

TRIP A4

KARST OF THE SCHOHARIE VALLEY, NEW YORK

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INTRODUCTION

Karst topography is well developed in a narrow band along the Helderberg Escarpment in Schoharie and Albany Counties, New York, where highly soluble Silurian and Devonian carbonate rocks are exposed (Figure 1). This is one of the finest examples of glaciated karst in the country. This field trip focuses on the valley of Schoharie Creek, which rises in the Catskills and drains northward to the Mohawk River. Where it intersects the carbonate belt, it is joined by Cobleskill Creek from the west and by Fox Creek from the east. Deep entrenchment by these streams has allowed a great deal of subsurface drainage to take place through the carbonate rocks, accompanied by karst topography and caves. During the latest phases of glacial retreat, about 14,000 years ago, a dam of glacial ice between the neighboring plateaus blocked the drainage and formed glacial Lake Schoharie, which inundated most of the caves and karst to depths as great as 150 m. The primary purpose of this field trip is to examine some of the major karst features of the area and how they have interacted with the effects of Wisconsinan glaciation. A secondary purpose of the field trip is to address cave resource protection considerations via land-use analysis and through use of the karst principles discussed in this paper. Further details about the Helderberg karst are provided by Kastning (1975), Baker (1976), Palmer (1976), Myroie (1977), Cullen et al. (1979), Palmer et al. (1991), Rubin (1991), Rubin (1995), Rubin et al. (1995), and Dumont (1995).

This field trip illustrates some of the major surface karst features of New York, including both surface karst and caves. Three different plateaus are visited, each with a different geomorphic setting and subsurface drainage pattern. No special equipment is needed for this field trip. However, many of the stops are on private property, and visitors at other times need written

permission. Caves that are not open to the public are rather dangerous and require special equipment and prior experience under supervised instruction. Appropriate contacts are listed in this guide.

GEOLOGIC SETTING

Stratigraphy

Strata exposed in the field-trip area range from Ordovician to Middle Devonian (Figure 2). In terms of karst development, these can be grouped into a few major sequences, described here only in terms of their geomorphic significance. Typical thicknesses are shown in Figure 2. The main karst-forming units of the Schoharie Valley are included in the Silurian-Devonian *Helderberg Group*, which overlies Ordovician-Silurian shales and sandstones. From bottom to top, the Helderberg Group includes the *Rondout Formation*, a shaly dolomitic limestone; the

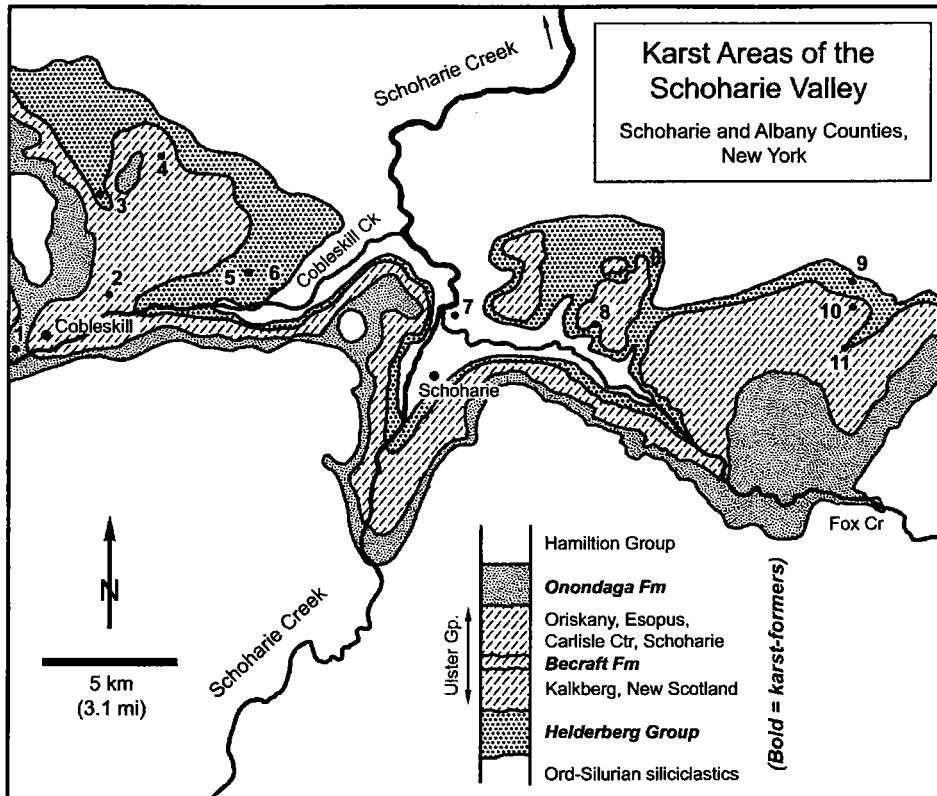


Figure 1: Geologic map of the field-trip area. Numbers indicate stops.

Period	Thickness (m)	Lithology	Formation	Group
Devonian	100-250		various shales and qtz sandstones	Hamilton
	30		Onondaga	Ulster
	2-5		Schoharie	
	15		Carlisle Center	
	17		Esopus	
	1		Oriskany	
	3-6		Becraft	Helderberg
	15-25		New Scotland	
	13-20		Kalkberg	
	15-20		Coeymans	
14-17		Manlius		
Silurian	2.5-12		Rondout	Helderberg
	0.3-3		Cobleskill	
	10-17		Brayman	
	500-700		Schenectady	

Figure 2: Generalized stratigraphic column in the Schoharie Valley region.

Manlius Limestone (Thacher Member), a thin-bedded limestone with thin shaly interbeds; the *Coeymans Limestone* (Dayville and Ravena Members), a massive, competent, cliff-forming limestone; the *Kalkberg Formation*, a poorly soluble, cherty, thin-bedded, shaly limestone; the *New Scotland Formation*, an even less-soluble shaly limestone, and the *Becraft Formation*, a thick-bedded limestone. The *Brayman Formation*, which underlies the Rondout, is a shaly dolomite that hosts very limited karst development, typically by entrenchment of cave streams below the Rondout.

Recently the Silurian-Devonian boundary in east-central New York has been redefined as a regional unconformity (Howe Cave Unconformity) that truncates progressively older strata from west to east (Ebert et al., 2001). In the western part of the field-trip area the unconformity lies between the Dayville and Ravena Members of the Coeymans Limestone. Farther east the Dayville is absent and the unconformity separates the Manlius from the Coeymans. This interpretation contrasts with that of Rickard (1962), who considered the Helderberg formations to represent correlative facies in an on-lapping sequence.

The *Ulster Group* is a sequence of mainly insoluble detrital rocks that includes the *Oriskany*, *Esopus*, *Carlisle Center*, and *Schoharie Formations*. Together they constitute a major barrier between karst groundwater in the Helderberg Group and the Onondaga Formation.

The *Onondaga Formation* consists of medium-bedded limestone with abundant chert beds. It is an important karst-former in Albany County, but in most of the Schoharie Valley it crops out mainly in steep slopes that are not conducive to karst. Large expanses of Onondaga are exposed in the eastern part of the field-trip area (see Figure 1), but the relief is low and known caves are small and few. However, solutionally enlarged surface fissures are common.

The *Hamilton Group* consists of siliciclastics that form the upper boundary of karst development in the area. It forms the steep slopes above the resistant Onondaga bench. East of the field-trip area, much of the groundwater recharge to the Onondaga is fed by runoff from this group. Sinkholes are common near the Hamilton/Onondaga contact.

Geologic structure

Throughout all but the eastern part of the field-trip area, the strata dip rather uniformly 1-2 degrees to the south-southwest. Minor structural flexures and depositional irregularities are superimposed but for the most part cannot be detected without precise surveys. There are two prominent joint sets with strikes of roughly N 22° E and S 70° E ($\pm 5^\circ$). Joints are essentially vertical, except for a few that have steep dips. Many small-displacement faults extend through the area, striking mainly WNW with dips of about 10-30 degrees either north or south. The most visible ones are simply ramping upward from larger underlying bedding-plane thrusts. Studies of the relationship between karst and geologic structure in the area include those of A. Palmer (1972), Gregg (1974), Kastning (1977), M. Palmer (1976), Mylroie (1977), Rubin (1991), Rubin (1995), and Rubin et al. (1995).

Geomorphology and Glacial Geology

The field-trip area is part of the Appalachian Plateaus geomorphic province. In contrast with other parts of the Appalachians, where the limestones tend to be valley-formers, in east-central New York the limestones are among the most resistant strata. Most of the insoluble rocks are incompetent shales and siltstones. Resistant sandstones are few and very thin. As a result, the Helderberg and Onondaga form two prominent structural benches.

The area is drained by the north-flowing Schoharie Creek and its major tributaries, Cobleskill Creek to the west and Fox Creek to the east. These two tributaries account for most of the karst development in the area by exposing the limestone in the down-dip direction (Figure 1). Without these stream valleys, karst development would be limited to a very narrow band along the Helderberg Escarpment.

The field-trip area offers a fine view of glacial derangement of karst. With substantially more glacial deposition, large areas of the karst would have been entirely buried. Wisconsinan ice reached its maximum extent about 22,000 years BP, with a thickness of about 1.5 km (Dineen, personal communication, reported by Rubin, 1991). The final ice retreat in the area was about 14,700 years b.p. (DeSimone and LaFleur, 1985). The main effects in the field-trip area are valley filling to a maximum of at least 30 m, partial or complete burial of small preglacial valleys by till, derangement of surface and subsurface drainage, glacial lake deposits, lineations in topography (drumlins, etc.), and development of meltwater channels. North-flowing Schoharie Creek was dammed by the retreating Wisconsinan glacier, creating glacial Lake Schoharie. LaFleur (1969) gives a sequence of elevations for Lake Schoharie ranging from 550 down to 230 m. Clays deposited in the lake are found in caves and on the surface throughout much of the lower Schoharie basin. At the surface and in caves these are found at elevations as high as 345 m. Along Fox Creek these clays were once used as a source of the bricks seen in many of the local houses.

The deepest stage of valley erosion in the field-trip area, now interred beneath as much as 100 m of glacio-alluvial sediment, is thought to be of "late pre-glacial age" (Dineen, 1987); but the age of higher-level erosional features is uncertain. Another effect of glaciation was to enlarge the effective catchment area feeding many caves. Invading glacial meltwaters, some of which were subglacial, may have significantly enlarged certain pre-glacial caves (Rubin, 1991).

LaFleur (1969) described four late-Wisconsinan glacial readvances in the Schoharie Valley. Ice invaded the area from the north and diverged at the Helderberg Escarpment. Drumlins and streamlined bedrock hills are oriented west-southwest in the field-trip area. The latest ice sheet receded from the area about 14,700 years ago (DeSimone and LaFleur, 1985).

North-flowing Schoharie Creek breaches the limestone between the Cobleskill Plateau and Barton Hill (Figure 1). Its valley was periodically dammed by ice during the latest stages of glacial retreat, forming glacial Lake Schoharie, which disappeared when the ice melted. This lake occupied the lower reaches of the Schoharie Valley in the field-trip area and most of the Cobleskill and Fox Creek Valleys. Hanging deltas, strand lines, and clay and marl deposits in the valleys indicate the former levels of the lake. Glacial geologists have asked whether there is evidence for the draining of lake water through caves, since the karst plateaus formed the northern boundary of the lake, but so far there is no evidence that it did.

There is considerable evidence that nearly all the caves and karst in the field-trip area pre-date the latest glaciation and are probably much older. Many caves and their feeder sinkholes are graded to valley levels now buried beneath glacial deposits. Cave sediments include not only the varved clays (rhythmites) but also large high-level gravel banks that speak of much greater water flow in the past. Speleothems in Schoharie Caverns exceed the 350,000-year limit of the uranium/thorium dating technique, indicating at least a mid-Pleistocene age (Dumont, 1995; Lauritzen and Mylroie, 1996). Many surface karst features are choked with glacial till. Scallops in some caves show evidence for discharges far greater than are available from their present drainage basin (Palmer, 1976, Rubin, 1991). Similarly, scallops in the bedrock walls of relict sinkhole insurgences indicate paleoflows far in excess of the drainage basin now available (Rubin et al., 1995). In many places karst drainage is being exhumed from its occluding glacial sediment.

KARST PRINCIPLES

Surface karst features, such as sinkholes, sinking streams, and large springs, owe their existence to the development of underground solution conduits (caves). When soluble rock is exposed in relief in a humid climate, groundwater selectively enlarges interconnected fractures, partings, and pores by dissolution, and a few flow routes eventually grow large enough to carry turbulent water. These highly transmissive conduits are generally fed by upland recharge and lead to outlets in nearby entrenched valleys. Laminar flow in surrounding unenlarged fissures and pores converges on the solution conduits. Some conduits serve as diversion routes for perched surface streams and may pirate the entire stream flow, leaving part or all of the surface channel dry. Sinkholes develop where paths of infiltration enlarge enough by solution that the soil subsides into the conduits and is carried away by turbulent groundwater. Sinkholes also form where a cave passage grows large enough to collapse. Unless the underlying conduits contain enough flow to carry detrital sediment, depressions in the bedrock surface tend to fill with overburden, revealing little or no surface expression. Caves and surface karst features therefore grow synchronously and interdependently. For further information on karst and cave development, see White (1988), Ford and Williams (1989), and Palmer (1991).

Solutional caves provide an important clue to the sequence and timing of geomorphic events in the area. It is a popular impression that caves are irregular pockets hollowed out of bedrock in a random sponge-like pattern. On the contrary, they consist of an orderly arrangement of discrete passageways that show great sensitivity to their structural, hydrologic, and geomorphic settings. Their most typical pattern is crudely dendritic, with sinkholes and other infiltration sources feeding tributaries that converge to form larger and fewer conduits in the downstream direction. The outlets are at lower elevations, generally near base level in entrenched valleys or perched at contacts with underlying less permeable strata.

In the vadose zone, above the water table, rivulets of water substantial enough to form caves are controlled by gravity. Passageways of vadose origin therefore descend along the steepest available paths. Where vertical fractures are available, the water forms vertical shafts, which are fissures or well-like voids with nearly vertical walls. Where the water is deflected from the vertical along inclined bedding-plane partings or faults, it tends to form downcutting canyon-like passages oriented down the dip. While these solution conduits are small initially, they become high and narrow, with sinuous bends controlled mainly by structural irregularities as they entrench downward. At the water table, the water loses its inherent tendency to follow the steepest paths and instead follows the most efficient routes to the nearest available surface outlet. Most phreatic cave passages are roughly strike-oriented tubes or fissures, which represent (in rather simplified terms) the intersection between the water table and the favorable parting or fracture that conducts the water. These initial openings diminish in width and number with depth, so most phreatic conduits form at or just below the water table, with some exceptions in tectonically disturbed areas. Even in presently dry caves, the transition from down-dip canyons to strike-oriented tubes is compelling evidence for a former level of diminished or interrupted valley deepening.

As rivers deepen their valleys, lower groundwater outlets become available, and the water table drops. New phreatic cave passages form at lower levels, and old ones either become vadose pathways or are abandoned completely. Groundwater patterns are greatly complicated in this way, because the old upper-level routes are temporarily reactivated during high flow and provide divergent paths for water. Younger passages can be formed by floodwaters (including glacial meltwater) above the normal low-flow water table. Drainage divides and flow patterns thus change not only with time, but also with flow stage.

There are exceptions to the rule that only a few select conduits achieve cave size. At the soil/bedrock interface, infiltrating water may contain so little dissolved carbonate that the water is solutionally aggressive enough to dissolve many interconnecting fissures at a rather uniform rate. The result is epikarst, a zone of enlarged fissures, either soil filled or open, in the top tens of meters of bedrock. The epikarst in New York may be entirely absent where it has been removed by glacial plucking or where lime-rich soil exhausts the solutional potential of the water before it reaches the bedrock. Another exception is where caves are fed by flashy recharge from sinking streams. During high flow, surface water pours into the caves, ponds upstream from passage constrictions, and is injected under steep hydraulic gradients into all available openings in the surrounding bedrock. Nearly all openings enlarge simultaneously, forming a maze of diversion passages around the constriction. Where vertical joints are prominent a network of fissures is produced, with a pattern like that of city streets. Where bedding-plane partings are prominent a braided (anastomotic) pattern of intersecting tubes is formed around the constriction.

Effects of Glaciation on Karst

Glacial effects on caves and karst include (1) changes in the rate and pattern of groundwater recharge, (2) changes in water-table level, (3) blockage or diversion of springs, accompanied by flooding and accumulation of sediment in their feeder caves, (4) changes in climate, affecting rates of solution, (5) partial filling of caves by glacial till, outwash, and lake deposits, and removal of some sediment by late-stage meltwater, (6) stagnation of groundwater in the vicinity of glacial lakes, (7) growth and modification of caves by subglacial and proglacial meltwater, (8) enlargement of fissures by glacial loading and unloading, and (9) development of now-relict surface channels and caves by meltwater.

These two powerful geomorphic agents – karst processes and continental glaciation – operated together in the field-trip area at different spatial and temporal scales. Karst is influenced by local drainage patterns and rock types and matures in time on the order of 10^5 years. Glaciation operates on a very broad scale (although with diverse local variations) in broad cycles with many smaller cycles of advance and retreat superimposed. The cycles of glacial advance and retreat that affected the New York karst had a time scale on the order of 10^4 years. Because of the shorter duration of glacial episodes, the effects of glaciation were mainly superimposed on preexisting karst systems.

Points to Ponder

Because the local karst is at least pre-Wisconsinan, several questions arise that rarely need to be considered in nonglaciated karst. They should provide considerable fuel for discussion during the field trip.

1. What happened to the karst features during glaciation? Did underground flow cease entirely? What was the geomorphic role of glacial meltwater?
2. How did preglacial drainage patterns compare to those of today?
3. What effect did glaciation have on base level and the position of the water table? Was there a potentiometric surface within the glacial ice, and did hydrostatic pressures in the ice translate into the underlying bedrock?
4. How did glaciation affect the groundwater geochemistry, both past and present? Do speleothems correlate only with interglacial periods? Can we relate karst features to Pleistocene climates?
5. Has there been substantial post-glacial karst development?
6. Do karst features contain clues to glacial events that cannot be recognized at the surface?

Recent Clues

Recent excavation on the Cobleskill Plateau may provide important clues to questions posed above. During excavation efforts, a relict sinkhole was uncovered beneath some 7 m of hard-packed till. The basal 2.5 meters is a low chroma dark bluish gray vs. a high chroma yellowish brown color of the overlying till. The gray color indicates a prolonged period of saturation or reduction (i.e., reduced iron) and either a raised potentiometric surface or water table condition. Post-glacial oxidation and aeration probably altered the color of the upper till. The till overlies a glacially striated limestone pavement indicating a glacial movement direction of S 68° W. A

sinkhole exposed within the limestone pavement is some 22 m long in the upslope direction, with a maximum width of about 2.5 m. The sinkhole is massively scalloped with small wavelength scallops, indicating turbulent, aggressive, water influx over a long period of time. The present glacially-deranged topography up gradient of this sinkhole provides only a small drainage basin, indicating that an alternate water source was once present (i.e., sub-glacial meltwater invasion beneath expansive glacier ice). Thus, the size of this partially occluded sinkhole indicates that today's surface runoff inputs are minor compared to formerly greater discharges below warm-based stagnant ice. Overconsolidated gleyed till surrounding this sinkhole suggests that this, and other sinkholes, may have served to locally under-drain a subglacial landscape into and through pre-Wisconsinan Cobleskill area caves before its final compaction. While not necessary, an elevated water table or potentiometric surface may have been present beneath glacial ice. According to conventional interpretation, water influx into the sinkhole discussed here, at about 300 m msl, may have, at times, brought sediment into a conduit back flooded to a lower elevational stage of glacial Lake Schoharie. However, an alternate explanation can also be considered to explain the physical conditions present during the formation and/or enlargement of the large and massively scalloped sinkhole discussed here. Because scallop formation in bedrock-walled sinkholes requires turbulent flow and development under non-saturated conditions, it is possible that this and other nearby sinkhole drains provided efficient outlets beneath glacier ice and through cave systems that were, at least at times, not back flooded by a glacial lake. Perhaps caves effectively under drained melting glaciers for long periods of time, prior to late glacial inundation by glacial Lake Schoharie. In this scenario, it is possible that much or all of the glacier ice that supplied water to scalloped sinkholes melted, leaving behind a basal till that covered and occluded numerous sinkholes. This melting ice may have contributed water to glacial Lake Schoharie that may then have saturated the till above this sinkhole. In this scenario, the gleyed till may reflect inundation by Lake Schoharie and not an elevated potentiometric surface within glacial ice.

DESCRIPTION OF FIELD-TRIP STOPS

For locations of stops, please refer to Figure 1, and to the road log following the stop descriptions.

Cobleskill Plateau

The Cobleskill Plateau contains some of the largest caves in the Northeast. This low, broad plateau consists mainly of limestone of the Helderberg Group, overlain in places by younger strata and by glacial till. The beds dip an average of 1.5 degrees to the south-southwest toward the Cobleskill Valley, which defines the southern boundary of the plateau (Figure 3). It is bounded on the east by the Schoharie Valley and to the north by the locally faint Helderberg Escarpment. To the southwest the limestones dip beneath base level and become covered with progressively thicker insoluble rocks, which prevent karst development. Northwestward the karst persists in subdued form along the Helderberg Escarpment.

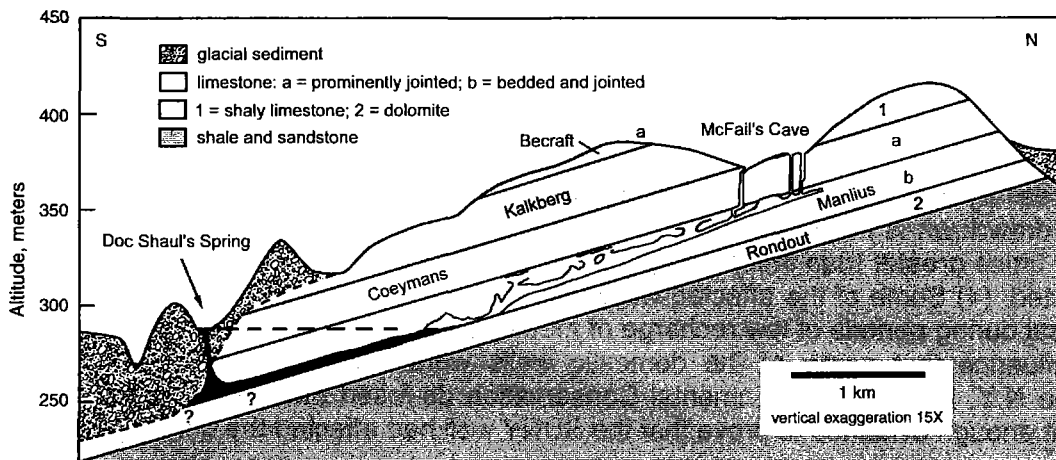


Figure 3: Geologic profile through the Cobleskill Plateau, showing the relationship of the Cobleskill valley and glacial deposits to the pattern of underground drainage.

The limestones of the plateau were completely truncated by erosion in the down-dip direction by Cobleskill Creek, but glacial-alluvial valley fill at least 100 ft (30 m) thick has covered their eroded surfaces in many areas. Because of the valley fill, the present stream course does not follow the exact path of its preglacial bedrock gorge, and in places it has established an entirely new path across exposed bedrock. Where it crosses the Coeymans Limestone near Barnerville, much of the water (all of it in dry periods) flows underground for several hundred meters.

The Helderberg Group contains a more continuous sequence of karstifiable limestones here than it does in most other areas of New York in which it is exposed. To the east the shaly and relatively impermeable New Scotland Formation occupies most of the upper Helderberg Group. In the Cobleskill Plateau, however, the purer Kalkberg Formation thickens westward at the expense of the New Scotland, so some karst groundwater is able to penetrate the entire sequence from Becraft to Rondout. Limestones of the Onondaga Group play a relatively minor role in karst development in the Cobleskill Plateau. In the western part of the plateau the Onondaga is exposed only in steep slopes and contains only a few small caves and springs.

STOP 1 – WELL FIELD ON CAMPUS OF SUNY COBLESKILL

This stop gives a chance to stretch and to review the geology of the region. The zero point in the field-trip log is at the nearby traffic light on NY Route 7. Note Cobleskill Creek (a western tributary of Schoharie Creek) and the broad plateaus rising to the north and south. To the northeast is the Cobleskill Plateau, which consists mainly of limestones of the Helderberg Group, and which contains the largest caves and karst drainage systems in the Northeast. Preglacial entrenchment of Cobleskill Creek allowed tributary subsurface drainage to develop in the plateau, and so the creek is directly responsible for much of the karst in the region.

In July 1991, SUNY Cobleskill began pumping from one of several wells at this stop to support an aquaculture program. The well penetrates several meters of glacial-alluvial sediments and extends into the underlying Onondaga Limestone. It was pumped intermittently

at about 30 gpm. Soon afterward many domestic wells in the vicinity ran dry or lost capacity. The SUNY well was widely held responsible. It seems to have an unusually broad effect, because wells lost capacity all over the county. Some ran dry even in Oneonta.

The blame was clearly misplaced. There are several reasons why the SUNY well could not have been responsible: (1) the volume of water pumped was many orders of magnitude less than that necessary to lower the water table over such a broad area. (2) In a cone of depression the greatest drawdown is in the center, but drops in water level in many of the affected wells were several times greater than in the pumping well itself. (3) Wells closest to the pumping well were unaffected. (4) Some of the affected wells also experienced an influx of hydrogen sulfide, which is typical during periods of low recharge of meteoric water. (5) The SUNY well was such a malignant influence that some of the domestic wells went dry even before SUNY began pumping. The N.Y. Dept. of Environmental Conservation de-fused the issue by declaring that there was not enough evidence to prove that the SUNY well had affected the domestic wells.

STOP 2 – DOC SHAUL’S SPRING

Many karst springs are located in the limestones along the southeastern end of the Cobleskill Plateau, near Howe Caverns (Figure 1). Farther west, however, the limestones exposed by pre-glacial erosion were covered by glacial and alluvial sediment, to depths as great as 30 m. Doc Shaul’s Spring, the main outlet for water in the western part of the plateau, rises upward from conduits in the limestone through a conical pit in the overlying sediment (Figure 3 and 4). It is one of the largest springs in the state. Much of the water that feeds it can be seen in caves higher in the plateau.

The original spring appears to have issued directly from the exposed down-dip edge of the limestone. The depth of the original spring has not yet been determined, but it is likely to be in the lower Manlius Limestone, about 25 m below the present spring. Divers have found the opening nearly choked with logs and sediment. Did the valley sediment accumulate slowly enough that it was continually swept away by the upwelling water, so that the spring remained open? Or was the spring inactive during aggradation and reactivated only by water forcing its way upward through the sediment? Saturated sediment has an effective specific gravity of about 1.1-1.2 (accounting for buoyancy and porosity), so a pressure head of about 30 m would have been necessary to balance the weight of the sediment. Dissolution of travertine in the tributary caves extends to levels at least 10 m higher than this. Thus either interpretation is feasible.



Figure 4: Doc Shaul's Spring, at the southern edge of the Cobleskill Plateau, rises from the limestones through glacial and alluvial sediment.

STOP 3 – BROWN'S DEPRESSION

Because of the difficulty of parking and land access, this "stop" actually consists of short views from the bus at several locations, accompanied by geologic discussions. Most of the depression is on private property where visitors are not welcome.

Brown's Depression (Figure 5) looks like an enormous sinkhole, but it is actually a remnant of a preglacial valley, which elsewhere has been completely obscured by glacial till. A second-order stream sinks into the limestone at the western edge of the depression. Either this part of the valley was never filled by glacial sediment or it was later exhumed by sapping through underground conduits. Laminated clays in the bottom of the depression at an altitude of 1100 ft (335 m) are probably deposits from glacial Lake Schoharie (Myroie, 1977).

Surface drainage was deranged by glaciation in many parts of the Cobleskill Plateau. Prior to glaciation, a prominent north-south stream valley extended through the western part of the plateau, reaching downward through almost the entire Helderberg Group in places. Deflection of glacial ice by the Helderberg Escarpment imposed a local west-southwest movement of glacial ice. As this direction was nearly perpendicular to the valley, the valley was almost completely filled with glacial till. As a final flourish, the ice camouflaged the valley with transverse drumlins that stand well above the surrounding terrain. Surface streams now follow a circuitous route around the drumlins, losing themselves here and there in wetlands.

The buried valley was detected with gravity surveys and well logs by Palmer (1976) and Milunich and Palmer (1997), and with reflection seismology by Myroie (1977). Its average depth of fill is 60-70 m. The generalized gravity profiles are shown in Figure 6. The deep part of the valley lies beneath the eastern flank of Brown's Depression, rather than directly beneath it.

The pre-glacial valley, of which Brown's Depression is part, separates two of the largest caves of the Northeast, McFail's Cave to the east and Barrack Zourie Cave to the west (Figures 6 and 7). Both discharge into Doc Shaul's Spring through water-filled passages. They probably join each other before reaching the spring, but whether the passages are open phreatic tubes or are occluded by glacial sediment is uncertain. Both caves contain extensive calcite-rich, fine-grained rhythmites that are convincingly demonstrated to have been deposited during the ponding of glacial Lake Schoharie (Palmer, 1976; Dumont, 1995).

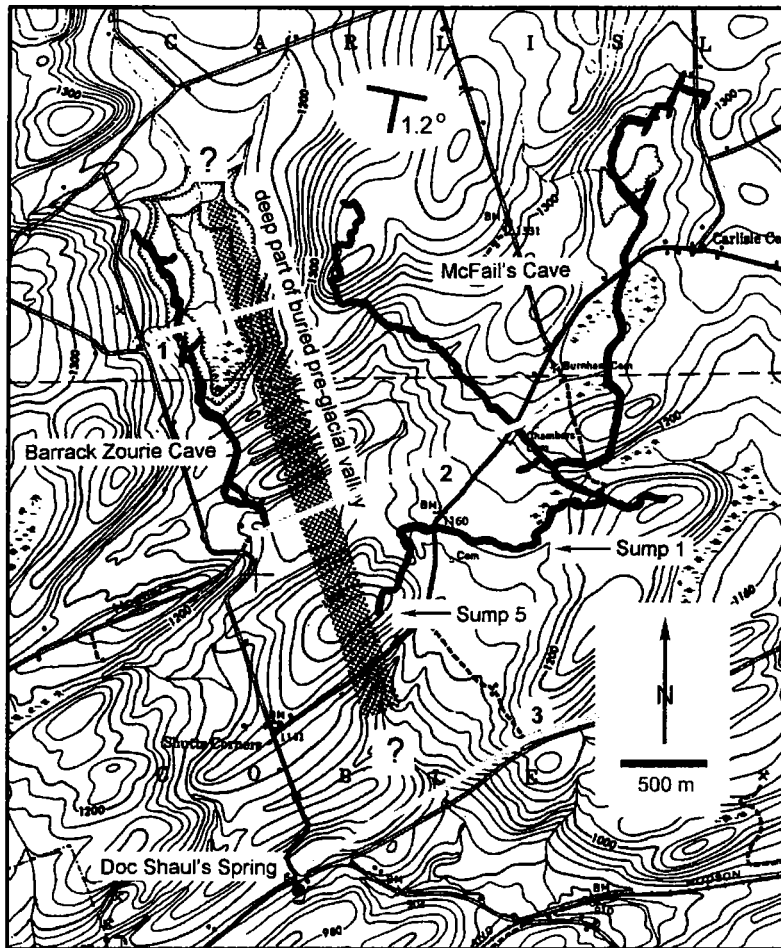


Figure 5: Relationship of the buried pre-glacial valley to Barrack Zourie and McFail's Caves. Only the deep part of the bedrock gorge is shown. Contour interval = 20 feet (6 m). Numbered gray lines refer to gravity traverses shown in Figure 6. Map of Barrack Zourie Cave from Dumont (1995). Map of McFail's Cave by A. Palmer, M. Palmer, E. Kastning, R. Zimmerman, K. Nichols, and others.

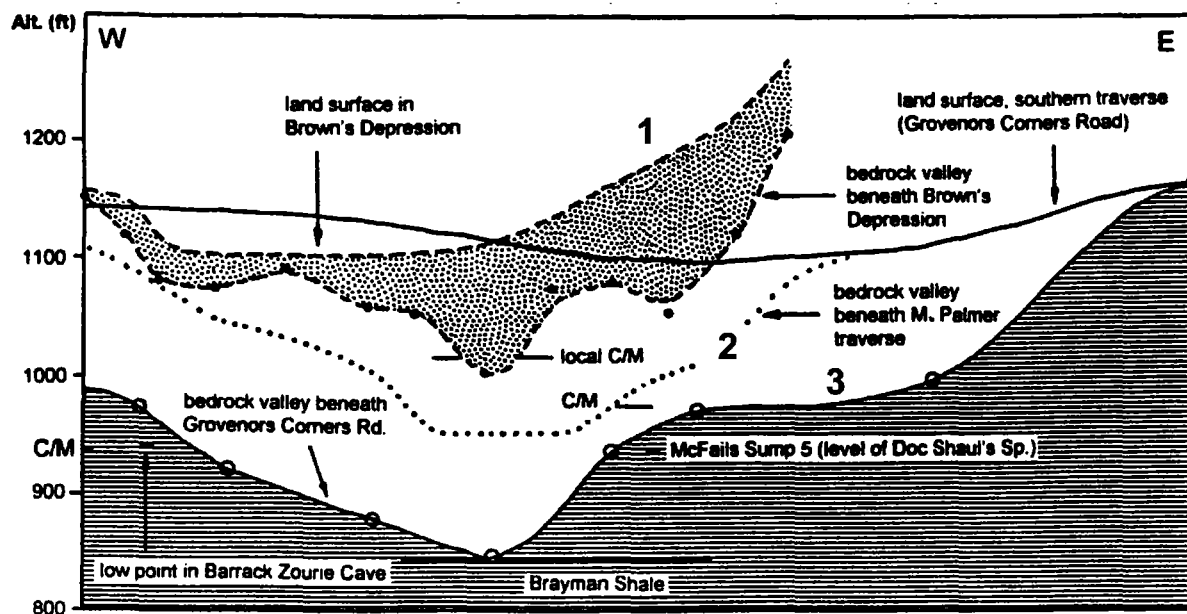


Figure 6: Cross sections of the buried bedrock valley at Brown's Depression, determined by gravity surveys. Elevations of nearby cave passages and local bedrock contacts are shown for comparison. C/M = Coeymans/Manlius contact. Dark line is the top of the Brayman Formation. No vertical exaggeration. Numbers refer to traverses shown in Figure 5. (From Milunich and Palmer, 1997.)

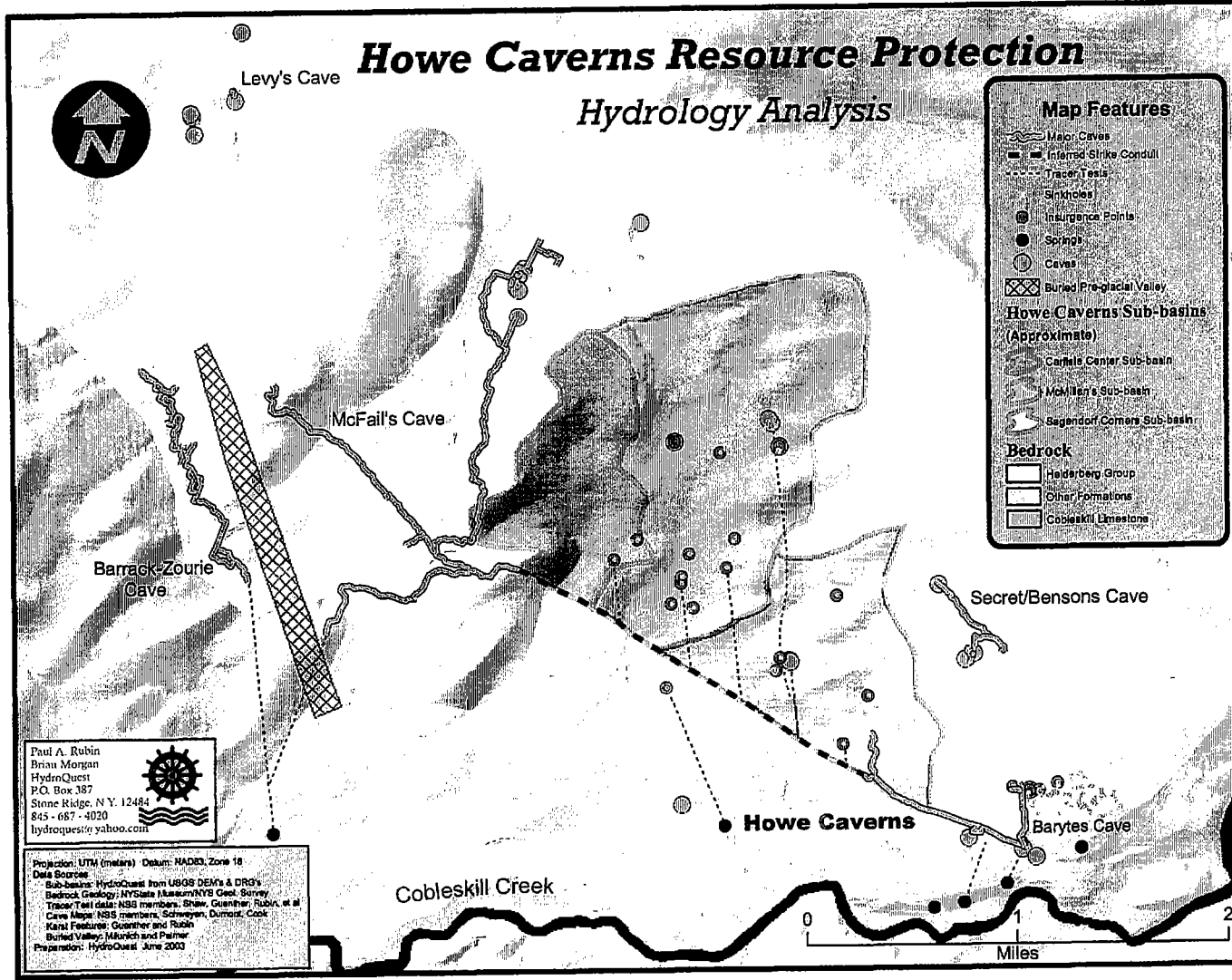


Figure 7. Howe Caverns Resource Protection hydrology analysis. Source: HydroQuest. Funding provided by Howe Caverns, Inc.

STOP 4 -- MCFAIL'S CAVE ENTRANCES

This property is owned and managed by the National Speleological Society, and a waiver is required to visit it. The cave entrances are all gated, and access is limited only to groups with the proper experience and equipment.

McFail's Cave has a mapped length of 10.7 km (6.6 mi.). Most of its water is contributed by two independent recharge areas. The main passage, which consists mostly of a single long down-dip canyon, is fed by numerous shafts and sinkholes in the vicinity of this stop. Its largest tributary, the Northwest Passage, is a remnant of a lengthy strike-oriented passage that formed at a former base level (Figures 5 and 7). Its strike orientation shows that water was prohibited from continuing down-dip to the Cobleskill Valley by the lack of limestone exposure in that direction. Dye tracing shows that the Northwest Passage is fed by drainage to the northeast of Brown's Depression, and that the water that sinks in the depression does not enter the Northwest Passage, but flows through Barrack Zourie Cave to Doc Shaul's Spring (Mylroie, 1977; Dumont, 1995).

The original flow was to the southeast to Howe Caverns, forming a passage graded to an outlet level at about 275 m elevation (Figure 7). In McFail's Cave the passage is at an elevation of 305-312 m. An alternate explanation is that the passage was formed or modified by glacial meltwater at a time when ice blocked the Cobleskill Valley, and does not represent a true erosional base level. The passage is partly relict today and has been segmented by breakdown and fill. The lower part of McFail's Cave is entirely water filled (Figure 3). Divers have penetrated to a depth of more than 6 m in the final sump (Sump 5 on Figure 5).

As Cobleskill Creek cut headward into the limestones south of the plateau, a new flow route developed to Doc Shaul's Spring, and the strike passage joining McFail's and Howe was abandoned (Palmer, 1976; Mylroie, 1977). Howe Caverns is simply a beheaded remnant of this passage. The accessible passage segments still contain perennial streams, but most are underfit because of the loss of drainage area. Small scallops indicate past discharges much greater than at present (Palmer, 1976). This may indicate modification by glacial meltwater (Rubin, 1991). Rubin (1999) used the Chezy-Manning formula and physical measurements to reconstruct Howe Caverns flows for 1938 and 1996 floods, using a range of Manning's n values between 0.03 and 0.04. This provided a range of peak flows between 270 and 365 cfs for a 1996 flood and between 1135 and 1520 cfs for the flood of 1938. Infrequent floods of this magnitude may still not have been sufficient to account for the conduit size present.

The cave entrances lie in a little woodlot that contains the densest cluster of vertical shafts in the Northeast (Figure 8). Each shaft feeds a tributary of McFail's Cave. Most of the tributaries are not traversable. A maintained trail snakes through the property past some of the more significant karst features. Please do not cross the open field, as this is private cultivated land.



Figure 8: Sinking stream at one of the entrances to McFail's Cave System, at the peak of a 50-year flood (January, 1996).

The tops of the entrance shafts are located in the thin-bedded Kalkberg Limestone. The Coeymans Limestone accounts for the vertical-walled sections of the shafts. The deepest shafts reach about 3-6 m into the Manlius Limestone. Much of the main passage originated at or near the Coeymans/Manlius contact and has since been entrenched downward as much as 12 m as a narrow sinuous canyon. In places its ceiling rises along prominent joints to heights up to 20 m. At the junction with the strike-oriented phreatic passage (see Figure 5) the cave stream has still cut only half way downward through the Manlius. Farther downstream the main passage descends through the rest of the Manlius and about a meter into the Rondout.

The main stream passage (Figure 9) extends almost exactly down the local dip, deviating significantly from this trend only where joint control is prominent. This orientation implies gravitational flow – hence vadose cave origin. Despite the prominent jointing in the limestone, the cave stream, as well as the original cave-forming water, was deflected laterally more than 1.5 km before reaching the potentiometric surface. From the standpoint of groundwater contamination, it is important to realize that infiltrating water in limestone, or probably any other bedded rock, rarely takes a perfectly vertical path downward to the potentiometric surface.



Figure 9: Down-dip vadose canyon in McFail's Cave.

The density of sinkholes and shafts in this small area requires some explanation. Limestones of the field-trip area are overlain in most places by relatively impermeable sedimentary rocks and by glacial till, which concentrate runoff into the few places where the limestone is exposed. Here, for example, is a narrow embayment in the low-permeability overlying rocks, where the relatively pure limestones are exposed beneath a thin soil cover (Figure 7). Runoff converges on this area from the high areas all around, so it is not surprising that so many karst features are concentrated here. The cave contains many small tributaries in its upstream end, fed by the openings seen at this stop. Where the cave extends beneath the cap of relatively insoluble rocks the only tributaries are mere trickles that deposit travertine.

Joint control of passages is strong in the area around the entrances, but farther down dip, where the cave is overlain by as much as 100 m of bedrock, joint control is much less prominent and bedding-plane control dominates. Minor bedding-plane thrusting may have aided the apparent lack of joint control. It appears that the abundance of fissures in the entrance area is due in part to erosional unloading. Glacial stresses may have played some part, although it must be kept in mind that the cave is known to predate at least the Wisconsinan glaciation.

Despite the great amount of surface runoff, the water entering the cave is not particularly aggressive. In fact, during most of the year the cave water is slightly supersaturated with calcite

and is unable to enlarge the cave by solution. Only during periods of high runoff, when sinking streams are fed by large amounts of overland flow, is the cave water solutionally aggressive.

Another small cave is located in the Becraft Limestone almost directly above McFail's Cave, which lies 60 m below. Water from the higher cave drains into one of the entrances of McFail's and thus traverses two different caves in entirely different limestone formations.

STOP 5 – HOWE CAVERNS

Howe Caverns is the largest Northeastern cave open to the public (Figure 10). The original entrance lies in the southeastern corner of the Howe Cave Quarry (Stop 6), whose operation has obliterated the connection with the main part of the cave. Howe Caverns is now entered through a 45 m elevator that descends from the visitors' center at the top of the plateau. From the elevator, the cave tour follows the main passage downstream for 500 m to a short boat ride on a dammed lake. Near the elevator, the main passage is joined by a down-dip tributary canyon, the Winding Way. The tributary water now follows a lower route, leaving the Winding Way dry during periods of low and moderate flow. However, during periods of high flow, the downstream outlet is inefficient. In response, floodwaters initially backup in the up gradient portion of the Winding Way, then aggressively flow through the Winding Way to the strike-oriented portion of Howe Caverns. It has not yet been determined whether this floodwater overflow route carries only water from the eastern portion of the Sagendorf Corners sub-basin or if it also serves as a bypass shunt around breakdown for floodwaters from upstream portions of the Inferred Strike Conduit (Figure 7). Upstream from the elevator the main passage can be followed a short distance to breakdown at the northwestern terminus of the West Passage, situated southeast of a small valley tributary to Cobleskill Creek. This passage appears to have once been the downstream continuation of the strike-oriented passage in McFail's Cave, whose terminus now lies about 3 km to the northwest (Figure 7). Tracer test results confirm that groundwater efficiently flows beneath this small valley prior to rising as the River Styx in Howe Caverns.

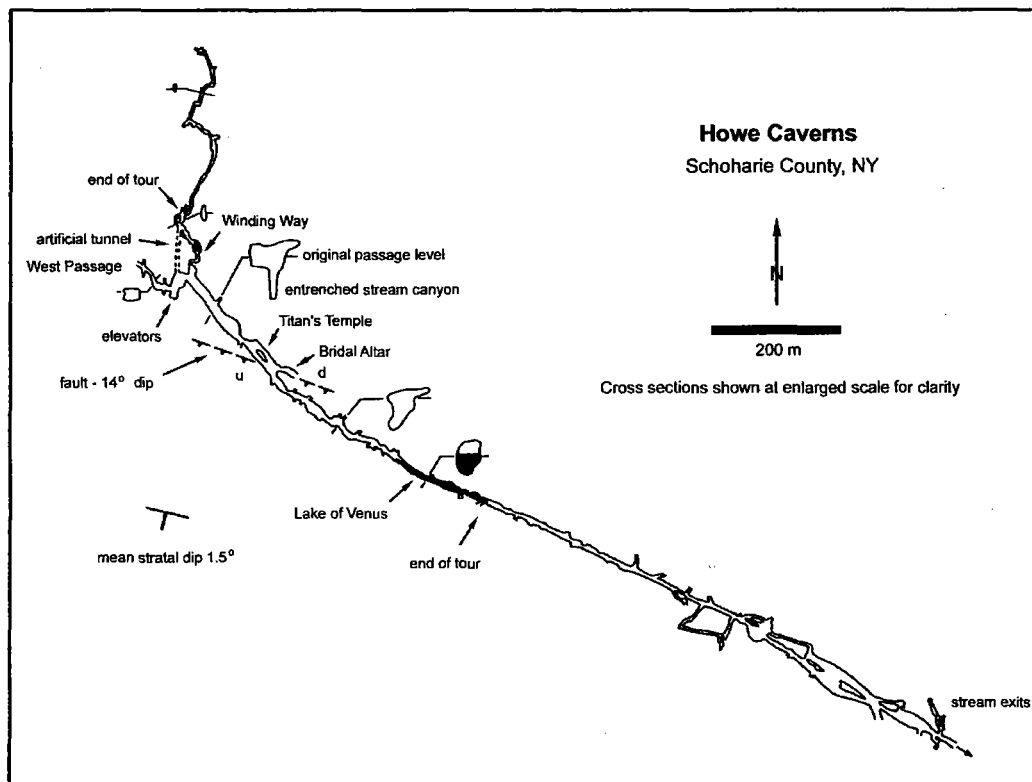


Figure 10: Map of Howe Caverns (from Mylroie, 1977).

Howe Caverns, now isolated from McFail's, receives its present drainage from those parts of the plateau updip to the north.

The main passage of Howe Caverns is a large tube up to 10 m wide and about 6 m high, with a 3-5 m canyon cut in its floor (Figure 11). The solutional ceiling is approximately at the Coeymans/Manlius contact. At Titan's Temple, the largest room in the cave, the lower-level canyon diverges from the tube. The stream, of course, follows the lower level, which assumes a more tube-like configuration farther downstream. The upper level is clay-choked in the former downstream direction. At the divergence point between the two levels, the cave intersects a subtly exposed reverse fault, also seen in the Howe Cave Quarry (Stop 6). Here the fault dips 14 degrees to the south-southwest. The fault can be seen in the chin of the bedrock feature known as the "Old Witch" (Gregg, 1974). It can also be seen extending downstream of the Old Witch along the southwestern passage wall.

The significance of the reverse fault (and smaller subsidiary ones) to the local cave development has been debated. Egemeier (1969) and Mylroie (1977) attribute a great deal of the local cave-passage orientation to the fault. Gregg (1974) and Kastning (1975) downplay the influence of the fault. The general consensus is that stratigraphy and base level exerted the main control over the Northwest-Southeast Passage in McFail's and the main passage of Howe, and that the fault controls local passage segments.

The main passage contains some noteworthy deposits. Most unusual are large remnant banks of thinly laminated clay rhythmites. These are clearly ponded-water deposits, and their abundance here and in other caves of the Schoharie Valley (but rarely, if ever, in other New York State caves) suggests that they accumulated in the late Wisconsin glacial Lake Schoharie. As in McFail's, they occupy the lowest level of vadose entrenchment. Almost the entire solutional history of the cave pre-dates the latest retreat of glacial ice. Since their deposition, the clay banks have been largely eroded away by the cave stream. Remnants of a thin flowstone covering conform to the eroded shapes of the clay banks. Most other dripstone and flowstone in the cave appear to be older than the clay fill. Many stalactites and stalagmites have been resurrected from fallen pieces, not necessarily in their original locations.



Figure 11: Main passage of Howe Caverns, showing the original strike-oriented phreatic tube with the later vadose canyon cut in its floor.

The Winding Way is the largest of several tributary passages, and the only one that can be followed for any great distance. Like most of them, it enters the main passage from the up-dip side. Although most of the tributaries are vadose canyons, their solutional ceilings descend steeply across the strata to join concordantly with the ceiling of the main passage. This is odd behavior for vadose passages. Should gravitational water have any greater tendency to cut downward across the strata above the main passage than it would anywhere else? The angle and direction of the discordant ceilings are suspiciously parallel to the trend of the reverse fault, however; initial solution of the main passage may have opened imperceptible fractures parallel to the main fault that guided the vadose water.

The tours exit the cave through an artificial tunnel between the Winding Way and the elevators. Much of this tunnel is excavated through a fault zone marked by small slickensided exposures and related calcite mineralization.

Howe Caverns Karst Hydrology and Cave Resource Protection

Howe Caverns, the most frequently visited commercial cave in New York State, is currently reviewing all factors that may impact their underground natural resource (Rubin et al., 2003). *HydroQuest* recently completed the first phase of this work for Howe Caverns, Inc. Phase I of this study examined surface and subsurface hydrologic and geologic factors that control groundwater influx into the cave (Figure 7). This is important because existing and potential land-uses within the drainage basins of commercial (and other) caverns may pose a threat to air quality, groundwater quality, cave-dwelling species, and surface streams. Protection of the cave resource requires knowledge of the karst principles discussed in this paper, coupled with information on tracer test results, bedrock geology, glacial geology, topography, land-use, and karst features.

Three surface drainage basins tributary to Howe Caverns were delineated by analysis of DEM and DRG data, and field verification. Land-use within these basins was photo-identified from high resolution imagery and mapped. Karst features (e.g., caves, sinkholes, sinking streams, springs) near Howe Caverns were located with a Global Positioning System receiver and plotted. Tracer test results were also plotted. All this was brought together in two unified GIS map documents (Figures 7 and 12). The tracer test flow routes plotted assume groundwater flow in a down-dip direction, often independent of surface drainage basins. Additional tracer tests may find that sub-basin boundaries extend further up-dip within Helderberg Group carbonates to the north-northeast beyond surface watershed boundaries. The southwestern sub-basin boundaries depicted truncate nearly coincident with the inferred strike conduit, reflecting a strike-oriented flow path into Howe Caverns. Actual sub-basin boundaries may also extend a short distance southwest and up slope of the inferred strike conduit. The exact subsurface path of the inferred strike conduit represents a reasonable hydrologic interpretation based on tracer test, geologic, and hydrologic factors.

The dominant land-use throughout the three sub-basins tributary to Howe Caverns is agricultural (Figure 12). Three active dairy farms operate in the watershed, each with large hay fields and row crops. Small horse farms are also present. Remaining portions of the watershed are largely comprised of forest, low intensity residential development, and wetlands. The greatest potential threats to Howe Caverns are manure/septic odors and related water quality degradation. Land-uses that pose this threat are primarily agricultural and residential based. What best management practices should be considered within each drainage basin?

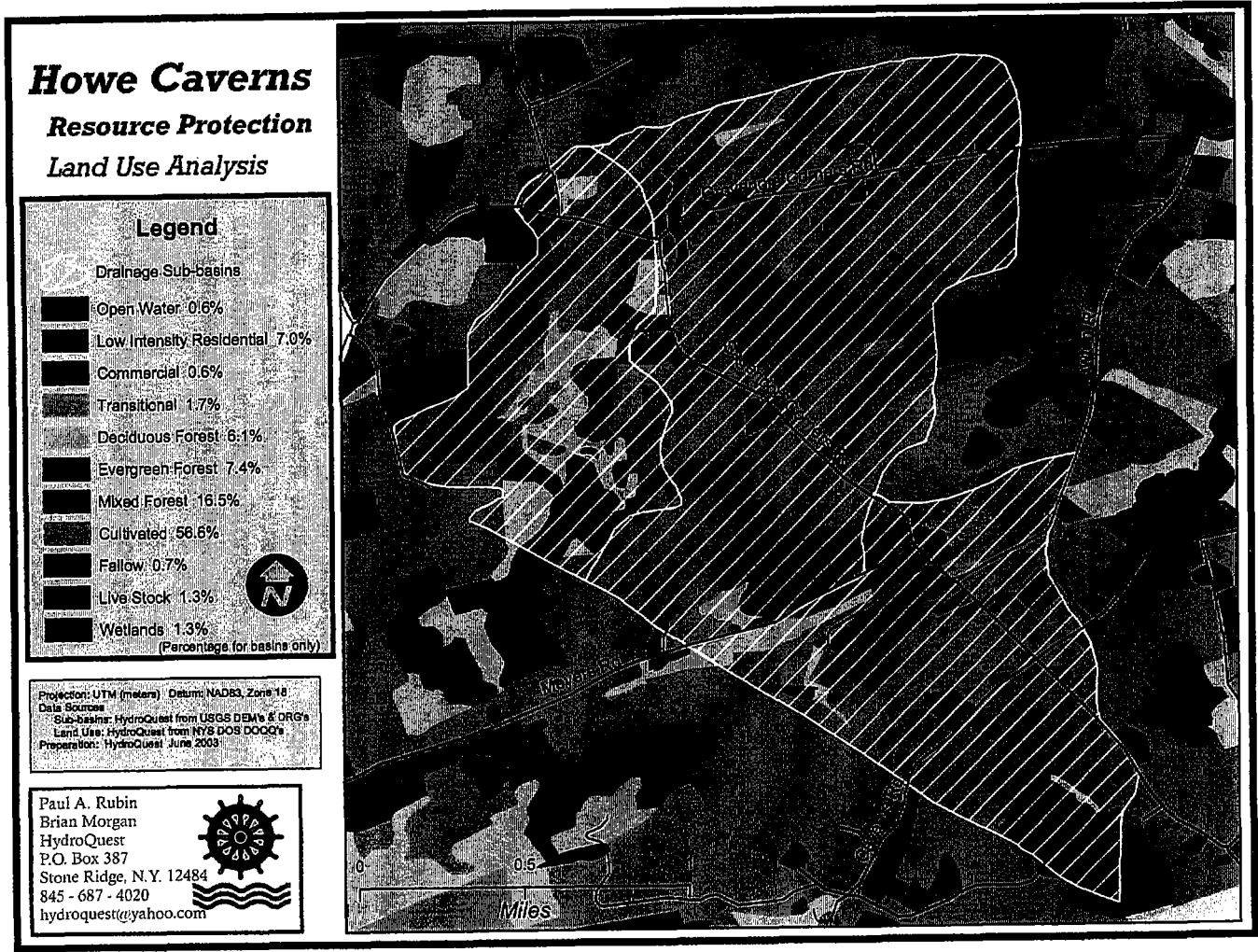


Figure 12. Howe Caverns Resource Protection land-use analysis. Source: HydroQuest. Funding provided by Howe Caverns, Inc.

STOP 6 – HOWE CAVE QUARRY

This quarry is now owned by Cobleskill Stone Products, which plans to develop part of the property into an educational center for use by school groups. The 150-year-old house on the property, once a hotel for visitors to the old Howe's Cave may become the visitors' center, with exhibits on mining and geology. The house was built next to the former natural entrance to the cave. However, quarry operations obstructed, partially filled, and removed portions of the roof of the downstream section of the cave. Cobleskill Stone Products is currently examining the feasibility of reopening this cave section for educational and commercial purposes.

The full thickness of the massive Coeymans Limestone and underlying thin-bedded Manlius Limestone is visible in the quarry walls, although their stratigraphic differences are masked in these artificial surfaces (Figure 13). A few meters of the Dayville Member of the Coeymans appears here. The highly jointed quarry floor is in the lower Manlius and is underlain by the Rondout Dolomite, which is extensively riddled with mine tunnels that extend below the quarry floor (Figure 14). The limestone was once quarried for crushed rock and the dolomite for cement.



Figure 13: Howe Cave Quarry is located in the Manlius and Coeymans Limestones. A small thrust fault is visible in the wall.

Note the diagonal fault trace in the quarry wall (Figure 13). This is the same reverse fault that is visible in Howe Caverns. Here it has about 0.45 m of displacement, and, as in the caverns, it dips 14° to the SSW. It is actually subsidiary to a much larger bedding-plane thrust exposed in the mine below the quarry floor. Strontianite and barite veins up to a meter thick occur along the lower fault zone.

The downstream end of Howe Caverns, beyond the limit of the tours, is contained within the narrow peninsula of limestone extending from the western side of the quarry. The truncated end of the natural cave is obscured by the debris cone at the base of the quarry wall. Water from the

cave drains through an artificial tunnel and shaft into the cement mine below. The resurgence of this water is located in the cliff below the quarry (Figure 14). The original drainage pattern of underground water has been disrupted considerably by the mining and quarrying.

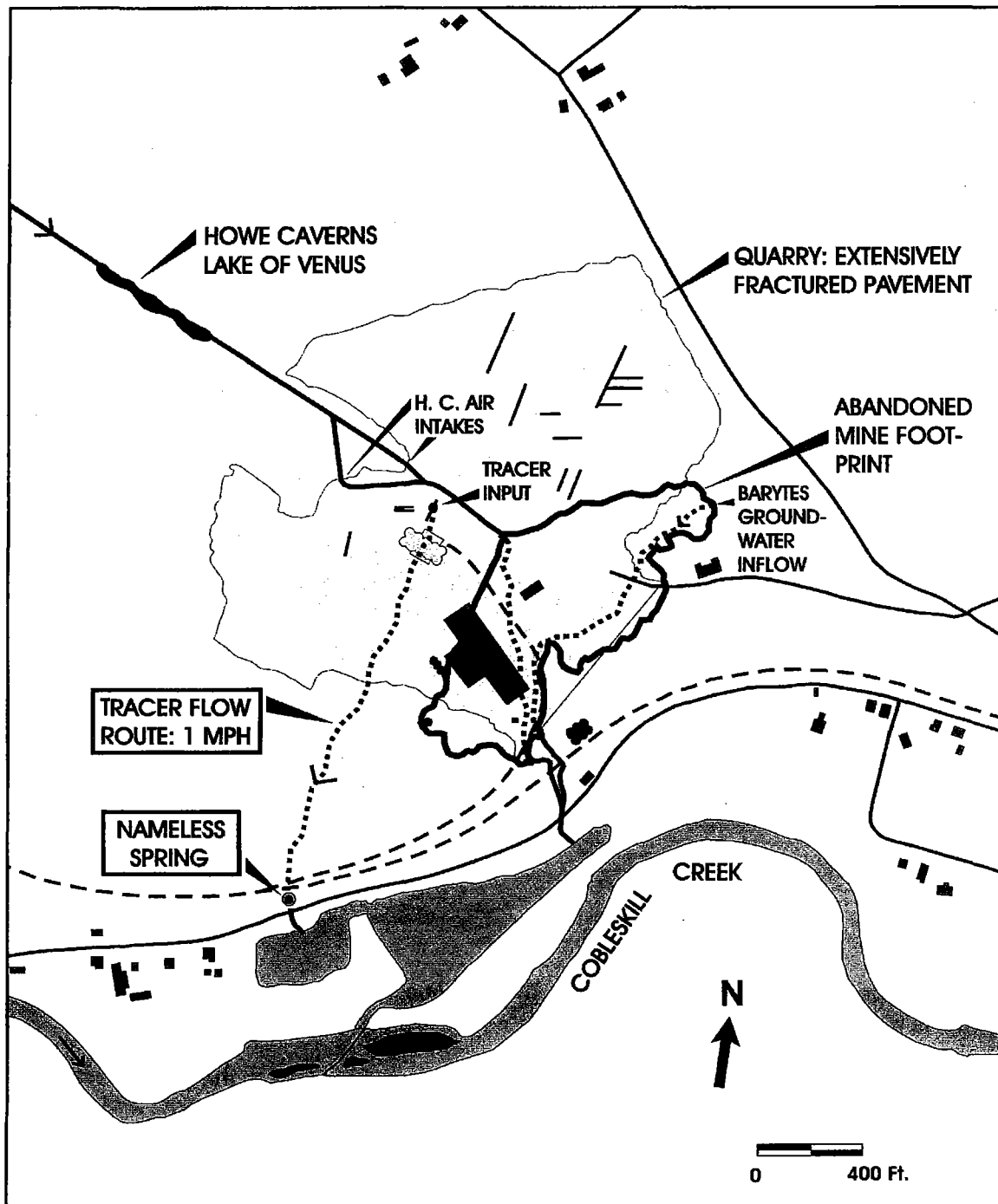


Figure 14: Hydrology of the Howes Cave Quarry area altered by historic mining operations. Source: Rubin and Guenther.

A large system of caves – the Secret-Benson-Barytes system – once joined Howe Caverns near its downstream terminus but has since been dismembered from the latter by the Howe Cave Quarry (Figures 7 and 14). Secret Caverns, which like Howe is open to the public, connects downstream with other caves in a discontinuous series that has been traced southward to a spring below the Howe Cave Quarry.

In the cliff face southwest of the quarry is a spring (Nameless Spring) at the base of the Cobleskill Formation, perched on the shaly Brayman. Guenther and Rubin have conducted tracer tests from an enlarged joint in the floor of the quarry north of the silo complex to Nameless Spring (Figure 14). Notable hydrocarbon residue and staining (perhaps diesel fuel) are visible at the tracer injection point. Travel time to Nameless Spring has been documented in as little as 23 minutes. This spring is a popular drinking source for the local community who assume that karst springs must be pure water. This conduit provides an example of stream piracy trending in a down-dip direction away from the strike-oriented Howe Caverns conduit. During high flow, water also emerges from an ephemeral spring farther west at the Rondout-Cobleskill contact (Figure 7). The latter has a larger alcove and small cave formed by extensive dissolution and collapse, suggesting that this spring was active for a longer time, and that Nameless Spring is a rather recent diversion route for this water (Myroie, 1977).

Glacial striae are visible in the limestone surface at the edge of the quarry, indicating that the amount of denudation since the last glacial retreat has been negligible here. This surface was covered by soil and till before the quarry was excavated, and the high carbonate content of the overburden prevented solutionally aggressive water from attacking the bedrock surface. Compare the almost complete lack of post-glacial denudation here to the rates of 20-30 cm per thousand years in some karst areas (Sweeting, 1973).

STOP 7 – TERRACE MOUNTAIN

We descend from the limestone plateau and follow Cobleskill Creek Valley downstream (Figure 1). Our route then crosses Schoharie Creek and follows it upstream for a few kilometers. Terrace Mountain looms to the right, with the gentle dip of its exposed limestone beds clearly visible. These strata look superficially like the Manlius and Coeymans Limestones, which are displayed so clearly in many escarpments in the region. However, the Terrace Mountain cliffs are actually composed of the Cobleskill and Rondout Formations, which reach their maximum thickness in this area. The Manlius and Coeymans occupy the higher slopes.

To the north, Schoharie Creek runs through a narrow gap in the limestone plateaus. This gap was once the site of ice blockage during the waning phases of Wisconsinan glaciation, about 14,000 years BP, which flooded the valley to create glacial Lake Schoharie. Prominent faults cut through the valley here, as shown by massive pyrite and strontianite bodies in the bedrock at the base of Terrace Mountain, but the main faults are probably obscured by the great thickness of valley sediment.

Although Terrace Mountain might seem an ideal location for karst features, today it does not have a catchment area large enough to supply recharge to more than a few shafts, sinkholes, and narrow abrasive caves, all of them small. The largest cave is recently discovered Ain't No Catchment Cave, named to poke fun at those who said there was insufficient catchment area

on the mountain to form a significant cave. However, this low, wet crawlway simply proves the point. Its main virtue is that it is one of the very few caves developed in the Cobleskill Formation.

We leave the Schoharie Valley and follow the valley of Fox Creek upstream to the east along the southern edge of Barton Hill.

Barton Hill

Barton Hill is large enough, compared to Terrace Mountain, to have developed extensive karst drainage and large caves (Figure 15). Some of the springs at the southern (down-dip) edge of the plateau are used as a water supply for the village of Schoharie, and with increasing development of Barton Hill many land-use management questions have arisen. Studies of the Barton Hill drainage pattern go back more than 40 years (Gurnee, 1961).

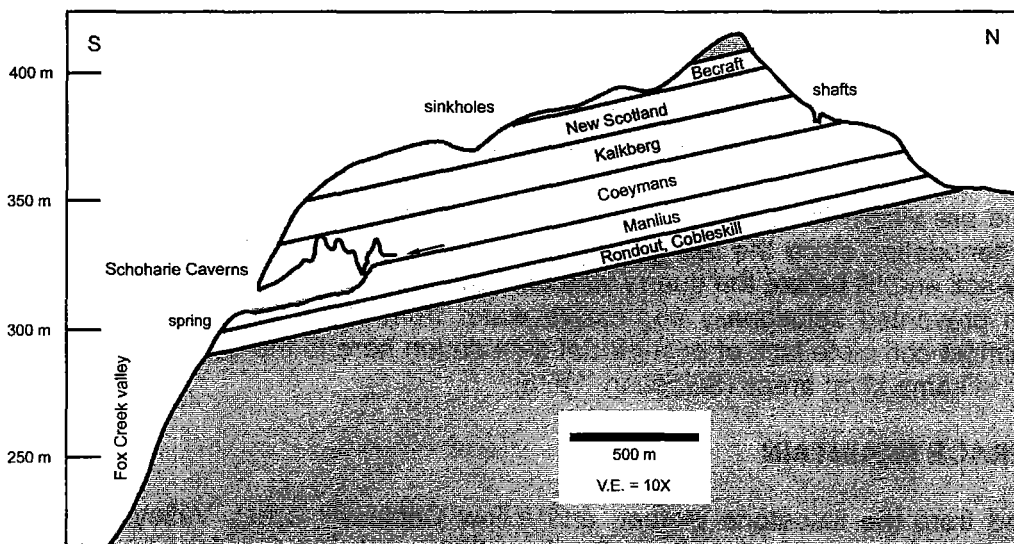


Figure 15: Geologic cross section through Barton Hill, showing the stratigraphic position of major caves.

The Helderberg limestones of Barton Hill are perched high above the nearby valleys, and the known caves have no apparent control by local fluvial base levels (Figure 15). As Fox Creek cut headward toward the east, the first opportunities for groundwater drainage to its valley were in the down-dip direction. The unrestricted down-dip drainage toward Fox Creek has prevented the development of phreatic tubes at large angles to the dip, at least as far as exploration has shown. In contrast, the limestones in the Cobleskill Valley were exposed only gradually by headward stream erosion as cave development progressed, allowing strike-oriented tubes to form before the erosion breached the limestones in the down-dip direction (Mylroie, 1977).

Glacial deposits have filled many sinkholes on the top of the plateau and have blocked some former spring outlets along the southern edge, but for the most part glacial disruption of the underground drainage pattern has been relatively minor. Virtually all of the major cave development is concentrated along the Coeymans/Manlius contact or in the upper half of the Manlius. Gage Caverns retains a large cross-sectional area upstream nearly to the northern margin of Barton Hill, suggesting that its catchment area must have been much larger when the

cave was forming. Proglacial or subglacial meltwaters could have helped to enlarge the cave independently of the normal topography.

STOP 8 – SCHOHARIE CAVERNS

This is the property of the National Speleological Society, and visitors must sign a liability waiver. Schoharie Caverns (or Shutters Corners Cave) consists almost entirely of a single large canyon passage that can be followed upstream about 600 m to a sump. The upstream 240 m of this section consists of a single joint-controlled fissure. The upstream sump and 5 others beyond have been dived (Schweyen, 1989). The traversable part of the cave terminates in two small branches containing infeasible streams. Flowstone is abundant in the cave, but it has been extensively redissolved by back-flooding and further damaged by mindless collectors. The spring opening was nearly blocked by glacial till, flooding the cave. The redissolving of the flowstone probably dates from that time. Beyond a number of sumps and upstream of a waterfall, a large stalagmite is positioned on the bedrock floor in the center of the stream passage (Schweyen, pers. comm.). This is unexpected because it could not have formed when the cave stream was flowing. Its presence suggests the cessation of water responsible for conduit formation, perhaps during an interstadial period, with long-term infiltration through the epikarst.

The former owner deepened the entrance by clearing away much of the till, still not reaching bedrock; so the cave became drained. Although subsidence of the entrenched till has begun to block the entrance again, a metal conduit installed in the bottom of the trench keeps the water level low, although during severe floods the Schoharie Caverns entrance fills nearly to the ceiling.

Uranium-series dating of speleothems from Schoharie Caverns (Lauritzen and Mylroie, 1996) gives dates greater than 350,000 years, the practical limit for the technique, suggesting that the origin of the cave is at least mid-Pleistocene. The semi-perched nature of the cave makes it difficult to relate this date to karst features elsewhere in the area or to fluvial events in the valleys below.

Solutional rills in the limestone face above and to the left of the entrance are up to several centimeters deep. The time required for this grooving was apparently not very long, although it is not certain exactly when the face was first exposed. It is almost certainly post-glacial.

Most or all of the water that feeds Schoharie Caverns has also passed through the soil, but the high degree of aeration in the cave causes calcite to precipitate inside as speleothems. The water still continues to lose carbon dioxide where it exits the cave, maintaining slight supersaturation, but not enough to cause detectable precipitation of calcite.

Another much smaller spring lies a few hundred meters to the southeast in the Rondout or Cobleskill Formation. It has deposited a large sheet of tufa on the hillside below the cave entrance. Evidently the water that feeds the cave has passed through soil rich in carbon dioxide and has reached equilibrium with dissolved limestone at these high CO₂ levels. The groundwater was not aerated through cave entrances, and so carbon dioxide rapidly degasses from the water only where it comes in contact with the low carbon-dioxide levels of the outside air. The water rapidly becomes highly supersaturated and deposits some of its dissolved load.

We continue eastward, leaving the Fox Creek Valley and climbing gradually onto the plateau in which Knox Cave is located.

Karst Systems of the Knox Area

The plateau in which Knox Cave is located is a broad, low-relief upland bounded on the north by the Helderberg Escarpment but with gradational boundaries in the other directions (Figure 16). The Helderberg Limestones form most of the surface, but here the New Scotland Formation in the middle part of the sequence is shalier than in the Cobleskill area. Much surface runoff collects on this shaly limestone and flows either south into tributaries of Fox Creek, or north into sinkholes in the Coeymans Limestone. The dip is about 2.5 degrees to the SSW, about twice as steep as at the other sites.

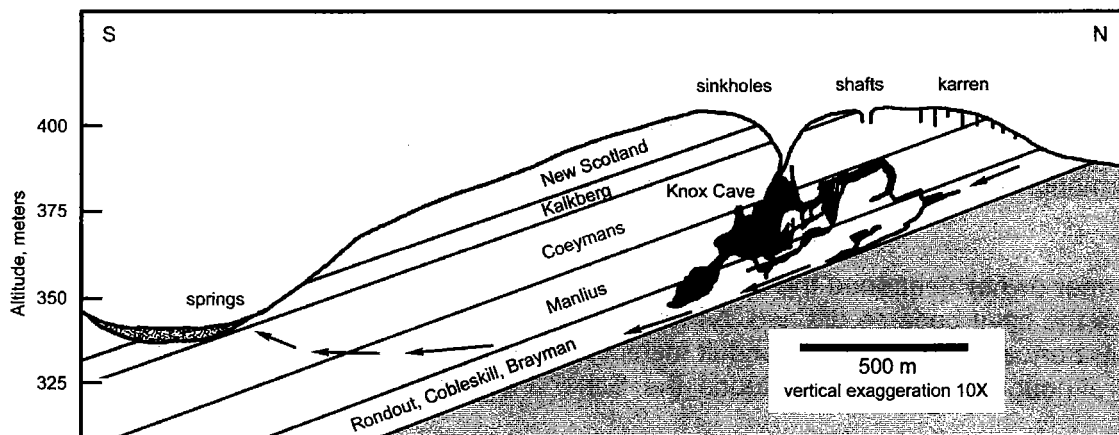


Figure 16: Geologic cross section through the plateau in which Knox Cave is located. For clarity, the cave is shown in black.

A rather limited recharge area feeds the cave-forming limestones along the northern border of the plateau, and they are not exposed in the down-dip direction. Instead, the springs are located in the Coeymans or Kalkberg Formations, which indicates that subsurface drainage must rise upward across the strata to the springs. Nevertheless, some of the largest caves in the state have formed here. Knox Cave, with 1.2 km of passage, is accessible only to persons through written permission of the Northeastern Cave Conservancy.

The springs that drain Knox Cave and other caves in the vicinity are located in a small tributary of Fox Creek (Hesler, 1984). Because they are partly blocked by thin glacial till and other sediments, and by massive collapse in places (as at the Knox Cave entrance), subsurface drainage is rather inefficient, particularly during floods, accounting for many features in these caves attributed to severe flooding, such as long fissures that intersect in a network pattern.

To the west is a large drainage system that feeds several springs on small northerly tributaries of Fox Creek. Water draining into several deep sinkholes has been dye traced to the springs over distances of several kilometers. Dye travels this distance in less than 16 hours during all but the lowest flow conditions. On the basis of measurements in 1984, Rubin (1986) reports that the largest of these springs (Bogardus Spring) had the third highest low-flow discharge in the field-trip area (3.2 liter/sec), in comparison to 6.3 liters/sec for Doc Shaul's Spring and 5.2 liters/sec for No Admittance Spring below Howe Caverns. This underground system appears to represent one of the largest undiscovered cave systems in the state.

STOP 9 – LIMESTONE RISE

Limestone Rise, owned by The Nature Conservancy, displays good examples of a “karst pavement” consisting of limestone surfaces with solutionally enlarged joints. The Coeymans and Manlius Limestones are represented here. The soil is very thin and absent in many places. Such fissures constitute one form of what is known as the epikarst – the uppermost zone of karst in which many paths of infiltration have been enlarged by dissolution (Figure 17). Exposed bedrock such as this is typical only of glaciated regions, which suggests either that the soil has been partly stripped off by glacial action, or that fissure enlargement has been enhanced by glaciation. Loading and unloading by glacial ice could have widened the joints and made them more susceptible to dissolution, perhaps aided by glacial meltwater, allowing soil to subside into the underlying fissures.



Figure 17: Solutionally enlarged joints in the Coeymans Limestone at Limestone Rise.

STOP 10 – ENTRANCE SINKHOLE OF KNOX CAVE

Knox Cave is owned and managed by the Northeastern Cave Conservancy, and visitors must sign a liability waiver. It is a complex of solutional fissures surrounding joint-controlled tubes and canyons (Figure 18). In places the passages coalesce into rooms up to 20 m high. The entrance sinkhole lies approximately in the middle of the cave and nearly blocks access to the northern half.

The entrance sinkhole extends through the upper 2/3 of the massive Coeymans Limestone and is rimmed by a slope of thin-bedded Kalkberg Limestone (Figure 19). From the pattern of underlying cave passages, it appears that much of the sinkhole's origin is due to collapse of bedrock between narrow fissures. The cave is located mainly in the Manlius Limestone, and the Coeymans/Manlius contact appears half way down the entrance canyon. Deeper, very inaccessible passages in the cave extend through 2.5 m of Rondout, about 30 cm of Cobleskill, and 5 m into the Brayman Formation, which is locally a dolomitic shale.

The cave was open to the public for several decades prior to 1961, and a few remnants of the old staircase into the entrance passage are still visible (Figure 19). The cave is basically a crude branchwork system of tubes and canyons strongly aligned in the NNE-SSW direction of the major joints (Figure 18). It originated as a southward-trending tube that was later blocked by collapse and sediment. At a later time, a canyon passage formed a bypass around the blocked tube. About half the cave's length consists of fissure-like passages along joints, probably formed by diversion and injection of floodwater resulting from blockage of the main passages by sinkhole collapse and sediment fill. Multi-level and discordant passage intersections support a floodwater origin for the fissures. Note the complexity and strong joint control of the cave, as shown in Figure 18, compared to the dendritic patterns of other caves in the regions (Figures 5 and 10).

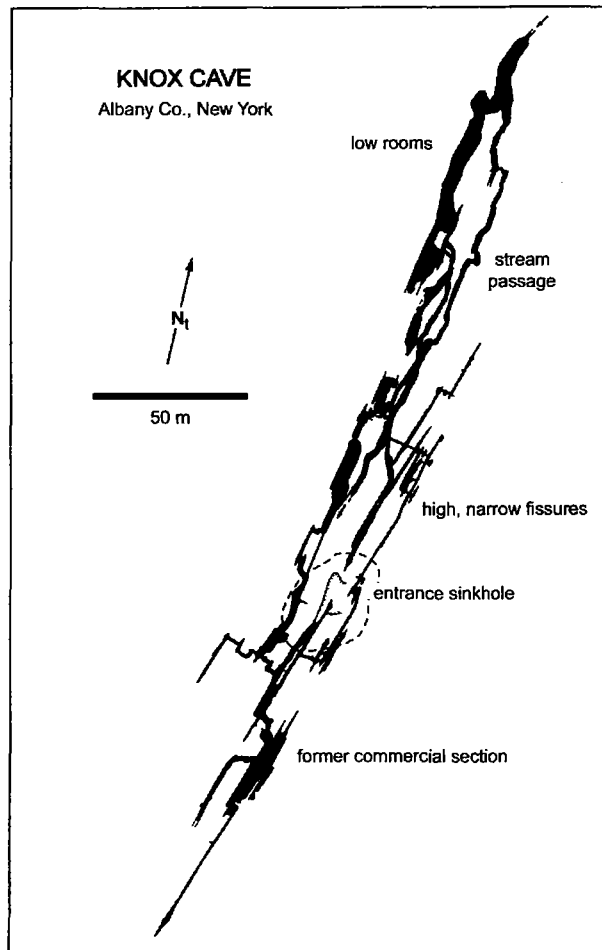


Figure 18: Map of Knox Cave, showing the relation of the cave passages to the entrance sinkhole. Mapped by A. Palmer and M. Palmer.

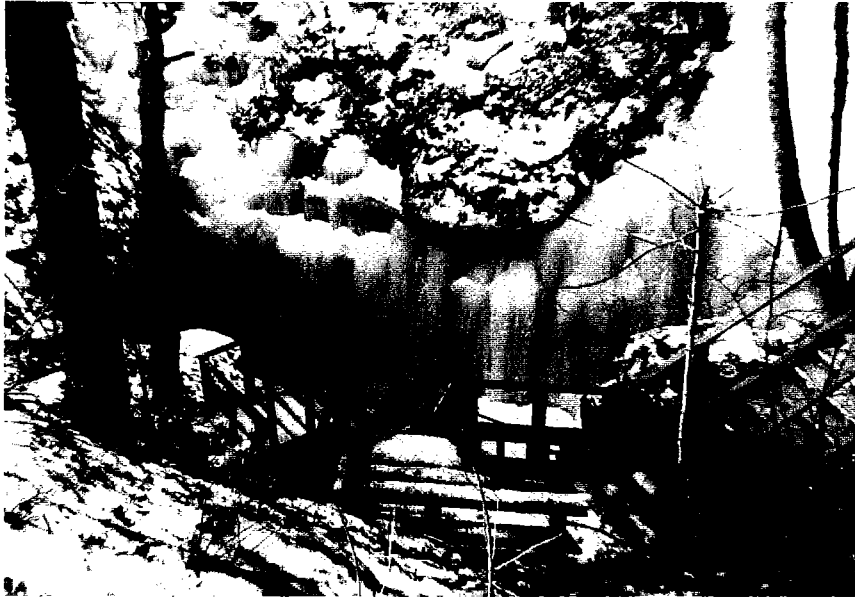


Figure 19: Sinkhole entrance of Knox Cave in 1959, when the cave was still open to the public. The upper slopes are in the Kalkberg, and the vertical-walled section is in the Coeymans.



Figure 20: Interior of Knox Cave, showing prominent joints and joint-defined breakdown blocks. This room is at the top of the Manlius Limestone.

STOP 11 – SPRINGS SOUTH OF KNOX CAVE

South of Knox Cave is a small park maintained by the Town of Knox. This area contains many springs fed by drainage from the caves in the plateau (Figure 16). Most are hidden by brush and are not readily visible. The springs rise from the underlying Coeymans Limestone through thin glacial, alluvial, and lacustrine deposits.

The nature of the karst drainage is cryptic. The overburden appears to average only a few meters thick, hardly enough to block large subsurface drainage paths, as in the valley of Cobleskill Creek at Doc Shaul's Spring (Stop 2). Yet, during high flow, water spurts upward in jets several decimeters high from minor fissures in the bedrock and soil, as though many subsurface conduits were semi-confined. Dye traces and geophysical field work are in progress in an attempt to evaluate the subsurface plumbing.

A short distance down the road from the parking lot is a wetland walkway that is ideal for bird-watching. It is hard to imagine a more pleasant and peaceful little park.

The field trip now follows Rte. 443 and I-88 back to Oneonta.

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ROAD LOG

This road log may deviate slightly from the actual field-trip route because of limited maneuverability of the bus.

Miles – Cumulative / (since last stop)

- 0.0 (0.0) Traffic light at main entrance of SUNY Cobleskill. **STOP 1** is just northwest of the intersection, west of the parking lot.
Follow NY Rte. 7 east through Cobleskill.
- 1.9 (1.9) Turn left onto County Rte. 7 just before railroad overpass (next to shopping center).
- 2.2 (0.3) Quarry on left is in Onondaga Formation.
- 3.2 (1.0) **STOP 2** is at junction with road to left. **Doc Shaul's Spring** is located below the junction in the trees on the right. Turn up the hill to the left onto the plateau, which consists mainly of limestones of the Helderberg Group.

- 3.5 (0.3) The top of the plateau has been streamlined by glacial ice, which was moving nearly westward in this local area. Deranged drainage, elongate bedrock hills, and drumlins are common.
- 3.8 (0.3) Continue straight at intersection.
- 4.2 (0.4) Road makes a wild curve around the nose of a prominent drumlin. The drumlins in this area are superimposed across a north-south pre-glacial valley in the Helderberg Limestones.
- 5.2 (1.0) **Brown's Depression (STOP 3)** is visible on the right. This is a remnant of the pre-glacial valley, now largely filled by glacial till. A second-order stream flows into the depression and sinks into the limestone along its western flank. The water emerges at Doc Shaul's Spring.
Continue straight ahead at junction.
- 5.6 (0.4) Another view of Brown's Depression on the right.
- 5.9 (0.3) Turn right onto Snyder Road.
- 6.2 (0.3) Cross the valley draining into Brown's Depression.
- 6.3 (0.1) Continue straight past junction with Crapser Road on the left.
- 7.3 (1.0) Turn right at junction onto unmarked road.
- 7.4 (0.1) **Barrack Zourie**, a prominent hill consisting of an outlier of Middle Devonian strata, can be seen to the right.
- 8.2 (0.8) Pass shale pit in the **Esopus Formation** on right.
- 8.6 (0.4) Turn left at T intersection onto Governors Corners Road (unmarked).
- 9.3 (0.7) Turn left onto Lykers Road.
- 9.8 (0.5) **McFail's Cave (STOP 4)**. Pull off road into a grassy parking lot on left. The various entrances to McFail's Cave can be visited by following the path through the woods at the northeastern end of the parking lot. The cave entrances and parts of the surrounding woods are owned by the National Speleological Society. Permission to visit the property can be obtained from the chairperson of the McFail's Cave Committee. Please do not cross the field, as it is private farmland.
When you leave the parking lot, turn right onto the paved road.
- 10.3 (0.5) turn left onto Governors Corners Road.
- 10.6 (0.3) Turn right onto Sagendorf Corners Road. Pass Myers Rd. on the right, then Lawton Road on the left.
- 12.4 (1.8) Four-way intersection. OPTIONAL: Turn right and go 2.9 mi to access road to Howe Caverns on left (**STOP 5**). After visiting the cave, return to the 4-way junction and resume road log from that point.
- 13.7 (1.3) **Howe Cave Quarry (STOP 6)**. This is private property and is accessible only for purposes of this field trip. Good exposure of **Manlius and Coeymans Limestones**. Note low-angle thrust fault on northwest wall. The down-stream end of the **Howe Caverns** is located in the rock peninsula projecting from the western wall of the quarry.
Drive or walk a short distance down Howe Cave Road, which branches to the right, to springs at base of plateau. This water is the drainage from Howe Caverns and neighboring cave systems.
Continue on main road across railroad tracks and past junction with other roads.
- 14.3 (0.6) Cross Cobleskill Creek again, then turn left onto Rte. 7 at traffic light.

- 14.8 (0.5) Cross **Schoharie Creek**. Deep entrenchment into the Helderberg Limestones by this north-flowing stream and its tributaries has been responsible for most of the karst development in Schoharie County.
- 15.6 (0.8) Turn right onto Rte. 30A.
- 16.6 (1.0) Rte 30 joins Rte. 30A from the left. **Terrace Mountain (STOP 7)** is on the right. It shows fine exposures of the gently dipping carbonate rocks. Although these look like the typical Manlius and Coeymans, in this exposure they are actually the Rondout and Cobleskill Formations.
- 18.0 (1.4) Turn left onto Rte. 443, following the Fox Creek Valley upstream. The steep bluffs of **Barton Hill** rise on the left.
- 19.2 (1.2) Turn left at 4-way junction onto a narrow road and drive up steep hill.
- 19.9 (0.7) **Schoharie Caverns (STOP 8)**. The property is owned by the National Speleological Society (2813 Cave Ave., Huntsville, AL 35810-4431).
- 20.6 (0.7) Drive back down the hill and turn left onto Rte. 443.
- 20.9 (0.3) On the left is a bank of semi-consolidated glacial till cemented by calcite, which is apparently deposited by emerging karst groundwater.
- 21.8 (0.9) Turn left onto Rte. 146 in Gallupville.
- 22.3 (0.5) Continue straight through intersection on Rte. 146.
- 23.1 (0.8) View of the **Helderberg Escarpment** immediately to our right, but here it has unimpressively low relief. There are good examples of glacially deranged drainage on the left.
- 25.5 (2.4) Albany County line.
- 26.5 (1.0) Pass Beebe Road on right. Continue straight on Rte. 146.
- 27.4 (0.9) **Limestone Rise (STOP 9)**. Park in the small parking lot on the left of the road, and follow the trail, which crosses the road and climbs to the top of the limestone plateau. Fine examples of limestone pavement and fissured epikarst.
- 27.5 (0.1) Continue east on Rte. 146 and turn right at 4-way intersection onto Knox Cave Rd.
- 28.2 (0.7) **Knox Cave (Stop 10)**. Park at the turnoff on the right. The cave and property are owned and managed by the Northeastern Cave Conservancy (Box 10, Schoharie, NY 12157).
Continue straight on Knox Cave Road.
- 29.1 (0.9) Turn right on Street Road.
- 29.8 (0.7) Park owned by the Town of Knox (**Stop 11**).

To return to I-88, continue on Street Road to T intersection and turn right onto Knox – Gallupville Road. This leads directly to Rte. 443. Follow 443 straight ahead (west) to Rte. 30. Turn right on 30, then left on 30A to I-88.