. Water is transported more slowly from the fracture network than the conduit network. Depending on the rock type, fractures can play a significant role in the transmission of water. They may also play a large role in the storage of water.

Bedrock porosity is the storage space between individual grains of rock, sometimes referred to as "primary porosity" or “matrix porosity”. This represents the largest potential storage reservoir in a karst aquifer. It is also the most difficult reservoir to access. Water trapped in the bedrock matrix is released very slowly and accounts for most, if not all of the water discharged from springs during drought conditions. The porosity (the available space for water to occupy) of the bedrock matrix varies by lithological unit, but as a rule, the older and more deformed the unit, the smaller the available space between grains of rock (Ford and Williams, 2007).

Aquifer Storage Capacity. The distribution of water between the three types of aquifer permeability controls the total volume of water stored in the aquifer. While scientists know that karst aquifers have
tremendous storage capacities, little is still understood about its role in the overall system or how to quantify the storage capacity.

**Epikarst.** Where soluble bedrock is at or near the surface, chemically aggressive water infiltrates into the subsurface, creating a network of enlarged joints and fissures near the soil-rock interface. This region along the soil-rock interface is called the epikarst. Sinkholes are a common surface feature that connects the surface to the epikarst. The jagged epikarst interface can frequently be observed in road cuts and quarry walls. In the Great Valley the epikarst is overlain by a thin soil layer. CO₂ within the soil (from plant respiration) enhances the amount of CO₂ in water as it percolates down to the water table. This increased CO₂ is instrumental in increasing the dissolutive capacity of water.

**Geochemistry of Karst Waters.** The geochemistry of karst water is dominated by the presence of the calcium plus carbonate ions in the form of H₂CO₃⁺ (dissolved CO₂), HCO₃⁻, or CO₃²⁻. The total inorganic carbon is the sum of the carbonate species; it is introduced via this generalized process:

1. Rainwater percolating through the soil layer becomes enriched in CO₂, forming H₂CO₃⁺ (carbonic acid plus dissolved CO₂) as per the following equation:

   \[ \text{H}_2\text{O} + \text{CO}_2 \rightarrow \text{H}_2\text{CO}_3^+ \]

   H₂CO₃⁺ naturally dissociates in water releasing ions of hydrogen (2H⁺). This is the source of the acid necessary to dissolve the limestone (comprised of the mineral calcite).

2. The acid rich water interacts with and dissolves some of the surrounding limestone as it percolates to the water table. This water is referred to as “aggressive” or “undersaturated with respect to calcite”. This is one of the means through which vertical fissures and sinkholes become enlarged.

3. Once the water reaches the water table, the aggressive water percolating downwards mixes with water already in the aquifer which is typically less aggressive. Once the acid is consumed, the water is no longer able to dissolve limestone.

**Karst Dissolution** - The degree to which water can dissolve limestone is controlled by the amount of available CO₂ to the system. This partial pressure of CO₂ (P_CO₂) varies throughout a karst system. At the surface, the P_CO₂ is equivalent to the atmosphere. As the water passes through the soil layer and to the water table, the P_CO₂ increases and the same volume of water is capable of dissolving additional limestone. When downward percolating water encounters an air filled chamber (i.e. a cave) it degasses, and calcite is precipitated as a cave formation. If degassing takes place at a spring the calcite is deposited on the surface and is called tufa or travertine. There are examples of tufa deposits at MM 92 along the C&O Canal (Southworth, 2008).

Most carbonate waters share similar geochemical characteristics. These have neutral to alkaline pH, moderate amounts of dissolved CO₂ and high concentrations of Ca²⁺ and bicarbonate (HCO₃⁻) ions as a result of the dissolution of limestone. If dolomite is present, moderate amounts of Mg²⁺ ions will also be present in the water. Specific conductivity, a measure of the electrical conductance of water, increases as more ions are present in solution. In a carbonate system, the measure of specific conductivity can be a good proxy for the total amount of dissolved chemicals in the water.
2.2 Introduction to karst system conceptual models.

A conceptual model of a karst system begins with an understanding of the inputs (recharge) and outputs (discharge) and how they are interrelated spatially and by process (Figure 3). Of the inputs, conduit input is strongly associated with surface water entering through sinking streams. The fracture-matrix systems, in turn, are associated with dispersed infiltration from autogenic recharge. The transmission rate for conduit water is several orders of magnitude quicker than the fracture-matrix water. Both components, however, are present in nearly all karst aquifers.

The combination of recharge, transmission and storage control the overall water flow system. Because the processes occur in series, not in parallel, one slow process can limit water conveyance through the system. For example, Scanlon and Thrailkill (1987) found that the density of sinkholes in Kentucky’s Inner Bluegrass Region controlled the discharge of springs during storm events.

Figure 3: Conceptual schematic of a karst system, showing various inputs and outputs (White and White, 2003).

Epigenic and hypogenic cave formation. The cave processes previously described fall under the broad category of epigenic caves. A cave is considered to be epigenic when the acid that drives cave development is derived from surface or near surface processes (Palmer, 2007). In hypogenic caves, however, the source of the acid-bearing water rises from depth. The process of cave development is reversed, with the lowest strata affected first and the surface affected last, if at all. As many as 15% of all existing caves may be hypogenic in origin. The idea of hypogenic cave development is fairly recent and first gained widespread interest in the U.S. in the 1970s and 1980s (Klimchouk, 2007). Hill (1987) proposed a rising sulfuric-acid speleogenesis for the caves of Carlsbad Cavern National Park. Recently Doctor et al. (2008) suggested a hypogenic origin for some of the caves of the Shenandoah Valley.

Hypogenic caves tend to be dominated by maze patterns and have little relationship to surface topography, making them difficult to locate. Additionally, established hypogenic caves may be subsequently modified by epigenic processes as topography is lowered, masking their hypogenic origin and making identification difficult. As a result, many caves which were formally thought to be exclusively epigenic in origin are being reevaluated from an hypogenic perspective (Klimchouk, 2007).
3 Study approach and resources

3.1 Definition of the study area

The study area covers the following national parks: Antietam National Battlefield (ANTI), Harpers Ferry National Historical Park (HAFE) and the Chesapeake and Ohio National Historical Park (CHOH) (Figure 4). The study area is underlain primarily by carbonate bedrock, with clastic (non-carbonate) sedimentary rocks of secondary importance. Precambrian crystalline rocks exist directly to the east of the study area. The eastward flowing Potomac River represents the primary drainage. Secondary drainage is from north-south trending streams: Opequon Creek, Antietam Creek, Conococheague Creek and Back Creek.

3.2 Geological resources

For this project, geological descriptions were compiled from the following published maps and guidebooks:

Maps

- Geology of the Harpers Ferry Quadrangle, Virginia, Maryland, and West Virginia (Southworth and Brezinski, 1996) (Map and folio)
- Geologic map of the Frederick 30’ x 60’ quadrangle, Maryland, Virginia, and West Virginia (Southworth, 2007),
- Geology of the Hedgesville, Keedysville, Martinsburg, Shepherdstown, and Williamsport quadrangles, Berkeley and Jefferson Counties, West Virginia (Dean et al., 1987).
- Keedysville, Shepherdstown, Harpers Ferry, Charlestown Quadrangle Geologic map (Brezinski, 2009)
- Geology of the Berryville, Charles Town, Harpers Ferry, Middleway, and Round Hill quadrangles, Berkeley and Jefferson Counties, West Virginia (Dean et al., 1990)

Guidebooks and Geology References

- Geology of the Chesapeake and Ohio Canal National Historical Park and Potomac River corridor, District of Columbia, Maryland, West Virginia, and Virginia (Southworth, 2008)
- Caves of the Eastern Panhandle of West Virginia (Gulden and Johnsson, 1984)
- Karst Hydrogeology of the Hagerstown Valley, Maryland (Duigon, 2001)
- Jefferson, Berkeley, and Morgan counties (Grimsley, 1916)
- The history and geology of the Harpers Ferry region, West Virginia, Maryland, and Virginia (Southworth and Brezinski, 2008)

3.3 GIS and mapping

Existing GIS data were provided by the National Park Service and were supplemented with data from additional sources (Table 1). Each organization has its own criteria for developing its GIS data and these do not always coincide. Additionally, each organization differs in what information it makes public and whether the information is freely available.
Figure 4: Overview map of the study area showing bedrock types: clastic (light grey) and carbonate (dark grey). Parks labeled by acronym.
Geological mapping standards vary from state to state, resulting in discrepancies in geologic units across state lines. Consequently, there are locations along political boundaries where the lithology appears to change because the corresponding state surveys interpreted or extrapolated the geology differently. Without fieldwork (beyond the scope of this project) to rectify these conflicts, it is impossible to determine which map is more accurate. Since most of these discrepancies occur on the sub-formation level, many of them were rendered moot by only showing formation or coarser divisions (Figure 5). Thus accuracy is sometimes attained at the expense of detail.

A subset of this problem is that formations do not remain laterally constant throughout their extent. Units which may be obvious at one end of a mapping area may be indistinguishable at the other end. This ‘facies change’ can render some units mappable only in portions of a mapping area. The Black River and Trenton Formations are an example. Within park boundaries they are mapped as the Martinsburg and Chambersburg Formations, but south of the parks they are mapped separately. This problem becomes more common as the mapped area increases.

Table 1: GIS Data Sources

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<thead>
<tr>
<th>Source</th>
<th>Location</th>
<th>Comments</th>
</tr>
</thead>
<tbody>
<tr>
<td>West Virginia GIS Technical Center</td>
<td>Department of Geology and Geography in Morgantown WV <a href="http://wvgis.wvu.edu/">http://wvgis.wvu.edu/</a></td>
<td>Free data</td>
</tr>
<tr>
<td>West Virginia Geological and Economic Survey (WVGES)</td>
<td>Morgantown, WV <a href="http://www.wvgs.wvnet.edu/">http://www.wvgs.wvnet.edu/</a></td>
<td>Charge for data. Sometimes georeferenced to older standards</td>
</tr>
<tr>
<td>Maryland Geological Survey (MGS)</td>
<td>Baltimore, MD <a href="http://www.mgs.md.gov/">http://www.mgs.md.gov/</a></td>
<td>Charge for data</td>
</tr>
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Figure 5: Geological formations in and around the study area. Lithological descriptions are included in Appendix A with the exception of the Trenton and Black River Formation. This unit represents a facies change from the Chambersburg and Martinsburg units and is only mapped south of the parks.
Cave maps and databases are less accessible than general geologic information because of concerns for safety, vandalism, trespassing issues and landowner relations. A cave database for West Virginia has been maintained by the West Virginia Association for Cave Studies (WVACS) since the 1960s and is the source of the West Virginia data. Additionally, Caves of the Eastern Panhandle of West Virginia (Gulden and Johnsson, 1984) was published by the West Virginia Speleological Survey (WVASS) compiling the then current data into one volume. At least one large cave system has been surveyed since that publication. Geospatial data for West Virginia caves are saved as UTM coordinates.

There is no similar organizational group for Maryland. The most recent information on Maryland caves was published in the Caves of Maryland (Franz and Slifer, 1971). Within that publication, cave locations are described as a series of repeated ninefold division within a named quadrangle. This method preserves the secrecy of the cave location at the expense of accuracy as the smallest grid is a rectangle several hundred feet to a side. A spreadsheet was constructed to locate the center point of an individual rectangle and assign it to a cave. This provides an estimate of cave locations in Washington County.

Due to non-systematic investigating, data from cave surveys are frequently clustered. As a result, certain lithologies may be disproportionately represented. In terms of average cave lengths, the mapping (or failure to map) one or more long caves may significantly skew the data. Cave data are, at best, a non-random sample of the overall population of conduits within a karst aquifer.

4 Regional geologic setting

4.1 Regional geology and geologic history

The Great Valley is a long lowland area stretching from the Champlain Valley in New York to the Coosa Valley in northern Alabama (Figure 1). CHOH, ANTI and HAFE lie at the border of two subdivisions of that valley – the northern Cumberland Valley and the southern Shenandoah Valley. The Cumberland Valley extends from the Potomac River northward to Harrisburg, PA and the Susquehanna River. Locally, the Maryland portion of the Cumberland Valley is known as the Hagerstown Valley. The Shenandoah Valley extends southward from the Potomac River to the James River in southern Virginia.

The rocks comprising the Great Valley are a Paleozoic sedimentary sequence of repeated carbonate and clastic rocks representing warm, shallow seas similar to the present Caribbean. The oldest rocks are the sandstones and conglomerates of the early Cambrian Period (~540 MYA) Harpers and Antietam Formations (Figure 5, Appendix A). These were deposited off the slopes of mountains to the west. By the middle Cambrian Period (~520 MYA) the western mountains had eroded away, leaving a warm, shallow sea in its place. Deposited into this sea were the limestones and dolostones of the Tomstown Formation. Immediately overlying them were interbedded carbonates and clastic rocks of the Waynesboro Formation followed by the deposition of a thick series of carbonates (the Elbrook and Conococheague Formations) at the end of the Cambrian Period (490 MYA). Carbonate deposition continued throughout the early to middle Ordovician Period (490-450 MYA) in the Beekmantown Group, St. Paul Group and the Chambersburg Formation. However as mountain building to the east neared, large amounts of shale were deposited from the encroaching mountains as the Martinsburg Formation (450 MYA). Further Paleozoic deposition consisted of clastics and carbonates which are not represented in the study area.

The Great Valley is bordered on its eastern and western edges by two massive thrust faults that transported older rocks from the east. Although a geologic cross-section is not available to exactly match
the study area, several are available for the larger region and provide an example of the general regional geology. Structurally, the North Mountain Fault forms the western boundary of the Great Valley and the Keedysville Fault forms the eastern boundary. (Figure 6, upper cross-section). As each event moved rocks west, strata from the east ramped over emplaced strata, repeating the sequence and thickening the crust (Figure 6, lower cross-section). Within the valley, numerous smaller faults and folds run in parallel from the northeast to southwest. As a result, many formations dip steeply to the southeast or northwest or are completely vertical. Geomorphic and hydrologic features preferentially orient along this northeast – southwest line. As research continues there is potential for uncovering additional faults and refining the location of currently known faults. With respect to water flow, the faults are not consistent. Some brecciated zones within faults can act as a preferential conduit for water, dramatically decreasing travel time. Other brecciated zones have been cemented shut through the deposition of calcite or other minerals. When this occurs the fault acts a barrier to flow and a potential drainage basin boundary.

Figure 6: Two cross-sectional views through the Great Valley. Both views are looking north. The upper view is the Great Valley along the C&O Canal route (Southworth, 2008). The lower view is through Berkeley and Jefferson Counties, WV (Dean et al., 1987). The lower view shows the deeper structure under the valley, with multiple thrust faults underlying the current topography.
4.2 Karst features

Springs are common in the region (Figure 7). Springs in the Great Valley can range from small seeps discharging much less than 0.001 m³/sec of water to enormous springs discharging hundreds of m³/sec of water. The largest spring in the Hagerstown Valley in Maryland is located east of Hagerstown and just north of the intersection of I-68 and County Highway 66. It discharges about 0.14 m³/sec (Duigon, 2001). The largest spring in West Virginia (Priest Field Spring 2.4 km SW of Middleway WV in the Chambersburg Formation) discharges around 0.4 m³/sec (Gulden and Johnsson, 1984).

In Berkeley and Jefferson Counties WV, most small streams on karst begin as springs and seeps, and this is probably the same in Washington County. That many of the springs arise in the middle of the carbonate exposures suggests that the water table is close to the surface and small changes in elevation or lithology are sufficient to bring water to the surface. This is supported by aquifer mapping reported in Kozar et al. (1991).

Sinkholes in the study area were inventoried by Duigon (2001). Duigon recorded 93 sinkholes on topographic maps in the Hagerstown Valley, ranging in size from under 30 m to over 300 m. He noted that sinkhole distribution was not linear, but clustered within specific units and at specific locations within those units. Of these, the Chambersburg Formation west of the Conococheague Creek had the most notable arrangement of sinkholes (Figure 5). These were arranged linearly, along strike, following strata conducive to sinkhole development. This sinkhole development can be seen to continue across the Potomac River into West Virginia. At least one sinkhole near the river has developed into a cave. This is Eby Cave which is described as a 7 m pit leading to an impassable hole.

Conduits and Caves are the primary subterranean expression of the karst terrain. The relationship between cave patterns and geological setting was defined by Palmer (2007). In fracture-dominated rock, Palmer defined differences in cave patterns by recharge - concentrated or dispersed.

Two caves in the area, both from the Tomstown Formation, illustrate these differences. The first example is Crystal Grottoes Commercial Caverns (Figure 8), located about 4 km northeast of ANTI. Crystal Grottoes is an excellent example of a maze cave formed along three dominant fracture sets - a north-south fracture set, a N30E fracture set and a N30W fracture set. Bedrock dips 20° here and strikes along the N30E fracture line. Passages are typically tall fissures with small rooms where passages meet. There is no flowing water present in the cave. However, water from ceiling drips accumulate quickly after precipitation events and pond in portions of the cave for considerable lengths of time (Franz and Slifer, 1971). Maze caves have been interpreted to have formed in systems with diffuse ground water flow, dispersed infiltration, along river banks, where dip is low, or where there is little hydraulic gradient.

The second example is Mt. Aetna Cave (Figure 9) about 15 km northeast of ANTI. This cave is also in the Tomstown Formation. Dip here is similar to Crystal Grottoes at 23° and strike is at N52E. This cave, however, is a single passage generally following strike. Passage dimensions are wide, with considerable breakdown lining the floor, suggesting this cave may be older. While no stream is reported in the cave, the presence of breakdown would obscure any stream morphology. Abundant speleothems indicate recharge from the surface was frequent after the passage was created. Linear caves have been interpreted to be the result of concentrated recharge, either through sinkholes or from streams sinking from nearby clastic rocks.
Figure 7: A map of caves and springs in or near the study area (no spring data were available for Maryland). The greater number of small streams in West Virginia is illustrative of the different mapping standards between states. Note also the scarcity of streams over carbonate bedrock as a byproduct of karstification.
Figure 8: Crystal Grottoes Commercial Caverns, an example of a maze cave (Franz and Slifer, 1971).

Figure 9: Mt. Aetna Cave. An example of a linear cave (Franz and Slifer, 1971).
In an effort to better quantify cave development within the area, some basic statistical analyses were performed. Because the caves within park boundaries represent an insufficiently large sample, all the caves within Berkeley and Jefferson Counties in West Virginia as well as Washington County, Maryland were considered. Data were acquired from *Caves of Maryland* (Franz and Slifer, 1971) and *Caves of the Eastern Panhandle* (Gulden and Johnsson, 1984). Only formations older than the Martinsburg Shale were considered, removing some of the westernmost caves in the CHOH.

Three formations account for 86% of the carbonate rocks in the Great Valley. These are the Beekmantown Group, the Conococheague Formation and the Elbrook Formation (Figure 10). Normalizing the number of caves and aggregated cave lengths by the area underlain by each formation allows for a comparison of cave distribution. Without this normalization, a poor cave-forming unit with many small caves appears more significant than a good cave forming unit.

![Figure 10: Percent of total carbonate area covered by individual formations.](image)

Because cave entrances are largely random, total length of surveyed passages is a more revealing statistic than number of cave entrances. It is useful to compare aggregated passage length for each formation against the normalized passage length by formation (Figure 11). Formations whose average length decreases when the total length is normalized by area have many short caves and very few long caves. This may be a result of heterogeneity within the lithology (lenses of more soluble bedrock are surrounded by relatively insoluble bedrock). Formations whose average length does not decrease when total length is normalized by formation area have relatively few caves, each of considerable length. The data show that the formations comprising the largest areas are also the three formations with the shortest average cave lengths. This suggests that these units do not exhibit many karst features over a large area. This agrees with the lack of observable caves and karst features over much of the Great Valley.
Cave systems may also be classified by pattern. The simplest division is to determine whether the majority of the cave is either (a) mazelike or (b) linear or branchwork. In a comparison of cave length per km$^2$ of formation separated by passage patterns the Chambersburg Formation and the St. Paul Group have the greatest length/km$^2$ (Figure 12). This agrees well with Duigon's assessment that the Chambersburg Formation is the most karstic by virtue of the greatest number of sinkholes per km (Duigon, 2001).

Figure 11: Distribution of cave lengths for individual formations (left) and normalized for area underlain by the formation (right).

Figure 12: A comparison of total length by passage pattern of caves (left) and total length by passage pattern of caves per km$^2$ (right) of formation.
However, different cave patterns are associated with different rock units (Figure 12). The Chambersburg Formation almost exclusively forms maze caves while the caves of the St. Paul Group are more linear. Another formation of interest is the Tomstown Formation. It also forms maze caves. However, these maze caves tend to be much shorter than caves in the Chambersburg Formation. Lastly, the importance of the Waynesboro Formation is overestimated by its small sample size.

ANTI and HAFE are underlain only by the Tomstown and Waynesboro Formations while CHOH is underlain by all four formations (Figure 13). It should be noted that the formations which appear to be the most cavernous are also the formations located adjacent to clastic rocks, suggesting that the influence of allogenic water onto karst terrains is an important factor in conduit development. Differences in lithology can only account for some of the variation in passage development. The two previous examples, both from the Tomstown Formation, illustrate this.

5 Karst geology at the parks

5.1 Antietam National Battlefield (ANTI)

ANTI is mostly underlain by the Elbrook and Conococheague Formations, and to a lesser extent, the Tomstown and Waynesboro Formations. The Elbrook and Conococheague Formations are among the least cavernous in the Great Valley (Figure 13). Duigon (2001) notes the Conococheague Formation in particular for having few sinkholes per km². Being farmland, sinkholes may be filled in quickly. This would be an effective strategy if the sinkholes did not transmit soil into the conduit system particularly quickly. Weeks (2002) reports two sinkholes in the West Woods reforestation area which have been used as dumping grounds by locals. Jones (2009) reports at least one small sinkhole in the North Woods. The difference in land use between forests and farms may account for the sinkhole distribution.

At ANTI, exposed bedrock illustrates the differences in epikarst development between and within formational units (Figure 14). Bedrock is exposed in long, thin parallel ridges, giving the landscape a striped appearance, particularly from the air. Informally known as "striped epikarst", it is most prominent in areas where dips are steep. Where dips are shallow the same unit is exposed over a wider area and the striping is absent.

Currently 10 springs have been mapped at ANTI (Figure 15). An older study by Duigon and Dine (1991) reported five in or near the battlefield. Of these 10, seven support streams. The most historically significant of these appears to be Mumma Spring with a maximum recorded discharge of 0.006 m³/sec (Weeks, 2002). Maximum reported turbidity for Mumma Spring from 1985-1988 was 2.7 NTU while the mean turbidity was 0.813 NTU (NPS Water Resources Division, 1995). Other springs on the Haines Tract and the Miller Farm are similar to Mumma Spring in that they are all located in relatively high topography. This suggests that either the water table is relatively close to the surface, that subsurface drainage throughout the battlefield is poor, or water flows upward under pressure. Drainage, however, is not so poor as to form wetlands. Two small areas designated as wetlands were noted by Weeks, one at the Piper Farm and the other at Mumma Spring. Neither are extensive and the only other wetland areas within the park are in the southeast corner along Antietam Creek.
Figure 13: Distribution of the carbonate Chambersburg, St. Paul, Tomstown and Waynesboro lithologies throughout the study area.
Figure 14: Parallel ridges ("striped epikarst") of the Conococheague Formation just west of the Observation Tower and north of the Piper Farmhouse. The road at left is MD 65, while the road at top is Richardson Avenue. The parallel lines of exposed bedrock occur by the preferential dissolution of some beds over others. Average dip here is approximately 75° to the southeast. The NW - SE break in bedrock exposure is a shallow valley created by the upstream extension of a small stream which flows into Antietam Creek. (Google Earth, 2011)
Figure 15: Map of the Antietam area. The red box marks the location of Figure 14.
5.2 C&O Canal National Historical Park (CHOH)

CHOH is a linear park crossing several physiographic provinces and many geological formations. The section of park within the study area is approximately 77 km long. Every carbonate unit in the Great Valley crosses CHOH at least once, giving it the greatest geological diversity of the three parks. Duigon and Dine (1991) report 15 springs along the C&O Canal towpath in the study area, with additional springs nearby. Two factors contribute to this. First, steep bluffs are frequently found along the Potomac River. With the sudden change in elevation, ground water comes to the surface as a spring. Second, the park’s long length along the Potomac River assures there are numerous opportunities for springs to develop.

As it follows the Potomac River, the park intersects many drainage basins, but completely occupies very few. The only location within CHOH where an entire surface basin is within park boundaries is at MM 109. There the canal cuts across a large meander loop in the Potomac River, with the NPS owning the entirety of the loop. This area is underlain by the Elbrook and Conococheague Formations. Recharge from this area is very likely to be localized and autogenic, providing an opportunity to observe an entire drainage basin within the park.

The remainder of the CHOH receives its source water upgradient from park property. Most subsurface pathways feeding CHOH springs are expected to be either fracture networks or immature conduit systems slowly moving water through the aquifer. This slow movement makes tracer testing difficult in many places, and other evaluative techniques, such as comparative biological diversity, may be more appropriate to delineate basin boundaries. In some locations (between MM 88-89, for example) enough of the conduit system is observable to expect tracers to be successful.

Previous studies by both park employees and volunteers have inventoried some of the karst features, including the caves, in the park. Franz and Silfer (1971) included C&O Canal karst features in Caves of Maryland. Maps of several caves were also included. However, they are considered antiquated by today’s standards. Southworth (2008) mentions several caves and karst features in Geology of the Chesapeake and Ohio Canal National Historical Park and Potomac River corridor, District of Columbia, Maryland, West Virginia, and Virginia. GIS information about karst features have been compiled by park workers. Pervious biological studies have listed several environmentally sensitive locations within the park (Bartgis et al., 1993; Feller, 1994) (Figure 16).
Harpers Ferry National Historical Park (HAFE)

HAFE is located at the eastern edge of the Great Valley. The park is mostly underlain by clastic and metaclastic rocks of the Cambrian Period as well as metamorphic rocks of the Precambrian. Karst represents a small but significant portion of the total park area (Figure 17). Karst bedrock is exposed in two places west of Harpers Ferry - first, in the easternmost exposures of the Tomstown Formation and second in the Tomstown Formation and calcareous strata of the Waynesboro Formation on the recently acquired Bolivar Heights parcel. Gulden and Johnsson (1984) report two karst springs immediately northwest of HAFE and three non-karst springs on the eastern side of the Shenandoah River.

The easternmost exposure is noteworthy for being the location of John Browns Cave (Figure 17). The cave is at least 800 m long. At least two-thirds of the cave remain unmapped and only sketched. The vertical extent of the cave is about 50 m, with many domes, which is unusual for caves in this area. The back portions of the cave have either not been explored or visited only once or twice due to standing water blocking passages (sumps). Passages within the cave are linear and predominately developed as tall fissures. This is particularly so in the rear of the cave. The rear the cave is a single passage, often with a stream in the bottom (Gulden and Johnsson, 1984).
Figure 17: Map of the Harpers Ferry area. The red oval (further west) represents the likely source area for John Browns Cave. The purple oval represents the location of the resurgences for Ditmer Squeeze / Harpers Ferry Caverns.
John Browns Cave is developed in a narrow band of particularly cavernous bedrock in the Tomstown Formation that is less than 30 m thick and dips at about 67° to the east (Gulden and Johnsson, 1984). Nearly all passages in the cave follow strike. Within the cave, the numerous sumps or near sumps occur in ‘phreatic loops’ - similar to U-bends in plumbing. Phreatic loops are developed beneath the water-table. As flow paths develop, they form most easily along particularly favorable fracture-sets, dipping deep along one fracture and then rising along another. The frequency and size of these loops is determined by the frequency of the fracture set distribution (Figure 18). The more frequent the fractures, the more numerous (and the smaller) the number of loops. Piracy of a surface stream (as in John Browns Cave) can remove many of the loops, leaving only the largest loops in place (Ford and Ewers, 1978). The source of the stream is unknown. Because of the linear nature of the flowpath, the source is likely either on Bolivar Heights or the adjacent properties on Prospect Ave.

The second karst area at HAFE is the recently acquired property along Schoolhouse Ridge. This land is underlain by cavernous units in the Waynesboro Formation. One of the properties acquired was the briefly commercialized "Harpers Ferry Commercial Caverns", that operated throughout the 1960s and 1970s as a tourist cave (Figure 17). This cave has ~230 m of natural passage. The lower levels of the cave have an intermittent or possibly permanent stream. The total extent of the lower levels is unknown, and the entire cave is unmapped, only sketched. Water from this cave along with the neighboring Ditmer Squeeze Cave (located just off HAFE property) likely resurge at springs along Elk Run to the north (Gulden and Johnsson, 1984).

Figure 18: The effect of fracture density on steeply dipping rock, resulting in phreatic loops. Diagonal downward-pointing dashed lines indicate preferential fracture sets. John Browns Cave would be classified as type (b) or (d) (Ford and Williams, 2007).
6 Conceptual model for the study area

The development of a regional conceptual model of karst evolution in the Great Valley remains a work in progress. In the past, karst development in the Great Valley has been compared to other karst regions with varying degrees of success. However, as more data are gathered it becomes apparent that this is a unique area.

The main constraints on karst development appear to be structure, fracture frequency, and flatness of the water table.

- **Structure** - Passage development in caves is highly dependent on structure. Where dip is high, passages tend to follow strike and development is elongated along strike, as at Mt. Aetna Cave (Figure 9). Where dip is low, passages more frequently adopt a network maze pattern.

- **Fracture frequency** - A high density of fractures is a prerequisite for the formation of network maze caves and the radial or semi-radial drainage which occurs in Jefferson County (Jones, 1997). The long, straight passages in John Brown's Cave, with infrequent phreatic loops, also suggest a high fracture density (Figure 18).

- **Depth-to-water** - The water table is generally high in the area – as shallow as a few meters below the surface in some places (Duigon, 2001). This creates a limited range for open-air (vadose) karst development. A high water table also helps to support bedrock weakened by dissolution and prevents sinkhole formation (Palmer, 2007).

The following constraints apply to a lesser degree:

- **Proximity to the Potomac River** - The Potomac River has cut deeply through the current landscape, lowering the hydraulic base level. To compensate for the change in hydraulic gradient, springs exist on banks and sinkholes may form along the tops of the bluffs. Springs are most commonly at or just above river level. Under favorable conditions, springs, particularly overflow springs, can occur farther up the steep sides. Furthermore, existing phreatic loops in steeply bedded rock can result in springs issuing directly under the river. The change in base level has increased fracture density and spacing due to bedrock unroofing, resulting in many small caves along the river's bluffs.

- **Heterogeneity of lithology** - Most cavernous units in the Great Valley are laterally heterogeneous with respect to cave development. This commonly results in 'pockets' of caves throughout the area. Clusters of caves are also a statistical artifact of the exploration process as explorers unevenly search an area due to interest, landowner relations, et cetera. As the karst landscape matures, these isolated areas tend to grow and ultimately connect to one another, provided that karstification proceeds quicker than surface erosion.

- **Proximity to clastic rocks** - The addition of concentrated, allogenic recharge is a driving force in enlarging a limited number of flowpaths when present as well as connecting nearby isolated cavernous pockets. The vast extent of carbonate rock within the Great Valley suggests that the only locations where this would be significant is at the edges of the valley or in units adjacent to the insoluble Martinsburg Shale.
• **Radial Flowpaths.** While most flowpaths within karst terrains are dendritic (similar to surface streams), distributary (like a delta) and radial flowpaths can occur. Of these, radial flowpaths are the least common. In a radial flow network, water is equally likely to travel in any direction from its starting point. A subset of the radial flow network is the half-radial (or semi-radial) flow. In half-radial flow, water may travel in any direction in a semicircle from its origin.

Several dye traces conducted in Jefferson County, WV exhibit half-radial flow (Figure 19) in conjunction with very slow travel times (Kozar et al., 1991). Other traces show a bimodal radial movement, with fast movement along strike and much slower movement perpendicular to strike (Jones, 1991). Flowpaths between different tracer tests frequently overlap and arrive at the same springs; particularly along the western edge of the carbonate outcrops in the county.

Several questions regarding the conceptual model of the karst system remain. They include the following:

• Doctor et al. (2008) suggest the karst of the Great Valley, particularly in northern Virginia, has a hypogenic origin which has subsequently been masked by epigenic processes. One characteristic of hypogenic karst is the development of a network maze; another characteristic is the development of a conduit system independent of the surface topography.

• Biological evidence, including the range of the troglobitic Madison Cave isopod, *antrolana lira*, suggest that an extensive conduit system must have been in place for many millions of years, at least in the Great Valley in Virginia (Hutchins et al., 2010). The Madison Cave isopod has been observed in three sites in Jefferson County, West Virginia - two wells and one cave. All three sites have been in the eastern carbonate basin, east of the Martinsburg Shale. The nearest site is less than 10 km southwest of HAFE (Hutchins et al., 2010).

• White and White (1974) tabulated average cave lengths from Franklin County, PA to Rockingham County, VA. When combined with more recent tabulations, it shows a trend in increasing cave lengths southward. This maybe attributable to more favorable lithological composition to the south, more favorable structure, closer proximity to a hypogenic source, more mature karst or a combination of any or all of these factors. Regardless of the cause, this helps place the study area within a larger, regional context. Future tabulations should further refine this trend.

Our understanding of the Great Valley karst is continuously evolving and numerous questions remain. The conceptual model and issues discussed above are based on our current understanding of the system. Given the structural complex city of the region, it cannot be readily compared to better understood karst areas, such as in and near Mammoth Cave National Park in Kentucky or the Edwards Aquifer in Texas. What is clear, however, is the critical factor that the geologic structure and stratigraphy play in the distribution, nature and behavior of karst features in the Great Valley region.
Figure 19: Radial flowpaths of several dye traces in Jefferson County superimposed onto the bedrock geology. Different colored arrows represent different tracer tests. Arrows which do not appear at mapped springs likely appear at smaller, less well known springs. Modified from Jones (1997).
Bartgis, R., Feller, D., Simoes, C., Thompson, E., and Wiegand, R., 1993, State and nationally significant historical areas of the Chesapeake and Ohio Canal National Historical Park: in Resources, Maryland Department of Natural Resources, Annapolis, 56 p.


Feller, D., 1994, Aquatic subterranean macroinvertebrate survey of the C&O Canal National Historic Park in western Washington County, Maryland, in Resources, Maryland Department of Natural Resources, Annapolis p. 39.


Jones, W. K., 2009, Select field guides to cave and karst lands of the United States, Charles Town, Karst Waters Institute, 177 p.


Southworth, S., 2007, Geologic map of the Frederick 30' x 60' quadrangle, Maryland, Virginia, and West Virginia, USGS, scale 1:100,000
Appendix A. Detailed stratigraphy of the study area

<table>
<thead>
<tr>
<th>Formation &amp; Thickness</th>
<th>Description</th>
<th>Cave or Karst former?</th>
</tr>
</thead>
<tbody>
<tr>
<td>Martinsburg Formation</td>
<td>This formation ranges from southern New York to Virginia with its type locality in Martinsburg WV. The unit is a dark brown to black shale which weathers to a characteristic yellow brown. Thin beds of limey shale occur near the base of the formation and there is considerable shaley limestone in the upper end of the formation. Cleavage is prominent and obscures bedding and most fossils.</td>
<td>NO</td>
</tr>
<tr>
<td>Chambersburg Formation</td>
<td>This formation is dark grey, fine to medium grained, 1-3 cm bedded limestone with irregular parting planes which results in a cobbly or nodular appearance. This unit can be both silty and argilaceous. This unit is both fossiliferous with numerous guide fossils present in addition to bryzoans. The upper contact with the overlying Martinsburg Formation is gradational.</td>
<td>Excellent karst former</td>
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<tr>
<td>New Market Formation</td>
<td>This formation is an unusually pure (CaCO3 content &gt; 98%) dove colored to light grey massive fine grained limestone. The limestone can be mottled or laminated. Few fossils present – mostly corals and particularly <em>Tetradium syringoporoids</em>.</td>
<td>Excellent karst former</td>
</tr>
<tr>
<td>Ron Park Formation</td>
<td>This formation is only locally mapped near the Potomac River. In other places, it is generally mapped within the New Market formation. It consists of a massive and dove-colored to light grey limestone intermingled with a dark, granular impure and sometimes cherty limestone. The characteristic fossil in this formation is <em>Malcurites magnus</em>.</td>
<td>Poor karst former</td>
</tr>
<tr>
<td>Pinesburg Station Formation</td>
<td>This formation is a light grey, generally fine-grained but sometimes coarse-grained, laminated dolomite with abundant chert nodules. Chert nodules in the lower portion of this formation tend to be white in color. The number of thin beds of interfingered limestone increase eastward; these limestone beds have characteristics similar to the underlying Rockdale Run Formation and overlying New Market Formation suggesting the Pinesburg Station Formation may be a facies change. This formation is approximately 150 m thick at the western edge of the study area but thins eastward to about 30 m.</td>
<td>Poor karst former</td>
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<tr>
<td>Formation &amp; Thickness</td>
<td>Description</td>
<td>Cave or Karst former?</td>
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<tr>
<td><strong>Ordovician</strong></td>
<td><strong>Beekmantown Group</strong></td>
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<tr>
<td><strong>Rockdale Run Formation</strong></td>
<td>330-350 m</td>
<td>This bottom of this formation is composed of 80-100 m of predominately cyclic sequences of aphanitic limestones and dolostones. Immediately above this are layers with interbedded chert nodules which preferentially weather out. 30-60 m above the chert zone are layers of oolitic limestone with fossilized gastropod <em>Lecanospira</em>. Above this are 250 m of interbedded limestones and dolomites, with larger percentages of limestones at the bottom and larger percentages of dolomites at the top. The contact with the overlying Pinesburg Station Formation is placed at the top of the highest bed of limestone.</td>
</tr>
<tr>
<td><strong>Stonehenge Formation</strong></td>
<td>300-380 m</td>
<td><strong>Unnamed Upper Member</strong> – This member is frequently subdivided into 2-3 distinct lithologies. The top portion consists of 50-80 m of ribbony carbonates and massive algal limestone. Below that are about 150 m of ribbony carbonates interbedded with massive flat-pebble limestone conglomerates. This section shows up as a low ridge topographically. Below this is about 100 m of massive, fine textured algal limestone which is a valley former and weathers dark blue-grey. This trilobite <em>Bellafontia</em> is often found within this member.</td>
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<tr>
<td><strong>Stoufferstown Member</strong></td>
<td></td>
<td><strong>Stoufferstown Member</strong> – This member consists mainly of shaley, ribboney mechanically-derived and algal limestones. Silty and argillaceous laminar stand out in raised relief within this member. Thin beds of flat pebble conglomerate also appear in this member This unit is a low ridge former along with the top beds of the underlying Conocochague Formation. Beds are sometimes discontinuous. This member weathers to thin brown and orange chips. This is the oldest unit in which fossils of gastropods and orthocerids can be found. Thickness of this member is between 50-100 m, increasing in thickness to the north.</td>
</tr>
<tr>
<td><strong>Cambrian</strong></td>
<td><strong>Conococheague Formation</strong></td>
<td>Unnamed Upper Member – This member is subdivided into two sections. An upper portion (between 100–125 m thick) is composed of thinner, ribbon-like grey limestone intermingled with dolomite in beds 1-2 cm thick. This limestone is a fine grained aphanitic limestone with chert nodules scattered throughout the formation. This portion of the unit can be seen as a general rise in topography. The contact between the upper end of this unit and the overlying Stoufferstown Member of the Stonehenge Formation is transitional and usually placed at the base of the first beds to contain silty and argillaceous laminar in the limestone. The lower portion of this member is a coarse grained oolitic and fossiliferous limestone with occasional intraformational conglomerated throughout. Its thickness varies between 500-600 m. Chert occurs throughout this unit. A sandy limestone portion occurs somewhere in the upper half of this portion.</td>
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<tr>
<td>Conococheague Formation 570-670 m</td>
<td><strong>Big Spring Station Member</strong> – This member is approximately 70 m thick. It is a fine to medium grained limestone and dolomite. Occasional quartzose sandstone occurs throughout this unit (although never more than 10% of the total unit thickness). Sandy dolomite makes up about a third of the unit as well. The total amount of sand in this member decreases from west to east.</td>
<td>Poor karst former</td>
</tr>
<tr>
<td>Elbrook Formation 700-800 m</td>
<td>This unit is composed of two distinct lithologies. The lower portion of the Elbrook Formation contains interbedded thin to thick bedded tan limestones and dolostones, which frequently weather shaly. These are interbedded with medium bedded dark grey limestone. This portion is very poorly exposed. The upper portion of the Elbrook Formation is composed of cyclically bedded grey thrombolitic limestones and ribbon to laminated limestones and dolostones. This portion makes up the majority of the unit. The Elbrook is characterized by numerous sinkholes and small caves. Abundant algal stromatolites can be seen within the Elbrook, particularly on the south side of the Potomac across from MM 71.5 on the C&amp;O Canal.</td>
<td>Mediocre karst former</td>
</tr>
<tr>
<td>Waynesboro Formation 290-320 m</td>
<td><strong>Chewsville Member</strong> - This unit is composed of a reddish to chocolate brown silty shale or siltstone to fine grained silty sandstone interbedded with thin, white sandstone beds (2-6 cm thick) and a tan to buff medium-bedded sandy duotone. This is the most recognizable unit within the Waynesboro Formation and is a good ridge former. Burrows of <em>Skoilithos linearis</em> are frequently seen within the thin sandstone beds.</td>
<td>NO</td>
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<td><strong>Cavetown Member</strong> - This unit is composed of alternating layers of a medium to coarse grained, medium to thick bedded, intraclastic grainstone, a laminated dolostone, medium grey, oolitic lime grainstone, ribbony carbonates and burrow mottled dolomites. Although this unit makes up the largest thickness of the Waynesboro Formation, it is frequently poorly exposed.</td>
<td>Excellent karst former</td>
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<td><strong>Red Run Member</strong> - This unit is a punky, fine grained sandstone interbedded with shales, grey sandy limestones, laminated dolostones and locally thin (0.6 - 5 cm) layers of red sandstones and siltstones. The sandstones frequently weather tan. This unit is typically a low ridge-former, with considerable less relief than the Chewsville Member.</td>
<td>NO</td>
</tr>
<tr>
<td>Tomstown Formation 365-415 m</td>
<td><strong>Dargan Member</strong> - This unit is composed of cyclical limestones and dolomites. Some dolomites and limestones are medium to dark grey, bioturbated and laminated. Other dolomites are tan and silty.</td>
<td>Good karst former</td>
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<tr>
<td><strong>Cambrian</strong></td>
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<tr>
<td>Tomstown Formation</td>
<td>Benevola Member - This member is a white to light grey, massive to poorly bedded, sugary in appearance, highly fractured dolomite. Faint crossbeds are common, but frequently only visible in polished samples. The lower contact is a gradational continuation of the Fort Duncan Member with the distinguishing feature being the lack of burrow traces in this member. The Benevola is frequently quarried. This member weathers white to a very light grey.</td>
<td>Good karst former</td>
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<td></td>
<td>Fort Duncan Member - This is a medium to dark grey, thick bedded mottled dolostone with vuggy voids filled with white, sparry dolomite. The sparry dolomite is typically 1-4 cm wide, while the voids are typically between 1-4 meters in diameter. The light/new contrast is very distinctive for this unit. The cone-shaped fossil Saletterella is evident throughout this member as are burrow traces. Contact with the underlying Bolivar Heights Member is sharp and indicates an erosional event. The weathered surface of this unit is frequently covered with a clotted algae. Sinkholes (possibly plaeo-sinkholes) within this member often contain collapse breccia and tufa deposits.</td>
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<td>Bolivar Heights Member - Is composed of 3 distinct strata. The basal layer is a tan, vuggy dolomite directly overlaying the Antietam Formation. This formation is between 9-12 m thick. It is rarely exposed and may not be present everywhere. Above this is a 12-15 m thick bed of very light grey to tan, laminated, sheared, mylonitic marble known locally as the Keedysville marble bed. According to Brezienski this bed is the result of a fault zone mapped from south central Pennsylvania to Berryville, Virginia. The top strata are about 60 m of thin to medium bedded, dark grey, ribbony, lime mudstone. These beds weather to a light grey and are burrow mottled. The burrows vary in abundance from almost none to abundant and resembling anastomosing tubes. In places, the Bolivar Heights Member is a high calcium limestone suitable for burning in kilns. Manganese nodules have also been found in this formation.</td>
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<td>Chilowee Group</td>
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<tr>
<td>Antietam Formation</td>
<td>The lowest beds of this formation are thin (2-6 cm) very light grey quartzites with considerable burrows of Skolithos linearis. Green-grey sandy metasiltstones are interbedded with the quartzites. These strata grade up into medium bedded, white to light grey, bioturbated, well sorted and crossbedded fine to medium grained sandstones. The top beds are light to medium grey, calcareous, crossbedded, coarse grained sandstones or granular conglomerates. The upper beds form a ridge typically lower than the underlying Weverton Formation but higher than the overlying Waynesboro Formation.</td>
<td>NO</td>
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<tr>
<td>Formation &amp; Thickness</td>
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<td>Chilowee Group</td>
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<tr>
<td><strong>Harpers Formation</strong></td>
<td>This formation consists of pebble conglomerates at the base, phyllitic metasiltstones in the middle and clean metasedanstones at the top. The basal beds are interbedded meta-arkoses and pebble conglomerates, similar to the top of the underlying Owens Creek Member, with which it is transitional. The metasiltstones are dark green-grey, highly cleaved and foliated and quartz laminated. Locally, the metasiltstones can be calcareous and/or sandy. These metasiltstones may also be phylonic, as on Catocin Mountain or at the Short Hill fault near Weverton Road. The top of the formation contains thin (3cm-5m) dark green-grey, fine to medium grained, highly argilacious or arkosic, ferruginous and magnetite rich metasedanstones. <em>Skolithos linearis</em> burrows are evident at the top of the formation. The top of this formation is transitional into the overlying Antietam. The upper boundary of the Harpers Formation demarcates the upper boundary of the Blue Ridge Province.</td>
<td><strong>NO</strong></td>
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<tr>
<td><strong>Owens Creek Member</strong></td>
<td>Consists of quartzites and quartz conglomerates interbedded with metasiltstones and greywackes. The metallic blue to green-grey quartzites range from a very coarse grain to conglomerate. The lower 8 m of this formation is a gun metal blue pebble conglomerate. This grades upward into 20 m of green-grey pebble conglomerate. This bed is the most poorly sorted of the formation. Crossbeds are present within the quartzite but are often difficult to detect. Large pink quartz crystals (1-3cm) are locally common within these beds and are characteristic of them. Also present are blue and red quartz pebbles, magnetite, opaque minerals and blue green phyllite clasts, which give the formation its dark blue tint. At the base of this member is a diagnostic 4 m bed of clean green-grey conglomeritic quartzite. This formation, when deformed (as at Chimney Rock) has been misidentified as the Harpers Formation.</td>
<td><strong>NO</strong></td>
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<td><strong>Maryland Heights Member</strong></td>
<td>Thin (&lt;10 m) layers of medium to dark grey, poorly sorted and crossbedded quartzite interbedded with metasiltstone and phylitic shale. This formation is usually poorly exposed, with the metasiltstones and shales almost never exposed. The metasiltstone is usually dark green to grey, highly cleaved and sometimes confused with the with the metasiltstones in the lower part of the younger Harpers Formation. The quartzites are well exposed at Weverton Cliffs. The type location for this member is along the railroad at Blue Ridge and Elk Ridge. This member represents a transitional change between the lower Buzzard Knob and the overlying Owens Creek. It is often mapped based on a swale between the topographic ledges of the Quartzite benches of those two members</td>
<td><strong>NO</strong></td>
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<tr>
<td><strong>Weverton Formation</strong></td>
<td><strong>140-225 m</strong></td>
<td></td>
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<tr>
<td>Formation &amp; Thickness</td>
<td>Description</td>
<td>Cave or Karst former?</td>
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<tr>
<td>Chilowee Group</td>
<td><strong>Weverton Formation</strong> 140-225 m <strong>Buzzard Knob Member</strong> - The Buzzard Knob Member is the oldest of the Weverton Formation. It is mostly metasiltstones and metagreywackes mixed with 2 ledge-forming quartzites. The base of the member may be a transitional conglomerate into the underlying Loudoun Formation. Where this transition is absent, the Buzzard Knob sits unconformably on the Loudoun. The lower quartzite bed is a well sorted, light to medium grey, medium bedded and contains dark grey argillaceous layers up to 4 cm thick. Crossbedding is frequently evident throughout this bed and is distinguished by purple or yellow-gold foreset bands. The upper quartzite bed is a well sorted, medium to thick bedded, and very light green to grey. Crossbedding is infrequent in this bed. This bed often appears as a ridge-former. The type locality of this member is at Weverton Cliffs. It is also evident at Purcell Knob.</td>
<td>NO</td>
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