Field Trip to Karst Landscapes of the Great Valley in Maryland, Virginia, and West Virginia

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Springhouse at Fay Spring (Field trip stop 2). Photo by Randy Paylor
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Introduction

The Appalachian Great Valley is among the great karst regions of eastern North America, stretching from eastern New York to northern Alabama. Karst development in folded and faulted carbonate rocks of the Appalachian Great Valley is less well described than that within the more cave-rich Appalachian Interior Low Plateaus or the mountains of the western Valley and Ridge. Within the western Appalachian plateaus and the high-relief areas of the Appalachian Mountains, well-known fluvial karst processes dominate the karst geomorphology; reviews of these areas and associated references can be found in Palmer and Palmer (2009). Within the low-relief Great Valley, however, the influence of fluvial karst processes on karst development is more muted. Sinking streams and flashy springs emerging from cave entrances are much less common. Caves tend to be short, tight, and muddy and often maze-like in passage pattern. Nonetheless, solutional conduits in bedrock are pervasive, and are often encountered in outcrops, quarries, and wells. The karst of the Great Valley is best characterized as a phreatic karst system developed within an artesian basin that has been overprinted by late fluvial evolution. The more recent fluvial evolution is generally manifested in the upper few tens of meters of exposed carbonate bedrock, especially in proximity to incised surface drainages and where ancient alluvial sediments have mantled the bedrock (Doctor and others, 2014). Within some caves, the fluvial phase of cavern evolution is expressed as late stream invasion that has imparted an overprint of paragenesis and vadose canyon formation on earlier phreatic development.

Although still somewhat controversial, the idea that the karst and caves of the Great Valley in Virginia and West Virginia originated as part of a deep, regional phreatic groundwater system has been around for more than half a century, beginning with the classic paper by William Morris Davis (1930), and furthered by the work of J Harlen Bretz (1942). Later, as karst studies in the Appalachian Interior Low Plateaus blossomed, these ideas began to fade from favor (White, 2007, 2009). In an influential paper, William Davies (1960) suggested that cave development in the folded and faulted rocks of the Appalachians occurred very near to the water table (within several meters), and further helped to lead many future workers away from the hypothesis that cave development could have occurred far beneath the water table. Davies (1960) based his conclusions primarily on the basis of coincident elevations of stream terraces to passage levels in nearby caves, and the observation that clastic fills in caves suggested alternately filled and excavated passages within a zone of fluctuating piezometric surface. However, when looking beyond the caves toward the greater hydrogeologic context in the Great Valley, one finds (1) a prevalence of solutionally developed conduits in deep boreholes as well as in outcrops, (2) larger springs that actively precipitate tufa and marl with temperatures often slightly elevated above ambient, and (3) a relative paucity of sinking streams. These facts indicate a karst system that likely developed far below the water table (several hundred meters) early in its geomorphic history. Subsequent lowering of the land surface through erosion has effectively exposed the fossil circulation system to surface weathering and capture of infiltration and focused runoff. An extensive unconfined flow system has also developed as a result of near-surface karstification, and is continuing to develop in the modern day. Where carbonate units occur near ridges of siliciclastic or crystalline rocks, ancient alluvial fans that once covered the carbonates have contributed to enhanced karstification (Doctor and others, 2014). In these areas, thickness of unconsolidated sediments overlying carbonate units can be tens of meters or greater. Areas stripped of regolith for development or quarrying reveal a well-developed epikarst, with bedrock pinnacles several meters tall. Although drainage through the regolith is relatively rapid, potential for water storage within conduits in the upper bedrock zone may be great, and may facilitate
storage of contaminants spilled at the surface for extended periods of time due to adsorption onto sediments and ponding within bedrock conduits both above and below the water table.

Subsurface movement of sediment in response to fluctuations in water level results in closed topographic depressions and surface cover-collapse sinkholes, often induced by human activities. In order to understand the processes that cause sinkhole occurrence in any karst region, a working knowledge of the regional geomorphologic history is needed to provide a basic conceptual context. In the case of the Great Valley, the geologic evidence suggests that the karst initially formed by rising groundwater flow under confined (hypogenic) conditions. In such settings, geologic structural information obtained through mapping and combined with geochemical and hydrologic data is crucial for understanding groundwater flow, and for predicting future sinkhole susceptibility.

This guide will provide localities where the phreatic development of the karst system can be examined in areas unaffected by more recent fluviokarstic modification, as well as areas near trunk base-level streams where fluviokarst processes have overprinted the initial phreatic karst development to a great extent (Fig. 1).

![Regional map showing area of the field trip and localities that will be visited. A–A’ location of the cross section shown in Figure 2.](image)
Many of the more accessible and extensive caves occur near modern base-level drainages, where vadose processes have both exposed the caves by removal of fill material and altered them by the action of vadose cave streams and deposition of subaerial speleothems. The localities covered in this guide include surface features such as sinkholes, springs, solution conduits, and minor caves. Much of the material in this field trip guide was originally generated as part of an earlier, more extensive field trip guidebook for the Geological Society of America (Doctor and others, 2015).

**Geological overview**

This field trip takes place within the Great Valley subprovince of the Valley and Ridge geologic province in Maryland, Virginia and West Virginia (Fig. 1). Strata exposed in this subprovince comprise Cambrian through Devonian sedimentary rocks that were deposited on the east side of the Appalachian basin and later folded and faulted during the Alleghanian orogeny. At the base of this sedimentary package are over 10,000 feet (>3,048 m) of Cambrian-Ordovician carbonate strata (Fig. 2).

**Great Valley Lithology**

The karst of the northern Great Valley occurs in the carbonate rocks of Cambrian and Ordovician age. These formations are nearly all composed of mixed beds of limestone and dolomite, and vary greatly in siliceous mineral content. The carbonate units exposed in the study area are, from oldest to youngest, the Tomstown Formation, Waynesboro Formation, Elbrook Formation, Conococheague Formation, Beekmantown Group (limestone and dolomite rocks), New Market Limestone, Lincolnshire Limestone, and Edinburg Formation (Fig. 2). Overlying the Edinburg is the Martinsburg Formation, a siliciclastic unit comprising shale and interbedded sandstone and siltstone (Fig. 2; see Brezinski and others, 2012, for a summary of the stratigraphy and depositional history of the Cambrian and Ordovician units of this region).

The Cambrian system is represented by the Tomstown, Waynesboro, Elbrook, and Conococheague Formations. The Tomstown Formation is primarily composed of dolomite, with intervals of dolomitic limestone. The Waynesboro Formation is a heterolithic unit comprised primarily of shaly tan dolomite, maroon shale, sandstone, and some limestone. Brezinski (1992) broke out several members within the Tomstown and Waynesboro Formations; these members are most readily recognized in Maryland, the Eastern Panhandle of West Virginia, southeastern Pennsylvania, and northern Virginia. The Elbrook Formation is composed of thin interbedded silty dolomite and limestone. Surface karst features are not as prevalent in the Elbrook as in other units; however, numerous springs as well as some significant caves do occur in the Elbrook, especially within the middle limestone member (Brezinski, 1996) of the formation. The Conococheague Formation consists of interbedded limestone, dolomite, and sandstone. This unit represents carbonate deposition within an intertidal to subtidal carbonate platform environment during a marine regression in the Late Cambrian into the Early Ordovician. The limestone is composed of cyclic bundles of subtidal to intertidal carbonate that contain dolomitic cycle caps, and laminae distributed through the limestone that range in thickness from a few millimeters up to 2 cm. Much of the limestone is layered with dolomite in wavy bands, which take on a ribbon-like appearance. Prominent coarse quartz-sandstone beds occur intermittently within the formation and hold up topographic ridges, while the purer limestones weather to valleys and often host karst features, including caves.

The Ordovician system is represented by the uppermost Conococheague Formation, the Beekmantown Group, a package of Middle Ordovician limestones, and the Martinsburg Formation.
Figure 2. Generalized stratigraphic column for the northern Great Valley in Virginia and West Virginia.
The Beekmantown Group is composed of limestone and dolomite, generally of higher purity (less siliceous mineral content) than the Elbrook Formation and Conococheague Formation. The units in the Beekmantown Group include the Stonehenge Limestone, the Rockdale Run Formation, and the Pinesburg Station Dolomite. Karst development is abundant, and many caves in the Great Valley occur within the Beekmantown.

The New Market Limestone, Lincolnshire Limestone, and Edinburg Formation are relatively thin units, and thus are referred to collectively as the Middle Ordovician limestones. These limestones all host significant karst development, especially farther south in Virginia. The New Market is an especially pure limestone, and solution features are commonly found in outcrops. The Lincolnshire notably contains interbedded black nodular chert. The Edinburg is mixed shale and limestone, with nodular limestone common in outcrops.

McCoy and Kozar (2008) determined that sinkhole density within the Great Valley of West Virginia is highest in the Beekmantown Group (3.5–4.0 sinkholes per square mile), followed by the Conococheague Formation (2.5–3.0 per square mile), and roughly equivalent between the Middle Ordovician limestones and the Elbrook Formation (1.3–1.5 per square mile). Hubbard (1983) found karst features to be most prevalent within the Middle Ordovician limestones farther south in Virginia. With the recent acquisition (between 2010 and 2014) of lidar data in the Great Valley of Virginia, West Virginia, and Maryland, high resolution topographic digital elevation models (DEM) are now enabling more accurate inventories of surface depressions in this karst region (Doctor and Young, 2013).

Stratigraphic units of the Cambrian–Ordovician carbonate platform are well exposed along the bluffs of the Potomac River and along the cuts made for the Chesapeake and Ohio (C&O) Canal (Southworth and others, 2008). Those units as well as solutional conduits and caves can be seen along the river bluffs and will be visited at Stop 5.

Great Valley Hydrology

Much of this field guide will focus on the karst of the Opequon Creek and Shenandoah River basins which drain northward into the Potomac River. Since 2002, the U.S. Geological Survey has been investigating the karst of the northern Shenandoah Valley to better characterize the carbonate aquifer system of Frederick and Clarke Counties, Virginia (Harlow and others, 2005; Kozar and others, 2007a, 2007b; Kozar and Weary, 2009; McCoy and others, 2005a, 2005b; Nelms and Moberg, 2010) (Fig. 3), and to provide relevant hydrogeologic information that can be used to guide the development and management of this important water resource. These investigations form the foundation of a regional study of the karst system that combines hydrologic and geologic information to improve the understanding of the aquifer system, its relationship to surface features, and potential subsidence hazards over a multi-county area of Virginia and West Virginia (McCoy and Kozar, 2008; Doctor and Doctor, 2012). Detailed geologic mapping along with fracture analyses, conduit analyses, and mapping of karst features are all components of this framework (Orndorff and Harlow, 2002).

Two groundwater modeling studies of the Opequon Creek basin have been published (Kozar and Weary, 2009; Yager and others, 2013), in which the groundwater basin was represented in the model domain as an equivalent porous medium (EPM) under steady-state flow simulation using MODFLOW (Harbaugh and others, 2000). The multi-layer models included anisotropy caused by attitudes of rock bedding, and incorporated the influence of fault zones and springs. Unless interrupted by cross-strike faults, fractures, or caves, anisotropic groundwater flow was assumed to be greater parallel to bedding planes and thus favors the dominant strike-parallel, north-northeastern direction.
However, the MODFLOW models were unable to incorporate changing dip along bedding planes (bedding was assumed to be vertical in the models); thus, fold structures are not represented in the models (Yager and others, 2013). Although both models attempted to describe some of the geologic complexity and its effect on flow paths to the major springs, it is evident from the detailed geologic maps and known karst development that the reality is more complicated than the models could simulate. As stated in Yager et al. (2013, p. 1215), “the individual flow paths simulated by the model do not necessarily predict actual local-scale movement of contaminants through this karst aquifer to individual springs.” Nevertheless, the MODFLOW modeling was generally successful in simulating the overall water balance for the Opequon basin, and perhaps more significantly, was able to reproduce the mean groundwater ages of spring waters estimated by using environmental tracers such as $^3$H/$^3$He and chlorofluorocarbons (CFCs) in a two-component binary-dilution model (Yager et al., 2013).

**Stop 1. Karst Window at Confluence of Opequon Creek and Potomac River**

(39°30’ 48.29”N, 77°52’11.61”W)

*NOTE: Landowner permission is needed prior to accessing this site. Park in the Izaak Walton League parking lot and walk back up the hill to the gate on the west side of the road. Follow path to the south-southwest, turning towards the west and then back north as it descends into the large sinkhole containing the karst window.*

The karst window is located in the base of a large sinkhole in the Lower Ordovician Rockdale Run Formation (Fig.4). The window consists of a very short reach of a small cave stream that emerges from the southeastern side of the sink and flows back underground to the north-northwest (Fig.5). The water then re-emerges at an underwater spring in the channel of Opequon (oh-PEK-un) Creek ~800 ft (244 m) to the north-northeast. Dye tracing shows that at least part of the water flowing through the window originates at Swan Pond Spring located ~3 mi (4.8 km) to the south-southwest of the window (Jones, 2005). This water flows north from Swan Pond in an unnamed surface stream for ~1 mi (1.6 km) and then sinks underground. All of the flow from that sink point to the karst window is subterranean, with flow taking ~4 days to cover the approximate 2 mi (3.2 km) straight-line distance remaining to the karst window (Jones, 2005).
Figure 4. Geologic map of the area around Stop 1 showing location of the karst window and qualitative dye traces to Opequon Creek (dashed arrow) (Jones, 2005). Geologic map from Dean and Lessing (1987). Geologic map units: Om—Martinsburg Formation; Oc—Chambersburg Limestone; Onm—New Market Limestone; Obps—Pinesburg Station Dolomite; Obrr—Rockdale Run Formation; Obs—Stonehenge Limestone; Obs—Stonehenge Limestone, Stoufferstown Member. Contour interval is 20 ft. Regional map showing area of the field trip and localities that will be visited. A–A’ location of the cross section shown in Figure 2.

Figure 5. Participants in the 2010 "Growing Communities on Karst and the Great Valley Water Resources Science Forum" field trip at the karst window. This conference was organized by the Potomac Headwaters Resource Conservation and Development Council, and also sponsored by the USGS, the West Virginia Department of Environmental Protection, and the West Virginia Department of Health and Human Resources (photo by D. Weary).
Flow from the karst window northward to Opequon Creek has been dye-traced more than once, including a spectacular demonstration using sodium fluorescein by W.K. Jones of the Karst Waters Institute in 2010, which colored the entire downstream Opequon Creek brilliant green (Fig. 6). During relatively low flow conditions, the dye took ~4 h to cover the approximate 800 ft (305 m) straight-line distance from the injection point in the karst window to the spring rising in the bed of Opequon Creek. The window is at ~425 ft (130 m) elevation, and the spring emerges at ~335 ft (102 m) elevation within the channel of Opequon Creek after completing a phreatic loop to an unknown depth below present base level. The straight-line path of the dye trace parallels the strike of a thrust fault located just to the west of the karst window. Fast groundwater flow and solution conduits may be enhanced due to fracturing of the rocks adjacent to the fault plane.

As you walk along the trail on the way back to the parking area, note the presence of occasional rounded quartz sandstone cobbles on the ground surface. These are evidence of a higher terrace level for Opequon Creek and the Potomac River, ~0.4 mi (0.6 km) to the northeast of the window. Sinkholes are common in the high bedrock terraces adjacent to the Potomac River in this area. This is because the local hydrologic gradient is steep, as the tributaries on the flanks of the Potomac River cannot keep up with the rate of its down-cutting. The steep gradient results in localized higher velocity groundwater flow, and both enhanced solution of the carbonate bedrock and physical removal of soil and sediment from the sinkholes.
Stop 2. Fay Spring

(39°12′18.28″N, 78° 7′50.37″W)

NOTE: Parking is available at the small paved lot adjacent to Redbud Rd. (State Rte. 661), and then it is a 600 m walk down the driveway to the spring house. Please do not park at the spring house without prior landowner permission. The discussion below is abbreviated from Doctor et al. (2011).

Fay Spring is located within the city of Winchester, Virginia, and was once used as a part of the municipal water supply (Harlow and others, 2005). The spring is situated on a northwest-trending strike-slip fault cutting Lower Ordovician dolomite and limestone (Orndorff and others, 2004) (Fig. 7). The spring was instrumented by the Virginia Water Science Center of the U.S. Geological Survey for continuous monitoring of discharge, water temperature, and specific conductance from late April 2007 to October 2010. For the period of record, the mean discharge was 2592 m$^3$/d (1.09 cfs) and the mean temperature was 13.1 °C. The continuous record for this perennial spring shows a broad seasonal rise and fall in water temperature and discharge, punctuated by rapid increases in flow with concomitant decreases in conductivity (Fig. 8). However, water temperature does not respond to discharge in as flashy a manner as does conductivity, and relatively little temperature change is observed even during periods of large changes in flow (Fig. 9).

Figure 7. Geologic map of study area showing traced connections to Fay Spring and Sempeles Spring. Geology based partly on Orndorff and others, 2004)

Conversely, conductivity does not show a pronounced seasonal trend as does water temperature, and tends not to vary greatly when above the long-term mean value of 711 μS/cm; however, a winter precipitation event in January 2008 caused an unusual dramatic increase in conductivity, with no corresponding change in discharge or water temperature at the spring (Fig. 8). This increase in
conductivity was hypothesized to result from sinking surface water carrying road salt to the spring, and may have been a delayed response to events ranging from several days up to two weeks earlier. In order to test this hypothesis, quantification of the influence of local, surface-water input to the spring was sought through a dye trace. On 30 June 2009, ~2 kg of 20% rhodamine WT dye solution was injected at the terminal sink point of Sunnyside Run, an ephemeral stream ~1 km west of Fay Spring (Fig. 7).

Figure 8. Continuous monitoring at Fay Spring, from 2007 to 2010. An unusual conductivity increase in January 2008 is highlighted in the vertical box and expanded in the right-hand panels. A similar, but more gradual conductivity excursion was observed during a major period of recharge as a result of above normal snowfall in 2010. Precipitation data obtained from National Weather Service Cooperative Station 449181 at Winchester, Virginia.

Water samples for dye and major ion analysis were collected by an automatic sampler with a frequency of every six hours to once per day during the trace period. The dye breakthrough curves at Fay Spring and Semples Spring are shown in Fig. 9. Dye was recovered at two springs (Fay Spring and Semples Spring, Fig. 7) located 300 m apart along a single fault, and within the channel of Redbud Run immediately upstream of its confluence with the spring run of Semples Spring; dye was not
recovered at any of the other sampling sites. The results of this tracing test provided evidence of initial rapid dye transit (>500 m/d) and low overall mass recovery (~12%) that includes sampling at both springs over the course of five weeks. In spite of the low tracer recovery in the water samples, monitoring by passive activated charcoal samplers verified a positive recovery of the dye at each of the three sites.

Figure 9. Dye recovery at Sempoles Spring and Fay Spring. Note pulsed recovery of dye released from shallow groundwater storage during later storm events. RWT—rhodamine WT.
The amount of dye recovered at Sempeles Spring was approximately ten times greater than that at Fay Spring; however, Sempeles Spring was not instrumented for continuous monitoring. The amount of dye recovered at Redbud Run was less than at either spring based on the passive charcoal sampler results.

In conjunction with monitoring for dye that had been injected for a different trace several miles to the north, chemical sampling continued at Fay Spring throughout the winter and spring of 2009–2010. Water samples were collected by an automatic sampler in the spring house every six hours to daily from January–April 2010, and a complete record of daily samples was obtained during the snowmelt period. The results of the continuous sampling during this period revealed a gradual rise in conductivity following the recession of a rainfall event in late January. Then in early February 2010, the largest snowfall on record for the region occurred. Melting of this snow began on 7 February, and during the melt period the rise in discharge was closely correlated to a rise in conductivity (Fig. 10). Rapid increases in discharge occurred with rainfall both prior to and following the snowmelt period, and resulted in concomitant decreases in conductivity. However, during the large snowmelt event, only a gradual, damped discharge increase was observed that corresponded with a similar damped increase in conductivity. Chemographs revealed dilution in NO$_3^-$, Cl$^-$, SO$_4^{2-}$, and Mg$^{2+}$ with most large rain events; however, Cl$^-$ increased during peak snowmelt with little change in other ion concentrations (Fig. 11).

![Figure 10. Record at Fay Spring before and after snowmelt in winter and spring 2010. Note the muted, gradual discharge increase in response to snowmelt versus the greater, more rapid discharge increase in response to rainfall events. Temperature decrease in response to rainfall is evident, though damped, and almost nonexistent in response to snowmelt.](image-url)
In comparison to the results of chemical sampling during the summer, the background concentrations of SO$_{4}^{2-}$ and NO$_{3}^{-}$ are similar in the winter; however, Cl$^{-}$ concentration is higher in the winter (up to 80 mg/L). This added chloride component may be due to the impact of road salt in the winter months. Based upon geochemical sampling in both summer and winter seasons, Fay Spring seems to be dominated by a diffuse flow component with elevated background levels of NO$_{3}^{-}$ (10–12 mg/L as NO$_3$), SO$_{4}^{2-}$ (45–50 mg/L), and Cl$^{-}$ (35–40 mg/L) that is noticeably diluted only after large rain events.

Greater chemical variability was observed during larger discharge events in the winter and spring than in the summer, but chemical variability was relatively muted overall. For example, conductivity values at Fay Spring do not drop below 600 μS/cm, even during the largest events. Sinking surface runoff might be a reasonable source of the water that causes dilution of the spring chemistry during peak flows, but this component does not dominate the spring discharge. The dye trace resulted in a demonstrable conduit connection between the sinking surface stream of Sunnyside Run and the spring, yet the overall mass recovery was low, and a large proportion of that recovery occurred with subsequent storm events long after the initial dye breakthrough. During the snowmelt period of 2010, Fay Spring showed a higher conductance and temperature than the water sinking at Sunnyside Run throughout the melt period. Thus, it is possible that another, as yet unidentified, source of sinking surface runoff affects Fay Spring.

**Figure 11.** Major ion chemistry at Fay Spring during summer and winter events. Left panel shows chemographs during the dye-trace monitoring in July 2009; right panel shows chemographs during the snowmelt period of 2010. Note that the ranges of each chemical constituent are shown on equivalent vertical axes for both sampling periods. Except for chloride and silica, the chemistry varies around average background values that are remarkably consistent between the two seasons, despite the recording of a nearly full range of discharge at Fay Spring.
Stop 3. Karst Features of Hupp’s Hill and Crystal Caverns

(39° 024.043N, 78°20259.693W)

Turn right into the entrance driveway for Crystal Caverns at Hupp’s Hill, park in front of museum building.

The area around Hupp’s Hill has many sinkholes and developed tour cave (Crystal Caverns, currently closed to tours) that illustrate the relationship of stratigraphy and structure to the conduit system (Fig. 12). The area sits on a topographic high north of the confluence of Cedar Creek with the North Fork of the Shenandoah River, and the karst development is related to the high hydraulic gradient. Numerous sinkholes occur within the vicinity of an adjacent quarry, many with open throats and evidence of soil piping. These sinkholes are generally subsidence sinkholes with gradual movement of sediments into the underground system, as opposed to the type that form due to catastrophic collapse. Two of the sinkholes on Hupp’s Hill have entrances to small caves that are developed along vertical joints in the bedrock. The active nature of these sinkholes can be attributed to their topographic position in relation to Cedar Creek, and also to the proximity of a large abandoned quarry to the north that previously had lowered the local water table. The geology of this area is gently dipping Middle and Upper Ordovician New Market Limestone, Lincolnshire Limestone, and Edinburg Formation that occur...
near the nose of a southward plunging anticline (Orndorff and others, 1999) (Fig. 12). High calcium limestone (as high as 98% calcium carbonate; Edmundson, 1945) of the New Market Limestone was mined from the quarry to the north. The contact between the Lincolnshire Limestone and Edinburg Formation runs northwesterly across the Crystal Caverns property.

Common to all karst regions, sinkholes can be entrance points for contamination into the groundwater system that may include agricultural runoff (pesticides, herbicides, and animal waste); industrial pollution; and seepage from underground storage tanks, landfills, and private septic systems, all of which can be found in the Shenandoah Valley. Historically, sinkholes have been used by landowners as dumping sites for waste. Slifer and Erchul (1989) estimated that there are nearly 1400 illegal dumps in sinkholes and 4600 in karst areas of the Virginia Valley and Ridge province. An example of this can be seen in the large sinkhole adjacent to the quarry located to the northwest of the parking area (Fig. 13).

Figure 13. David Weary examines a trash-filled sinkhole, perhaps once used as a domestic landfill adjacent to the quarry at Hupp’s Hill.

Crystal Caverns

Crystal Caverns is formed within the Lincolnshire Limestone and Edinburg Formation, and is located on the nose of a broad, south-plunging anticline (Fig. 12). The cave has very sparse speleothem formation, and passage morphologies can be closely inspected. The entrance to the cave is through a collapsed sinkhole leading into a network maze of passages developed mostly along joint planes and to a lesser extent along bedding planes (Fig. 14). The major northeast-trending passages parallel the local
northeast-trending joint set. The intersection of the joints with the bedding planes must be important to conduit development, because this lineation has a major southwest trend and shallow plunge, and a secondary southeast trend and plunge that are consistent with the cave passage orientations.

Descending into the cave, one moves down-dip in the Edinburg Formation, with a phreatic half-tube formed within the ceiling along the bedding. The dominant flow direction inferred from passage

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**Figure 14.** Map of Crystal Caverns and locations of photographs taken within the cave. (A) Joey Fagan examines a phreatic solution channel in the ceiling of the cave. Bob Denton admires cupolas (B), a large rising phreatic tube (C), and fine sediment that fills the passage to the ceiling (D) in the lower parts of Crystal Caverns. (E) Laminated clay and silt deposits on top of a flowstone shelf that spans a passage. (F) Dense, layered calcite that once lined the lower portions of Crystal Caverns and was partially excavated from the cave.
morphology is consistently one of rising from depth. At the lowermost end of the tour path, water rises after heavy rains and may flood the lower passage levels; however, there is neither a cave stream nor any known connection to a surface stream. The normal water level at the bottom of the cave is slightly elevated above the water level of the lake within the abandoned quarry adjacent to the cave. Divers have explored submerged conduits within the lake to a depth of several meters (B. AmEnde, 2014, personal commun.). Cupolas, ceiling channels, and blind passages are common and well developed in the cave, and these solutional bedrock forms are indicative of phreatic speleogenetic conditions (Fig. 14A–E). In the lowermost level of the cave, calcite wall coatings up to several centimeters thick helped to give the caverns its name (Fig. 14F). Today, only remnants of the coatings remain, as they were partially removed by solution, and partially excavated by humans.

**Epikarst at Hupp’s Hill**

The surface of a karst terrain is termed the epikarst, or the “skin” of the karst. It is within this zone that intensive solution of bedrock tends to occur, resulting from the production of weak acids from soil carbon dioxide in solution (carbonic acid) and other organic acids in the soil and regolith atop the bedrock surface. Excellent epikarst development can be seen in the exposed pinnacles of rock at the edge of the quarry. Many of these pinnacles are exposed due to the loss of soil and regolith that filled former closed depressions. The depressions were intersected by the quarrying activity, and cut open. In at least one place along the quarry edge, the ancient sediment fill can be examined. This fill shows evidence of root casts and other burrows preserved within an oxidized, iron-oxide rich residuum.

Distinctive weathering rinds are present in numerous bedrock outcrops exposed through blasting above and immediately adjacent to the cave, both within the quarry and in a recent development project located within 500 m of the cave that was halted. Two types of rinds occur: some are coatings of calcite-cemented sediment that once filled voids in the bedrock (Fig 15A), and others are simply corroded bedrock (Fig. 15B). The rinds of calcite-cemented sediments are present in both the Edinburg Formation and overlying New Market Limestone, react vigorously to acid, and have a distinctive yellow-orange

![Figure 15. Bedrock corrosion rinds and calcite-cemented sediments coat solutional fracture partings in the Edinburg Formation on Hupp’s Hill. (A) Dark-gray Edinburg limestone with light-tan, calcite-cemented corrosion rinds coating the surface along solutional fracture and bedding plane partings. (B) A hand sample of the rock shows a distinct zone of incomplete solutional weathering extending into the limestone to a thickness of 5–10 mm. (C) The thickness of calcite-cemented sediment fill within solutional openings and attached to the bedrock surfaces may be up to several centimeters.](image)
color when freshly broken (Fig. 15C). The rinds form primarily along vertical solutionally enlarged fracture partings.

Numerous anastomotic solution channels are present in the walls of the quarry, some clearly having been formed by horizontal flow, and others having been formed by vertical flow (Fig. 16). Certain geometries of the voids in the bedrock could only have arisen from solution occurring upward into the bedrock from below, such as open and sediment-filled pockets with no discernible connection to the overlying surface (Fig. 16A). Some solutional features appear to have once been solutional “feeder tubes” for rising flow under artesian head in a phreatic environment, and are now exposed at the surface (Fig. 16B).

Figure 16. Epikarst features seen in the quarry exposures at Hupp’s Hill; see text for descriptions (photograph in panel A by C. Khourey; all others by D. Doctor).
The intersection between surficial weathering and subsurface or subsoil weathering is evident in places where rillenkarren are juxtaposed against smooth-walled overhanging pockets in the limestone (Fig. 16C). Early inception horizons along bedding plane partings that have been enhanced by solution, later filled with sediment, and then partially excavated once more are exposed in the quarry walls (Fig. 16D), with apparent “feeder tubes” providing rising flow into the bedding plane horizon (Fig. 16E), which are partially coated with calcite-cemented sediment. These early phreatic pathways have all been overprinted by later solution and excavation from waters infiltrating at the surface and descending through and along these former phreatic pathways, thus resulting in a very complex epikarst (Fig. 16F). This epikarst zone serves as its own aquifer, with abundant detention storage within dead-end voids, as well as storage capacity in voids partially filled with sediment.

These features contain remnants of the corrosion and cemented sediment rinds similar to those seen in bedrock outcrops in the zone of recent blasting. The anastomotic channels contain sediment fills of both clay-size and coarse, angular fragments. The angular fragments are interpreted to have resulted from an earlier phase of collapse, which provided the sediment that fills the bedding plane conduits. Such paleocollapse features are also present as sinkhole fills in the walls around the quarry.

**Hupp Spring**

Hupp Spring is located at the base of Hupp’s Hill about .5 mi (.8 km) south of the quarry and below the elevation of the cave entrance. This spring has very high hardness, has been observed to effervesce, and precipitates tufa within riffles in the spring run. The spring continues to flow even during drought periods, and the flow has been estimated at 1135 L/min (300 gal/min). It appears that the spring drains the conduit network that may have flowed out through Crystal Caverns at one time, but to our knowledge, this has not been confirmed through dye tracing.

**Stop 4. Groundwater hydrology and geohydrologic framework at the USGS Leetown Science Center**

(39° 21′6.59″N, 77°55′52.10″W)

**Introduction**

This field trip stop provides a venue for examining karst features and discussing the karst hydrology and geohydrology of rocks at the Leetown Science Center (Fig. 17) in the eastern limb of the Massanutten synclinorium in West Virginia. Most of the content of these discussions is documented in a series of reports produced during extensive interdisciplinary studies conducted by the U.S. Geological Survey (USGS) and partners in the early and mid-2000s (Kozar and others, 2007a, 2007b; Kozar and Weary, 2009; McCoy and others, 2005b). The material for this part of the field trip guide is largely excerpted from Kozar and others (2007b). These studies, and others in various locations in the Shenandoah Valley, were prompted by concerns over groundwater supplies after the region was effected by a severe drought from November 1998 through February 2000.
The Leetown Science Center is a research facility operated by the USGS that occupies approximately 455-acres in the eastern panhandle of West Virginia near Kearneysville, West Virginia (Figs. 18 and 19). Landsurface elevations in the study area vary from 410 to 630 ft above NAVD88. Hopewell Run and Opequon Creek (Fig. 18), the major streams in the area, drain to the Potomac River. The primary mission of the Center is to conduct research needed to restore, maintain, enhance, and protect aquatic species and terrestrial organisms and their supporting ecosystems. The Center is also the site of the National Fish Health Laboratory, where research is conducted on fish diseases, declining species, invasive species, genetics, aquaculture, and ecology. The Center is co-located with the U.S. Department of Agriculture (USDA) National Center for Cool and Cold Water Aquaculture (NCCWA). Large quantities of high-quality, cold groundwater are needed to support operations at these two research facilities. In 2006, water for research operations at the facilities was obtained from three large springs on Center property. A production well provides additional water during periods of high demand and during periods of low spring yield in late summer and early fall. A second production well provides primarily water for restrooms, water fountains, and a dormitory. Continued operation of the facilities as well as possible expansion of research activities depend on locating additional sources of high-quality, cold water to augment existing supplies.
Figure 18. Location of study area and boundary conditions for the groundwater flow model of the Leetown area, West Virginia. Figure modified from Kozar and others (2007b).
Previous Investigations

The relation between geology and groundwater supply and quality in the Great Valley of West Virginia was first discussed by Jeffords (1945a, 1945b). Further work by Graeff (1953) and Beiber (1961) defined the lithologic control that carbonate units have on the groundwater quality, quantity, and direction of flow in the aquifers of the Great Valley. Large springs discharging in excess of 1,000 gal/min (76 L/s) from these carbonate units were correlated with the location of numerous faults in the area by Hobba and others (1972). Taylor (1974) concluded that systematic fracturing of the carbonate bedrock, attributed to a four-phase deformation history, is partially evident from topographic analysis of the area. He found well yields and spring locations in lowland areas to be related to structural features capable of transporting large quantities of water. Joints, faults, and fractures in lowland settings serve to move groundwater downgradient, although seasonal and annual fluctuations in storage and base flow to streams associated with these features can be high (Hobba, 1976; 1981). Base-flow stream discharge data, and water level data from Kozar and others (1991) and Shultz and others (1995), were used to calculate aquifer transmissivity values that range across an order of magnitude, which was attributed to the anisotropic nature of fracturing in the carbonate rocks. Preferential flow in the direction of strike was verified by the dye-racing work of Jones (1991), Kozar and others (1991), and Shultz and others (1995). Previous dye-tracing work by Jones and Deike (1981) at the Center showed the aquifer to dominantly consist of steeply dipping bedding planes and a diffuse network of fractures that may retard travel times and force circulation to depths below that common to other karst systems. In nearby Frederick County, Virginia, Harlow and others (2005) applied the conceptual model of Wolfe and others (1997) to describe the influence of structural features on the spatial arrangement of various rock units and active karst development at moderate depths.

Groundwater flow patterns in the Great Valley are complex. The once flat-lying sedimentary rocks have been folded, faulted, and intensely weathered such that a variably thick layer of regolith overlies steeply dipping deformed bedrock units. Recharge in the form of infiltrating precipitation initially moves into the regolith where much of it is stored. Movement of water to the underlying bedrock occurs via leakage to open fractures, faults, and bedding planes or direct runoff into surficial karst features. Flow within the bedrock is controlled by the orientation and connectivity of the fracture system and location of solutionally enlarged conduits. The geologic structure in the regolith and continuous bedding planes serve to apparently force flow parallel to regional gradients (Jones, 1991). Frazier and others (1988) conducted a detailed surface-water assessment of the Center and adjacent property using watershed models primarily to assess the potential for flooding and long-term stability of engineered ponds proposed for construction. A major finding of the study, which is also supported by the work of Jones and Deike (1981), was that sufficient water for operations at the Center is not available during drought years; operations at the Center would have to be adjusted to compensate for reduced water availability during droughts or other periods of prolonged below average groundwater levels. Additional borehole log and surface geophysical data were collected by Mayle and Schnabel (1998) as part of geotechnical investigations prior to construction of the NCCWA facility.

Historical Notes

The unincorporated hamlet of Leetown is named for British-born Charles Lee who built an estate here called Prato Rio in 1773; the main house rests today just southwest of the Science Center. Lee served in the Revolutionary War as a general in the Continental Army. Charles Lee is not closely related to the more famous American Civil War general, Robert E. Lee. Another notable late colonial manor in the Leetown area is Harewood, located about 3.5 mi to the south-southwest of the Leetown Science
Center (Fig. 17). This house was built in 1770 for Samuel Washington, the brother of George Washington, the 1st president of the United States. Dolly and James Madison, a framer of the U.S. Constitution and 4th president, were married there in 1794. Like many of the older mansions in the Great Valley, Harewood is located near a large karst spring, known as Washington Spring.

**Hydrogeologic Setting**

The Center is uniquely situated with respect to the availability of abundant, high-quality, groundwater resources. The area is dominated by folded and faulted karstic carbonate rocks of Cambrian and Ordovician age that strike predominantly to the northeast (Fig. 19). Low-permeability lithologic units such as the Martinsburg Formation and the Conococheague Limestone control groundwater flow by acting as barriers to water flowing down the hydraulic gradient and across the strike of the bedding (Figs. 19 and 20). This retardation of cross-strike flow is especially pronounced in the Leetown area, where the bedding typically dips at steep angles. Geologic structures that disrupt the rocks in cross-strike directions, especially highly permeable fault and fracture zones, provide avenues through which groundwater can flow laterally across or through strata with low primary permeability.

![Figure 19.](image-url) **Figure 19.** A) Geology, B) geologic cross section, of the Hopewell Run surface water drainage basin, and the USGS Leetown Science Center, West Virginia. [Geology modified from Dean and others, 1990]. Figure taken from Kozar and others (2007b).
Geology

A geologic map of the Leetown area was published at a scale of 1:24,000 by the West Virginia Geologic and Economic Survey (Dean and others, 1990). This map was used as a basis for more detailed mapping of the area near the Center (Fig. 19). This detailed mapping includes locations of bedrock contacts and the orientation of cleavage, fractures, and other structural features. Updates to the geologic interpretation include slight adjustments in the location of some of the bedrock contacts within the Hopewell Run drainage basin and the recognition of two west-northwest trending cross faults in the Leetown area (Fig. 19).

The bedrock in the Leetown area consists primarily of fractured limestone and dolomite of the Lower Ordovician and Upper Cambrian Conococheague Limestone; the Lower Ordovician Stonehenge Limestone, Rockdale Run Formation, and Pinesburg Station Dolomite; the Middle Ordovician New Market and Chambersburg Limestones; and the Upper and Middle Ordovician Martinsburg Formation (Fig. 20). The geologic structure of the area is complex, with many thrust faults oriented parallel or sub parallel to the regional bedrock strike (approximately N. 20º E.). Cross-strike longitudinal and oblique faults also occur at orientations of approximately N. 80º W. and N. 65º E.,
respectively (Fig. 19). The rocks are tectonically deformed and many upright and overturned folds occur throughout the area. The overturned folds verge to the northwest, with limbs dipping steeply to the southeast. The variable permeability of the different lithologies, especially the more resistant Martinsburg Formation and Conococheague Limestone, together with geologic structure and topography, govern the pattern of groundwater movement through the study area.

Joints, bedding, and cleavage within the bedrock also can exert control on groundwater flow patterns and rates. A compass rose diagram (Fig. 21) clearly illustrates the dominant directions of joint planes within the surficial bedrock in the Leetown area. Two dominant joint directions are apparent: one oriented approximately normal to and the other oriented approximately parallel to the regional strike of bedrock. The joint set normal to the bedrock strike (cross-strike joints) trends in a range from about N. 90° W. to N. 60° W., and the joint set parallel to bedding (strike-parallel joints) trends in a range from about due north to N. 40° E. The cross-strike joints may be over-represented in this sampling as a result of the linear geometry of the outcrops, which typically causes them to be better exposed than the strike-parallel joint set.

![Figure 21. Compass-rose diagram showing the dominant joint trends in the Paleozoic rocks of the Leetown area, West Virginia.](image)

Bedding attitudes in the Center area are complex as a result of the tectonic events that occurred in the geologic past. Bedding varies from nearly flat-lying in some areas to nearly vertical in others. In some areas, the bedrock dips steeply toward the southeast, and in other areas, beds have been overturned as a result of complex folding within the bedrock. Cleavage is produced by alignment of platey minerals in rocks under tectonic stress and commonly occurs in planes normal to the direction of maximum compression. Some fanning of cleavage direction with respect to the fold axes has been observed in the study area. In the Center area, cleavage is common in the Martinsburg Formation and in the Stonehenge
Limestone, which has a silty component, especially within the massive, basal Stoufferstown Member where it is so well developed that it completely obscures the bedding locally. The predominant attitude of cleavage planes in the Center area is a strike of about N. 20° E. and near-vertical dips to either the east or west, with eastward dips prevalent. Although dissolution of limestone in karst areas is common, the Leetown area does not exhibit classic mature karst topography. Sinkholes are found in the study area, but they are sparse and confined primarily to the Conococheague Limestone (Figs. 19 and 22).

![Image](image.png)

**Figure 22.** W. Palmisano (USGS retired) looks into the throat of a sinkhole developed in the Conococheague Limestone near the Leetown Science Center. Photo credit D. Weary.

**Hydrology**

Groundwater flow occurs predominately in diffuse fractures within the study area. Water that flows through the intricate network of lower-permeability diffuse fractures is collected over a broad area in higher-permeability conduit drains that primarily coincide with the major thrust, normal, and cross faults. Zones of high conductivity are also indicated by surface geophysical and LiDAR data in some areas. These geophysical anomalies are most likely highly transmissive fracture zones but may be unmapped faults. Investigators who developed conceptual models of groundwater flow in the area (Kozar and others, 1991 and Shultz and others, 1995) recognized the importance of solutionally-enlarged bedding planes in governing groundwater flow. The importance of cross-strike faults or other complex geologic structures, such as overturned or tightly folded structures, in controlling groundwater flow, was not fully realized until it was viewed in context with recent work (McCoy and others, 2005 a;b), and as a result of the detailed data collected for this investigation.
Dye Traces

This section of the field trip guide is abridged from Field (2007).

Groundwater flow rates and trajectories at the Leetown Science Center are an issue of concern. The primary concern is the low flows of Balch, Blue, and Gray Springs during drought, but additional concerns are the recharge areas to the springs. For example, to the east and upgradient of the Center are the Leetown Pesticide Site and the Jefferson County Landfill (sites A and B on Fig. 17 respectively), a U.S. Environmental Protection Agency Superfund site (CERCLIS I.D. WVD980693402) where pesticide compounds consisting of isomers of benzenehexachloride (BHC), DDT, DDE, DDD, endrin, endosulfan, and aldrin were detected. Potential transport of contaminants in groundwater from these sites is a concern.

A later study of low level halogenated volatile organic compounds (VOCs) found that nearly all the water sampled in the vicinity of the LSC has been affected by addition of halogenated VOCs from non-atmospheric source(s) (Plummer and others, 2013).

Tracer tests conducted in the Leetown area were limited to the immediate vicinity of the Center to provide additional hydraulic information about general flow directions, velocities, and mean residence time of groundwater. Additionally, the tracer tests provided information on how groundwater flow is affected by withdrawal of water from springs and wells for use at the Center.

Previous Tracer Work

Seven tracer tests were conducted at the Center in 1979 and 1980 using fluorometric tracers (Jones and Deike, 1981). These early tests were among the first efforts to characterize the flow of groundwater in the conduit-dominated portion of the flow system in the region. Travel times for the previous tests all indicated long groundwater transit times for a karst aquifer; 58 to 81 day travel times for dye recovery at resurgence points were documented over distances ranging from 0.9 to 2.21 miles (1.45 to 3.56 km). These early tracer tests show that travel times are uncharacteristically long for a karst aquifer and typically are detected at multiple resurgence points. This is indicative of a karst system with a significant slow-flow component connected to a rapid-flow component typical of more permeable, solutionally enlarged fractures. The tests also determined that dye recovery usually occurs following significant precipitation.

Tracer tests were also conducted at the Leetown Pesticide Site and the Jefferson County Landfill area (NUS, 1986, p. 5–48). Although these tests focused primarily on detecting leakage from the landfill with expected detection in the site monitoring wells, at least one test resulted in recovery at the Center (Gray Spring) 42 days after injection, which is slightly more rapid than reported by Jones and Deike (1981). The time required for tracer breakthrough indicates a flow velocity of 171 feet per day (52 meters per day), which is extremely slow compared to a more conduit-dominated karst aquifer such as that in Greenbrier County, West Virginia with typical velocities of about 1,000-10,000 feet per day (305 to 3,050 meters per day) (Jones, 1997). Each of the tracer test recoveries were found to coincide with precipitation events (NUS, 1986, p. 5-55–5-56) suggesting increased flow velocity or flushing after heavy rainfall.

These findings were also seen in additional tracer tests conducted in other areas of Jefferson County, West Virginia in 1987 and 1988 (Kozar and others, 1991). Heavy precipitation was found to mobilize dye that was being sequestered within the subsurface, allowing it to flow down gradient (Kozar and others, 1991). When not mobilized by precipitation, it is believed that a significant portion of the dye either remains relatively immobile within the subsurface or moves slowly down gradient through
the less permeable, diffuse-flow matrix of interconnected fractures that are drained by the more permeable solution conduits (Kozar and others, 1991).

**Tracer Tests**

Eight dye tracer tests were conducted during 2004-2005 in the Leetown area (Fig. 23) as part of this investigation to provide additional data on groundwater flow paths, determine typical groundwater flow velocities in the conduit dominated portion of the aquifer, to better understand the role of conduits and large solutionally interconnected fractures in conveying ground water, and to refine the conceptual model of groundwater flow in the Leetown area. These tests were conducted using fluorescent tracers, primarily Rhodamine WT (C.I. AR 388) and to a lesser extent sodium fluorescein (C.I. AY 73).

![Figure 23. Groundwater flow connections determined from dye tracer studies conducted in the Leetown area, West Virginia.](image)

Dye was either released into surface streams (Tests 1 and 2) or injected into piezometers in highly permeable bedrock (Tests 4 and 6) or sinkholes (Tests 3, 7, and 8) (Fig. 23). An additional tracer test (Test 5) was conducted by injecting dye into a hole augered in a dry stream bed coincident with a thrust fault (Fig. 23). Except for the stream tracer tests, which were conducted to assess the potential for stream leakage into ground water, all injections consisted of mixing and delivering the dyes into the piezometers or sinkholes with a constant stream of water, followed by flushing with 2,000 gallons
(7,570 L) of water to assure dye entry into ground water. Suspected resurgence points, consisting of down-gradient springs, wells, and streams (Fig. 23), were then monitored by grab samples or by use of automated water samplers on a regular schedule. The sampling interval ranged from less than one per hour to one per day depending on the length of the test and distance from the point of injection. A prime emphasis of the tracer tests was to determine transport rates and trajectories from injection points to recovery sites. No effort was made to determine tracer mass recoveries because the manner in which water is pumped from the Center springs made it impossible to establish accurate discharge estimates needed to assess flux changes.

**Results of Tracer Tests**

Tracer tests conducted as part of this investigation agree well with the groundwater flow velocities and directions obtained by Jones and Deike (1981) and NUS (1986). The data provided by these tracer tests indicate relatively slow groundwater flow velocities when tracer dye was released into sinkholes and augured holes (See Kozar and others, 2007b, table 14). Coupled with the relatively poor tracer recoveries obtained, this indicates that much of the tracer dye is retained in the epikarstic zone, flow in phreatic solution conduits may be blocked by fallen rocks or sediment, or both. [It is also possible that some of the dye is retained as recharge to the diffuse fracture network.] These slow flow velocities and weak recoveries are not typical of most karstic terranes but are easily explained.

More rapid transit times and generally stronger recoveries were obtained when dye was released in the losing stream reaches providing evidence that losing stream reaches provide some recharge to the Center springs and wells. Leakage through losing stream beds complicated the design of the first two tracer tests and led to the modification of the Efficient Hydrologic Tracer Test Design (EHTD) program which was used for this project to assist in the design of the tracer tests (Field, 2003, 2006). It is possible that leakage from the losing branches of Hopewell Run caused groundwater to accumulate in subsurface flow channels beneath the site slowing groundwater transit times and resulting in poor dye recoveries when dye was released into sinkholes and augured holes. The effect of leakage of streamflow to subsurface flow channels would be to reduce the flow of the dye released in the sinkholes or augured holes and dilution of the dye resulting in very low recovery concentrations.

A significant issue associated with the dye-tracing study relates to the repeated use of Rhodamine WT for most of the traces (sodium fluorescein was only used for Tracer test No. 6). The potential for recovery of dye from earlier traces (Rhodamine WT and sodium fluorescein) may confound the results. (It should be noted that all rhodamines have been found to degrade to a greenish compound that fluoresces very near to that of sodium fluorescein.) However, the potential cross-over effect is not considered to be a major problem because dye injections were intentionally staggered between the north and south sides of the east branch of Hopewell Run to avoid cross contamination. On the date that both sodium fluorescein and Rhodamine WT were released, a large separation distance coupled with injection locations north and south of Hopewell Run were used to prevent cross contamination.

Most significant in addressing potential cross contamination was the reliance on conducting dye analysis on water samples using a fluorometer. Dye concentrations in water are additive so increasing background is not a concern when directly analyzing water samples for dye. In this sense, a positive dye recovery may be established if the recovered dye peak(s) represent a proportionally significant increase over the background.

Travel times obtained for this study were found to be comparable with those obtained by Jones and Deike (1981) which further leads to confidence in the results obtained for this study. Although some of the transport velocities obtained for this study were slightly faster than those obtained by Jones and
Deike, the difference is most likely a reflection of differing hydrologic conditions which facilitated faster transport velocities.

Consider, for example, on April 20, 2004, 288 g of Rhodamine WT was released into the south branch of Hopewell Run. The vast majority of this release was recovered downstream at the Hopewell Run at Leetown gaging station on the south branch of Hopewell Run where peak recovery exceeded 15 μg/L and Balch Spring where peak recovery exceeded 10 μg/L, but where a reasonably good recovery was also obtained from the ball field well where peak recovery exceeded 1.5 μg/L (Fig. 22). On May 17, 2004, another 288 g of Rhodamine WT was released into the east branch of Hopewell Run. The vast majority of this second dye release was recovered at the downstream Hopewell Run at Leetown gaging station on the east branch of Hopewell Run where peak recovery was approximately 6 μg/L (Fig. 23) and a reasonably good recovery was obtained at Blue Spring where where peak recovery was approximately 3 μg/L (Fig. 23).

Spring Recharge Delineation

An analysis of potential recharge areas to the Blue and Gray Spring complex and to Balch Spring was conducted as part of development of the groundwater flow model for the Leetown area (Kozar and others, 2007a). Results of the previously discussed dye tracer tests were also used to better define potential recharge areas to the Center’s springs and well fields. The probable and possible areas of recharge to the springs are shown in figure 24. Available hydrogeologic data and results of the tracer tests and current groundwater flow model indicate primary recharge areas for Balch Spring to the east and north, for Blue Spring to the south, and for Gray Spring to the east and southeast (Fig. 24). It is possible, although unlikely, that an undetermined amount of ground water may flow towards Balch Spring from the area south of Kearneysville, West Virginia, in the northern part of the model area (possible recharge area on Fig. 24). Also, the likely contributing area of recharge to Balch Spring is much larger than that of either Blue or Gray Springs. This is due to the fault which crosses the Center and is connected to Balch Spring, providing an added area of potential recharge to the spring that would not normally be considered if the fault were not such a dominant groundwater flow path.

Due to the nature of bedrock fractures in the area, it is difficult to assess the exact recharge areas contributing water to Blue and Gray Springs. Results of a particle-tracking analysis conducted as part of the development of the groundwater flow model indicate that areas to the south are likely to contribute water to Blue Spring and areas to the east are likely to contribute water to Gray Spring. It is probable that the recharge areas to the springs overlap and may shift with changing hydraulic head within the aquifer. While there is some uncertainty in the results of current (Fig. 23) and past dye tracer tests (Jones and Deike, 1981; NUS, 1986) conducted in the Leetown area, they generally confirm the results of the mathematical simulation of groundwater flow to the springs.

Another potential problem in delineation of recharge areas to Blue and Gray Spring is documented flow of ground water across the East Branch of Hopewell Run from an area northeast of the Center (Leetown sinkhole on Fig. 23). This appears to contradict the general directions of groundwater flow indicated by the particle-tracking analysis. It does, however, confirm the connection established by a previous tracer test conducted a few decades earlier (Jones and Deike, 1981). The reason for this anomalous connection is uncertain but is most easily explained by leakage of water from the East Branch of Hopewell Run to thrust faults that trend southwest to northeast through the area and connect with Blue and Gray Springs. Additional data would be needed to better quantify the contribution of ground water to Blue, Gray, and Balch Springs. Even with these data; it may be difficult or impractical to determine the exact areas contributing water to the various springs. While a better understanding of the connection between ground water and leakage of water from tributary streams is important, the
general groundwater recharge areas delineated here should be more than adequate for purposes of watershed management and protection.

Figure 24. Estimated recharge areas of Balch, Blue, and Gray Springs during average hydrologic conditions, Leetown, West Virginia. [Recharge areas delineated are approximate and vary with changing hydrologic conditions.]

Audio-magnetotelluric survey

This part of the field trip guide is abridged from Pierce (2007).

Audio-magnetotelluric Methods

Audio-magnetotelluric soundings are made to determine variations in the electrical resistivity of the earth with depth (Cagniard, 1950, 1953; Tikhonov, 1950; Wait, 1962; Keller and Frischknecht,
The AMT method uses natural-source multi-frequency electromagnetic signals from lightning or atmospheric disturbances as an energy source. These natural signals diffuse into the earth and the diffusion governs the electromagnetic (EM) induction. For this survey, a controlled-source transmitter was used to supplement natural source EM signals when signal strength from the natural source was low.

These AMT soundings consist of electric and magnetic field measurements over a range of frequencies from 10- to 100,000-Hertz with fixed receiver and transmitter locations. The distribution of currents induced in the earth depends on the earth’s electrical resistivity, earth’s magnetic permeability, and frequency measured. Since low-frequency signals penetrate to greater depths than high-frequency signals, measurements of the electromagnetic (EM) response at several frequencies contain information on the variation of resistivity at depth. These soundings are inverted and stitched together to form pseudo sections at various angles to the geologic structures. For a more complete discussion of AMT theory the reader is referred to publications by Weaver, 1994; Simpson, and Bahr, 2005; Mackie and others, 1997; Tikhonov, 1950; Cagniard, 1950, 1953; Swift, 1962; and Gamble and others 1979a, 1979b.

During calendar years 2003, 2004, and 2005, fifty-three AMT soundings were collected within and outside of Center property (Fig. 25). Three soundings were parallel sensor tests of the AMT instrument providing a total of fifty AMT soundings. The soundings were collected to improve understanding of the geologic structure and hydrology, as well as provide widespread vertical resistivity information for the Center area (Fig. 25). AMT soundings were recorded using a Geometrics EH-4 system. No vertical magnetic field (Hz) data were recorded because the system is limited to four channels (Ex, Ey, Hx, and Hy). About 2,000 frequencies were collected and then reduced to approximately 40 frequencies for each direction (Ex and Ey) from 10- to 100,000-Hertz. The EH-4’s system includes a 60 Hz notch filter to reduce power-line interference. Inversions that are invariant with rotation angle were chosen to assess the data and models in order to compare them with the known geology and hydrology.

Limestone rocks in the Leetown area are broken into roughly orthogonal blocks that have weathered along fractures. Soils and clays (karstic weathering products) between the fractured limestone blocks are not as permeable as the fractured rocks and are more electrically conductive than the limestone blocks.

Once a series of 2-dimensional models are generated, resistivity cross sections and associated maps can be made of different depths below land surface. The electrical sections and maps at various depth slices are compared with the known surficial geology, borehole data, and hydrologic information. Additional hydrologic and geologic interpretations can be made and the interpretations modified.

Results of AMT Surveys

AMT profiles and maps of inverted data reveal gradients and areas of high and low resistivities coincident with the prevailing geologic strike, N. 20-E. (Figs. 25 and 26). Low resistivity examples were measured along a syncline and near faults and fractures, especially thrust and cross-strike normal faults. Springs show up as low resistivity areas due to the increased porosity and higher fluid content in the rock. Even with the uncertainty of interpreting electromagnetic data in a two- to three-dimensional environment, electrical sections correlate well with the mapped geologic units, geologic structure, and the sub-surface hydrology. Resistivity lows positively correlate with areas (Figs. 24 and 25) where water is near-surface (near springs) or near high yield wells verified by drilling or aquifer tests (syncline well – Jef-0588).
Figure 25. Two-dimensional invariant electrical-resistivity section (LL1) across the USGS Leetown Science Center, West Virginia. The resistivity low on the left of the section is coincident with steeply dipping (60º E.) thrust faults and the low on the right is coincident with a synclinal fold and the well (Jef-0588) with the highest transmissivity on the Center. The broad resistivity high is coincident with an area of exposed low-permeability limestone ridges. ft, feet; NAVD88, North American Vertical Datum of 1988; AMT, audiomagnetotellurics. Figure taken from Kozar and others (2007b).

A map view of the invariant resistivity model at a depth of 197 ft below land surface shows correlation to features identified by surface geologic mapping (Fig. 26). A general pattern of high resistivity in most areas underlain by the Stonehenge Limestone and lower resistivity in areas underlain by other units demonstrates differences in resistivity between the lithologic units. Faults appear as linear anomalies or gradients separating areas of relatively high or low bedrock resistivity.

The resistivity map (Fig. 26) indicates three areas of profoundly lower resistivity. Anomaly 1 is an area in the north central part of the study area and coincides with a synclinal feature in the Rockdale Run Formation that underlies the Stonehenge Limestone. This resistivity low is interpreted as groundwater saturating and filling the trough of the syncline with perhaps some solution enhancement of the porosity of the bedrock, especially the Rockdale Run Formation. One of the monitoring wells, drilled in the axis of a syncline (Jef-0588), is located in this anomaly and is one of the most productive wells based on aquifer tests conducted on the Center and in the surrounding area (McCoy and others, 2005b). A second area of low resistivity (Anomaly 2) is located at the north edge of the study area. This area may indicate a groundwater saturated fracture zone, or may be the result of edge effect from contouring the data, as it is outside of the extent of the AMT stations. A third area (Anomaly 3) lies beneath the Center facility in the western part of the study area. This area has enhanced groundwater flow and storage associated with springs, multiple thrust faults, and two cross-strike faults. After the initial report on the AMT survey was published (Pierce, 2007), results of other investigations in the area suggest that a holding pond located just to the east of Anomaly 3 is leaking water, caused by evaporation to be more electrically conductive, into the groundwater. This leakage of conductive water may be another factor producing low resistivity Anomaly 3.
AMT apparent resistivity data in the study area can be correlated with the mapped geology and provides a series of useful cross sections and map images around the Center. These data can be used to identify areas of low resistivity that correspond to productive areas for wells. Potential productive areas for development of additional wells include the synclinal axes of fold structures, especially overturned structures. In this case, cross-strike faults also enhance the productivity of wells drilled near or through such structures. Thrust faults with enhanced fracturing may enhance well yields, but conduit development along some faults may preclude water-supply development as conduit dominated areas may yield water with high turbidity levels that are unacceptable for the Center.

![Composite map showing the 2-dimensional audio-magnetotelluric resistivity at 197 ft below land surface, surficial geologic map, and USGS topographic map of the Leetown area, West Virginia. (White numbers are low-resistivity anomalies; anomaly 1 is the productive syncline area, anomaly 2 is a potential productive area but may be influenced by grid edge effects, and anomaly 3 is the productive area near Balch Spring), Figure taken from Kozar and others (2007b).](image-url)
Ground water flow

The conceptual model of groundwater flow in the Jefferson and Berkeley County, West Virginia, areas has been modified on the basis of borehole- and surface-geophysical data and geologic mapping conducted as part of this investigation. A groundwater flow model developed as part of this investigation (Kozar and others 2007a) also helped test and revise the conceptual model of groundwater flow in the Leetown area. Focused recharge is a dominant process only when surface runoff occurs, typically as a result of intense local rainfall. While focused recharge to sinkholes can and does occur, the dominant process of groundwater recharge is areally diffuse precipitation over a broad area. Precipitation falling on the surface quickly infiltrates the soil and regolith and percolates into the upper epikarst, a zone of intense weathering that extends from land surface to a depth of approximately 30 to 60 ft.

The groundwater flow system is a triple-porosity aquifer matrix with intergranular porosity and small micro fractures providing minor storage of water. A dominant set of diffuse fractures provide the majority of storage, and a system of solutionally enlarged fractures act as drains for the intricate network of primary and secondary porosity features (Kozar and others, 2007a). The majority of solutionally enlarged conduits occur in a relatively thin zone within the epikarst but some smaller conduits occur at depths as great as 200 ft (61 m) below mean sea level. The epikarst is characterized by the presence of solutionally enlarged bedding plane separations and high angle joints which allow rapid infiltration of water to the deeper bedrock aquifer. Below the epikarst, a zone of less weathered bedrock is present.

This zone of moderately fractured bedrock does not typically contain a high density of solutionally enlarged conduits and the hydraulic conductivity is approximately 2 or 3 times less than the epikarst. Below a depth of about 250 ft (76 m) below mean sea level, the aperture of bedrock fractures decreases significantly, and the estimated hydraulic conductivity, from aquifer test and surface geophysical resistivity data, is approximately half that of the intermediate zone (Kozar and others, 2007a).

Flow of ground water through the epikarst can be rapid, on the order of weeks, especially if flow is concentrated in solutionally enlarged conduits. Flow within the intermediate zones is much slower, with estimates of groundwater age for the carbonate rocks in the region ranging from 15 to 50 years (McCoy and Kozar, 2007). A more recent study on spring water ages at the Leetown Center was published by Plummer and others (2013), with ages averaging about 2 years in 2004 following a wet climatic period (2003–2004), and ages in the range of 4–7 years in periods of more average precipitation (2008–2009). There are few data to approximate the age of ground water in the deeper portions of the aquifer. It is likely that ground water flows slowly at depths greater than about 300 ft below mean sea level; one chlorofluorocarbon (CFC) analysis of water from a 475 ft (145 m) deep well on the Center indicated an apparent groundwater age of approximately 50 years. Water from this well is likely a mixture of some older and younger components and water from greater depths is likely much older (Yager and others, 2013).

Topography also has a major effect on groundwater flow. Depth to water on hilltops is typically greater than that in valley or hillside settings. Uplands usually are in areas of more resistive rock. The bedrock underlying the Center comprises sedimentary rocks, chiefly limestone, dolomite, and shale. The bedrock, especially low permeability lithologic units such as the Martinsburg Formation and the Conococheague Limestone (Fig. 19), controls groundwater flow by acting as a barrier to water moving down the hydrologic gradient and across the strike of the bedding. This retardation of cross-strike flow is especially pronounced in the Leetown area, where the bedding typically dips at steep angles. Geologic structures that disrupt the rocks in cross-strike directions, especially highly permeable cross-strike faults
and fracture zones, provide avenues through which water can flow laterally across or through strata with low primary permeability. These rock formations also typically have lower hydraulic conductivities than the more permeable bedrock in lowland areas. The Conococheague Limestone and Martinsburg Formation, and to a lesser extent the New Market Limestone (Fig. 19), are the principal lower-permeability units which tend to act as barriers to water flowing down topographic gradients (Kozar and others, 2007a). Within the study area, local groundwater flow is primarily from the topographically higher uplands to the east towards Opequon Creek to the west. Although groundwater flow in lower permeability units is impeded on its westward path, cross-strike faults and oblique faults provide avenues along which ground water can flow either across or through the less permeable units. The lower permeability Martinsburg Formation exerts a dominant control on groundwater flow, impeding westward flow of groundwater towards Opequon Creek and forcing water to flow along solutionally enlarged faults which cross the Center (Kozar and others, 2007a). The hydrogeologic setting of the Leetown and Center area is desirable from a groundwater development perspective, as large quantities of ground water are funneled through the area. This is the primary reason for the highly productive springs that have historically supplied ground water for Center operations.

Numerical groundwater flow model for the Leetown area

This section of the field trip guide is abridged from Kozar and others (2007a).

Various methods were used to develop the numerical groundwater flow model for the Leetown area. Extensive data from aquifer tests conducted in Berkeley and Jefferson Counties (McCoy and others, 2005a;b), including the Leetown area, were analyzed to provide the hydraulic data (transmissivity, hydraulic conductivity, and storativity) needed for model development. These tests were conducted as part of fracture-trace and lineament analysis projects to assess hydrogeologic factors controlling the distribution of high-yield wells in the complex karst aquifer system of the region. A streamflow-gaging station (Fig. 27) was constructed on Hopewell Run to provide discharge data for calibrating flow components of the model, for developing a budget of water resources, and to determine recharge for the Leetown area by hydrograph analyses (Rutledge, 1998).

The computer software package Visual MODFLOW, version 4.0.0.131 (Waterloo Hydrogeologic, 2004), was used to develop a steady-state groundwater flow model of the Leetown area. Visual MODFLOW is a commercially derived graphical user interface to the USGS MODFLOW 2000 three-dimensional finite-difference groundwater modeling software (Harbaugh and others, 2000) and to MODPATH (Pollock, 1998), a USGS particle tracking software package. MODFLOW was used to simulate groundwater levels and provide a hydrologic budget for water in the Leetown area. The MODPATH interface was used for particle-tracking analyses to identify potential recharge areas to the springs and wells at the Center.

The use of a three-dimensional numerical model such as MODFLOW for simulating groundwater flow in a karst setting is a subject of much debate. Because MODFLOW assumes an equivalent porous media, its use in simulating groundwater flow in complex karst hydrogeologic settings may be inappropriate for many karst aquifers. This is especially true for cavernous and large conduit dominated karst systems such as that found in the Mammoth Cave area of Kentucky. MODFLOW, however, may be effectively used to simulate groundwater flow in the karst in the Leetown area. Although the aquifer does exhibit some karst features, sinkhole development is sporadic and caverns when encountered are limited in length and width.
Figure 27. Location of study area and boundary conditions for the groundwater flow model of the Leetown area, West Virginia. Figure from Kozar and others (2007a).
The majority of the rock mass in the region, especially in upland areas, is drained by an interconnected network of bedrock fractures with little or no solution development. The small conduits which develop, predominantly in low-lying areas, act as drains for the interconnected fracture network in the more areally extensive fractured rock mass. Thus, the conduits may be simulated as a network of interconnected drains by assigning higher hydraulic conductivities to these more permeable features and the aquifer can effectively be simulated by an equivalent porous medium approach using MODFLOW. It is necessary to simulate the anisotropy in the aquifer which can easily be done using MODFLOW.

Groundwater Flow

When model calibration was completed, analyses of groundwater flow were conducted to (1) develop a water budget for the Leetown area to assess effects of current and future groundwater withdrawals on long-term water availability to the Center, (2) evaluate hydraulic heads simulated by the model to better understand groundwater flow in the watershed and to delineate the recharge areas to the Center’s principal springs, (3) simulate drawdown from production wells and assess potential pumping effects on the Center’s springs, and (4) assess effects of drought on water availability by reducing the recharge to the model and comparing simulated water levels for pumping periods to water levels simulated under pre-pumping conditions. The groundwater flow model prepared for this investigation is a steady-state simulation concentrated on the Leetown area but also incorporates a larger area to define hydrogeologic boundary conditions. The focus of the model is the 455-acre USGS Center and adjacent USDA land (Fig. 27). The Hopewell Run watershed from its headwaters to the gaging station at Leetown (drainage area of 8.95 mi²) was also an important part of the simulation. Areas of less importance include the minor tributary streams that drain to Opequon Creek and the area west of the Center, which includes areas where the Martinsburg Formation and Chambersburg Limestone crop out. These areas were included to provide distinct, easily delineated boundary conditions for the model and because of their importance in controlling groundwater flow in the Leetown area.

Design and Assumptions

The groundwater flow model is based on the conceptual model of groundwater flow previously discussed. It consists of three layers to simulate 1) the epikarst zone; 2) the primary intermediate zone in which most wells are completed; and 3) the less fractured, deeper portion of the bedrock aquifer (Fig. 28). Water-level and streamflow data collected from May through November 2004 provided the basis for development and calibration of the model. Assumptions were made for areas where data were limited or unavailable (mostly peripheral areas of the model) and with respect to the overall depth reached by groundwater flow. Aquifer tests on and adjacent to the Center indicate that the faults that cross the area tend to have higher hydraulic conductivity than the individual lithologic units. This area also exhibits hydraulic properties more characteristic of a homogenous, isotropic aquifer with multiple faults and regolith acting as a uniform aquifer system rather than as discrete fractures. Therefore, the area near the Center was simulated as a broad, highly conductive area within the model. Because no data were available to characterize the peripheral areas of the model, faults in these areas were assumed to exhibit hydraulic properties similar to those near the Center. The minimum base elevation of the Potomac and Shenandoah Rivers near the study area is approximately 240 ft (73 m) above mean sea level at their confluence in Harper’s Ferry, West Virginia. Because rivers are the primary discharge zones in the region, it is unlikely that substantial groundwater flow occurs at depths much greater than the base level of these major rivers; however, the groundwater flow model was extended vertically to sea level to account for any deeper groundwater flow that may occur. The average depth of
groundwater flow simulated was approximately 450 ft (137 m) below land surface. Land-surface elevations simulated in the model range from a minimum of 395 ft (120 m) in the northwestern part of the model along Opequon Creek to 628 ft (191 m) above NAVD 88.

Grids, Layers, and Boundaries

The model grid extends from the Kearneysville area to the north (Fig. 27), to Middleway to the south, and is bounded by Opequon Creek to the west and by bedrock ridges near Johnsontown and Browns Corner to the east. The model consists of three layers that represent the three layers in the previously discussed conceptual model of groundwater flow (Figure 28). Hydraulic conductivity is highest in the top layer which represents the epikarst (Fig. 28) and extends from land surface to a depth of 100 ft (30.5 m). The middle layer represents the fracture-dominated portion of the aquifer in which most wells are completed (Fig. 28 and 29); it represents 200 ft (61 m) of bedrock and was assigned hydraulic conductivities approximately 2.0 to 2.5 times lower than that assigned to layer 1 based on results of aquifer tests conducted in Jefferson and Berkeley Counties. The third and lowermost layer (Fig. 28) represents mostly fractured rock with low permeability and little or no assumed conduit development. This layer represents approximately 150 ft (48 m) of bedrock, and was assigned hydraulic conductivities approximately half that of the middle layer based on results of resistivity surveys and aquifer tests. The lower layer of the model extends to sea level in most areas.

Figure 28. Layers simulated in the groundwater flow model of the Leetown area, West Virginia, and their approximate elevations referenced to NAVD 88. [Dark areas are active nodes and light areas are inactive nodes within the model].

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Figure 29. Hydraulic conductivity assigned to geologic formations, faults, and fracture zones in the middle layer of the groundwater flow model of the Leetown area, West Virginia. Figure from Kozar and others (2007a).
Groundwater simulations

Various scenarios were simulated to estimate the potential effects on future water-supply alternatives. The model was initially developed and calibrated based on recharge estimated from streamflow data collected at the Hopewell Run streamflow-gaging station from October 2003 through September 2006, results of aquifer tests conducted in Jefferson and Berkeley Counties over the same period, and water-level data collected in the summers of 2004 and 2005. The model was developed as a steady-state simulation. Analysis of head equipotential lines and flow paths from particle-tracking analysis clearly illustrates that surface and ground water over a very broad area are funneled through the Leetown area, including the area occupied by the Center, and are responsible for the large springs at the Center.

The model was then adjusted to simulate current conditions with production wells withdrawing water at average annual rates based on records of pumping maintained by the Center staff. The purpose of the simulation was to assess the effects of current groundwater withdrawals on streamflow and drawdown in the vicinity of existing wells and springs. Analysis of drawdown caused by the current production wells and pumping from Balch Spring indicates that current operations (in 2006) have a minimal effect on groundwater and surface-water resources in the area under normal conditions. Current pumping of ground water diverts 0.51 ft³/s (14.4 L/s) or 4.7 percent of available streamflow under normal climatic conditions, produces less than 0.10 to 0.20 ft (3.0 to 6.1 cm) of drawdown around Balch Spring, and approximately 0.10 ft (3.0 cm) of drawdown in the vicinity of Blue and Gray Springs.

The model was then used to assess the potential effects of the addition of two new production wells at the Center with average withdrawal rates of 50 and 100 gal/min (189 and 378 L/min), respectively. With an additional 150 gal/min (568 L/min) augmenting existing supplies, drawdown of approximately 0.30 to 0.40 ft (9.3 to 12.4 cm) was simulated near Balch Spring and 0.10 ft near Blue and Gray Springs. This represents an additional 0.10 ft (3.0 cm) of drawdown near Balch Spring. The simulated net effect on streamflow is approximately 0.83 ft³/s (1,410 L/min) compared to non pumping conditions, or 7.7 percent (an additional 3 percent compared to current withdrawals) capture of ground water that would have discharged to the stream as base flow. Most of the streamflow captured is groundwater that would have discharged to Hopewell Run or its tributaries. Also, the captured groundwater is only temporarily diverted, because most of the groundwater withdrawn (minus negligible evaporative losses) is returned to the stream by the wastewater-treatment plant at the Center.

Similarly, with an additional 200 gal/min of withdrawals (100 gal/min each) simulated from the two new production wells to augment existing supplies, 0.50 ft of drawdown was simulated near Balch Spring and 0.10 ft of drawdown near Blue and Gray Springs. This represents an additional 0.40 ft of drawdown near Balch Spring and 0.10 to 0.20 ft of drawdown (compare to non pumping conditions) near Blue and Gray Springs. The net effect on streamflow (compared to non pumping conditions) is 0.93 ft³/s or 8.6-percent (an additional 3.9-percent compared to current withdrawals) capture of ground water that would have discharged to the stream as base flow.

The potential effects of two large subdivisions were assessed by simulating a production well in the northern portion of the study area withdrawing water to supply 300 homes and a production well in the southern portion of the area withdrawing water to supply 500 homes, for a combined additional withdrawal of 117 gal/min. Simulation results indicated an additional drawdown of 0.10 ft near the springs and wells at the Center under normal climatic conditions and current demands increased by an additional 200 gal/min in withdrawals from two new production wells on the Center and 117 gal/min withdrawals from the simulated subdivision wells. Simulated streamflow in Hopewell Run was lowered...
to 9.65 ft3/s or 10.9-percent compared to non pumping conditions (an additional 2.3 percent resulting from the subdivision wells). The simulation was conducted assuming interbasin transfer of water by means of a sewage collection system. The effect on water availability would be reduced if the water is returned either to Hopewell Run or by septic-return flows.

Finally, a drought was simulated by reducing recharge within the model to 8.3 in/yr, a rate that approximates the recharge during the prolonged 16-month drought that affected the region from late 1998 to early 2000. In these simulations, existing wells were pumped at current rates but two new production wells were pumped at rates of either 50 and 100 gal/min, 75 and 100 gal/min, or 100 gal/min each. Regardless of the specific scenario simulated, drawdown produced by the augmented withdrawals during a drought is substantial in the vicinity of the Center’s springs. Simulated drawdown in the vicinity of Blue and Gray Springs ranges from 0.3 to 0.5 ft and drawdown in the vicinity of Balch Spring ranges from 0.6 to 1.0 ft for the withdrawal scenarios. This amount of drawdown would substantially decrease the flow of Balch, Blue, and Gray Springs and the streamflow in Hopewell Run. Therefore, on the basis of the simulations, withdrawal of water at current rates combined with the additional 150 to 200 gal/min withdrawal from the new production wells likely is not sustainable for long periods during droughts. Close monitoring of water levels and streamflow during critical low-flow periods would help avoid adverse impacts of groundwater withdrawals.

Stop 5. C&O Canal at McMahon's Mill

(39°31′50.59″N, 77°49′24.25″W)

NOTE: No hammering of outcrops or sample collection of any kind is allowed in the C&O Canal National Park without a permit.

Solutional features within the bluffs of the Potomac River along the C&O Canal provide clues as to the initiation of conduits for cave formation in the Great Valley. The caves along the C&O Canal are generally formed along intersections between joints and bedding plane partings. Passage cross sections are parabolic or elliptical, indicative of solution at significant depth under conditions of stress favorable for creation of these geometries (Davies, 1960). Preferred zones for the development of such early-formed solution features have been termed “inception horizons” (Lowe, 2000; Filipponi et al., 2009; Lowe and Waters, 2014). The rock exposures along the Potomac River were created in the mid-nineteenth century (between 1828 and 1850) during canal excavation; blasting revealed many conduits that had remained relatively free of the influence of surface weathering and erosion. Conduit initiation has taken place along both vertical and horizontal fractures, yet fracture intersections are most favorable for accentuated solution (Fig. 30). Several conduits exposed along the bluffs show evidence of calcium carbonate rinds and fills of calcareous cemented fine sediments. In places, distinct corrosion rinds, vesicular in texture and bright orange-pink colored, coat the conduit walls and are several millimeters thick (Fig. 31). These rinds indicate alteration of the carbonate host rock by chemically aggressive fluids. A qualitative X-ray diffraction (XRD) mineralogic analysis of the fill, corrosion rind, and unaltered rock showed that all three were composed primarily of calcite, with the unaltered rock and rind containing minor quartz and trace feldspar, while the fill contained a greater proportion of quartz, feldspar, dolomite, and trace (unidentified) clay. In places, the solutionally enlarged fractures have become partially filled with calcium carbonate re-precipitated as travertine. However, later solution is also frequently evident where the corrosion rinds and travertine have been partially removed.

In a well-exposed outcrop along the C&O Canal near Mc Mahon’s Mill, the bedding is nearly vertical, and solution occurs horizontally along joints and/or stylolites, as well as vertically along
bedding plane partings (Fig. 32). Sub-horizontal fractures tend to be the focus of greatest solution, and early-formed anastomoses often become occluded by sediment, and later by travertine.

Figure 30. Solutional conduits exposed along the bluffs of the C&O Canal. (A) John Repetski points to enhanced solution that occurs at the intersection between vertical and horizontal fractures in dolomitic limestone of the Rockdale Run Formation; here, bedding is subvertica and solution initially occurred along a sub-horizontal joint. (B) Closeup of section in A, showing upward paragenetic solution guided by vertical bedding plane partings, having originated along the horizontal joint that became occluded with sediment and calcite cement prior to the paragenetic solution phase. (C) Elliptical profile of a solution conduit along sub-vertical fractures near a fault zone (conduit is ~6.6 ft [2 m] tall).

A number of explorable caves exist on both sides of the Potomac River within the Cambrian and Ordovician carbonates (Davies, 1958; Franz and Slifer, 1971; Tudek and Vesper, 2011). These caves are generally limited in extent, and passages are generally occluded by fine sediment within a short distance of the river. However, some caves have been excavated of their sedimentary fill material as a result of sinking surface flow. Howell’s Cave is one of these, where an ephemeral free-surface stream exits the entrance to the cave. This cave occurs near the axis of a broad syncline, where axial-planar fracturing has provided preferential pathways for flow and bedrock solution (Fig. 33).

A supposed sequence of events that may have formed the small caves along the C&O Canal is: (1) early solution occurred along fracture planes that left behind corrosion rinds and fine sediment fills that were subsequently cemented with calcite; (2) preferred solution along sub-horizontal fracture partings and at fracture intersections; (3) enlargement by focused flow after breakthrough into the turbulent regime along more isolated flow paths, yet still deep within the aquifer under stress conditions favorable to elliptical and parabolic cross sections. A latest stage of conduit enlargement may have occurred through paragenesis as fine sediments continued to fill the passageways, and solution occurred upward into the bedrock thereby enlarging the early-formed conduits into cave passages (Fig. 34). Given the geometric character, sedimentary fills, and degree of calcite precipitation within the conduits and caves along the C&O Canal, it would seem that these solutional features did not form, for the most part, due to back flooding of the Potomac River.
Figure 31. Sediment fill and vesicular bedrock corrosion rinds that are cemented with calcite frequently occur within solutional conduits along the C&O Canal. (A) This vertical solution conduit is filled with fine-grained sediments that had become cemented with calcite to form a very hard rock, now partially weathered into blocks (circled in white dashed line) floating within clay-loam fill. (B) Rinds of the initially corroded bedrock remain in only a few places, having survived later solution. (C) The early-formed corrosion rinds are vesicular in texture and vuggy, and are cemented with calcite; note the sharp contact between the intact bedrock and the corroded bedrock at the apex of the fracture where solution took place. (D) A hand sample from which a thin section was made to examine the corrosion rind at the bedrock surface. (E) A scan of the thin section, illustrating the vuggy, red-orange corrosion rind at the upper left of the thin section image.

Figure 32. Vertical rising solutional forms in bedrock and anastomoses along a sub-horizontal joint in the bluffs along the C&O Canal near McMahon’s Mill. Bedding is near vertical as shown by the bedding plane surface to the right of the field notebook. The reddish-orange sediment fills are cemented with calcite to the bedrock walls.
Figure 33. Geologic sketch from McMahon’s Mill near milemarker (MM) 88 westward to Lock 41, depicting rocks of the older Conococheague Formation thrust above limestone of the younger Rockdale Run Formation. The rocks are folded into a syncline in the footwall of the thrust fault. There are several limestone quarries and caves in the bluffs, including Howell’s Cave, which is shown near the axis of the syncline in this sketch (modified from Southworth et al., 2008).

Figure 34. David Weary and Robert Mason at the entrance to Dam Number 4 cave along the C&O Canal. This cave is an enlarged cavern due to the focused flow of surface water sinking and flowing through the early-formed phreatic proto-conduits, and ultimately forming a large paragenetic cave passage.

References


