INTERACTION BETWEEN KARST AND GLACIATION IN THE HELDERBERG PLATEAU, SCHOHARIE AND ALBANY COUNTIES, NEW YORK

ARTHUR N. PALMER
State University College
Oneonta, NY 13820-4015

PAUL A. RUBIN
Environmental Sciences Division
Oak Ridge National Laboratory
Oak Ridge, TN 37831-6037

MARGARET V. PALMER
R. D. 4, Box 82
Oneonta, NY 13820

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. The purpose of this field trip is to examine some of the major karst features of the region and to show how they have interacted with Wisconsinan glaciation.

Karst features, particularly solutional caves, preserve the record of past groundwater flow patterns and geologic events very clearly -- much more so than surface topography, whose relict features tend to be easily destroyed or buried. Deposits in caves can also be dated radiometrically or paleomagnetically. Paleontological studies are beginning to show the colonization history of certain species, some of which are now extict. Karst features may therefore aid in unraveling the complex glacial history of the area. Conversely, a knowledge of the glacial history helps to explain certain unusual aspects of the Helderberg karst. Several open-ended questions will be posed on the field trip, and participants are encouraged to contribute to the discussion.

Four sites will be visited on this field trip: from west to east these are (1) the Cobleskill Plateau, (2) Barton Hill, (3) the Knox area, and (4) the Clarksville area. Although in general they share a common geomorphic setting, each has unique characteristics. Further details about the Helderberg karst are provided by Kastning (1975), Baker (1976), Palmer (1976), Mylroie (1977), Mylroie and Palmer (1977), Palmer et al. (1991), and Rubin (1991a, 1991b).

The area includes two major karst-forming rock sequences: the Helderberg Group (Upper Silurian - Lower Devonian), and the Onondaga Group (Middle Devonian). A generalized stratigraphic column is shown in Figure 2. Each contains about 30-40 m of cavernous limestone and dolomite. These two aquifers are separated by about 40 m of shale and siltstone, and so they are hydrologically independent. Throughout most of the field-trip area the

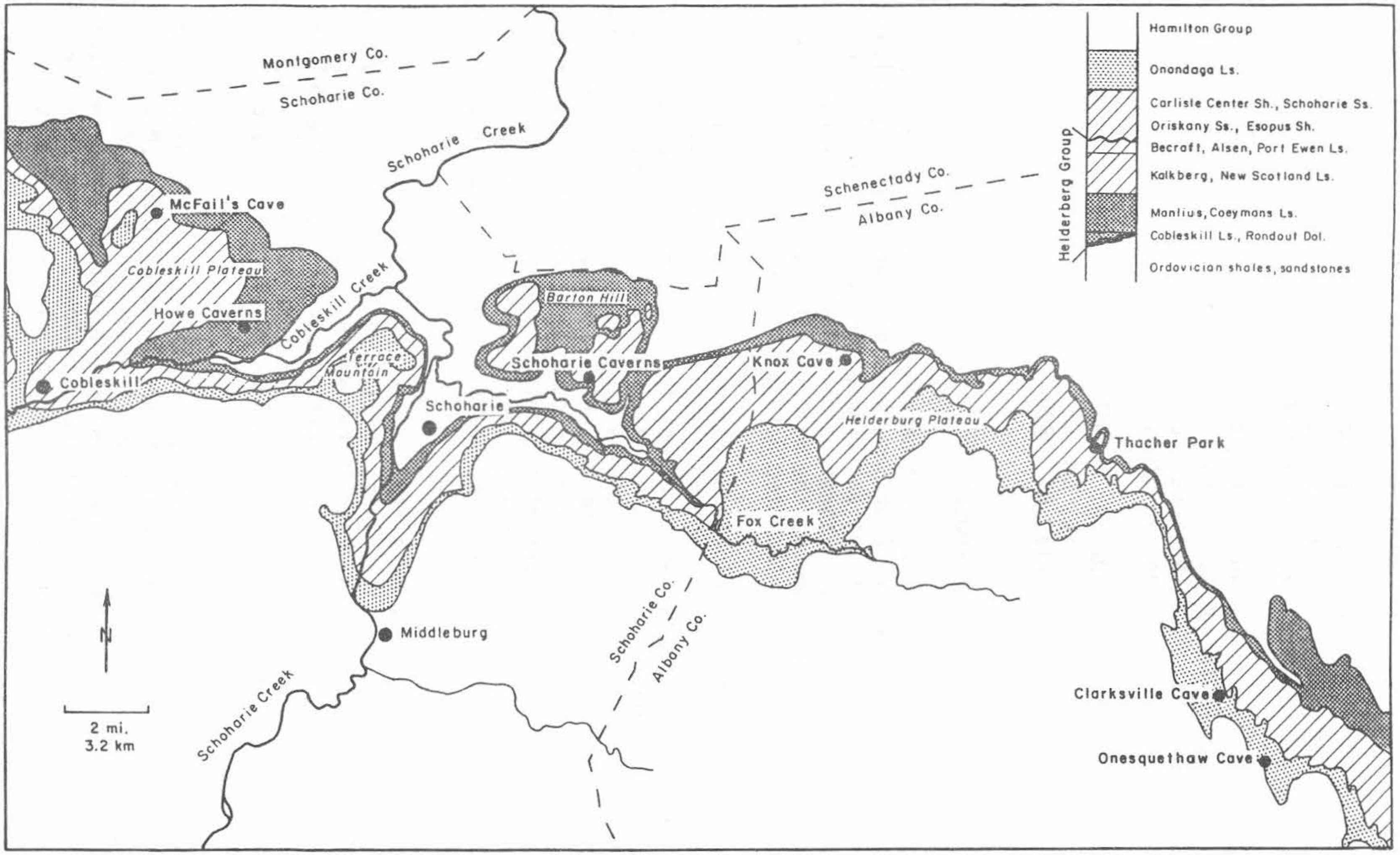

Figure 1: Geologic map of the field-trip area, showing location of stops (from Palmer, et al., 1991).

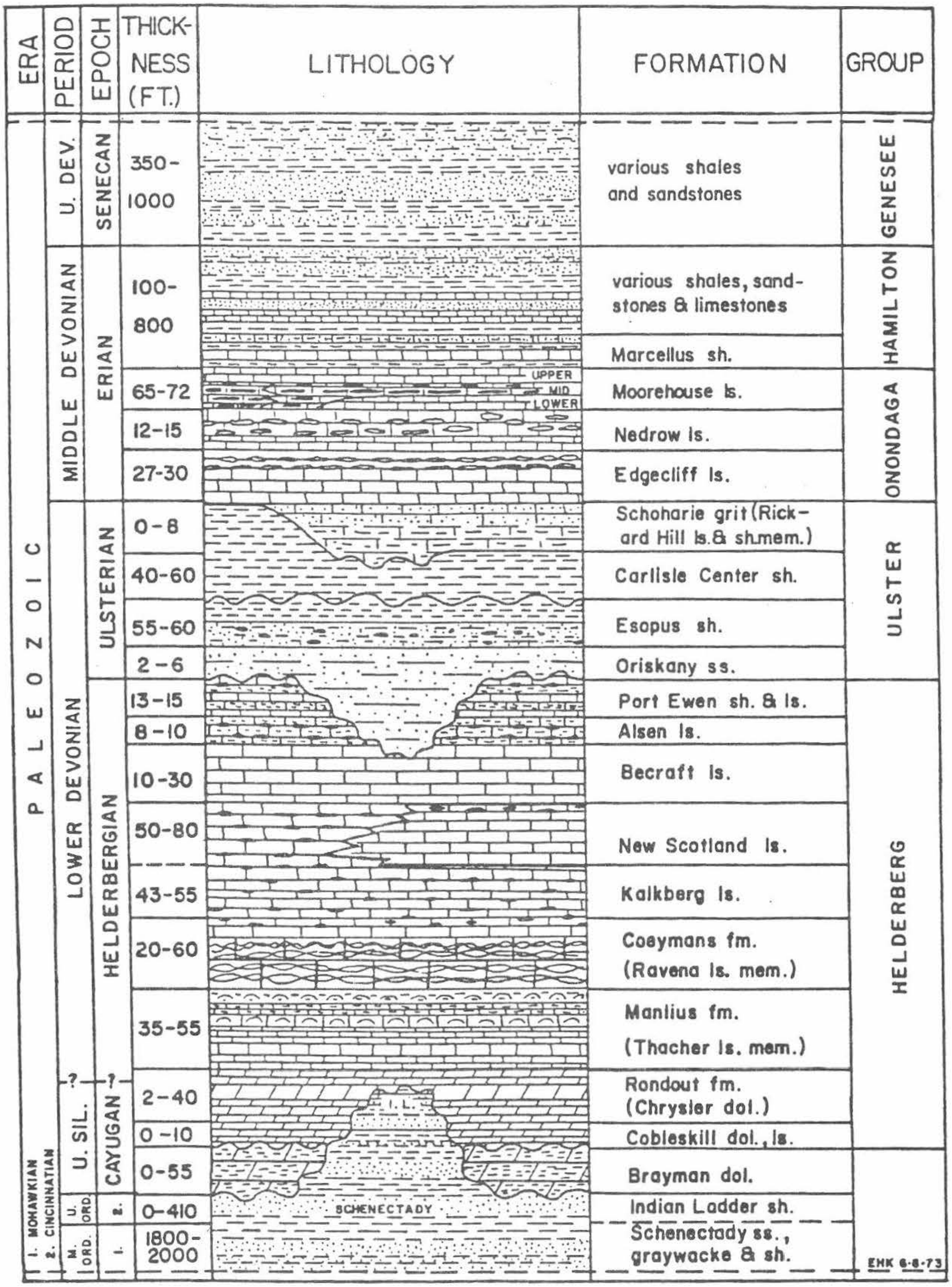

Figure 2: Stratigraphic section in the field-trip area (Kastning, 1975).

rocks dip less than 2 degrees to the south-southwest, but toward the east they are folded and faulted with a roughly north-south structural trend as the result of Appalachian tectonism.

GEOMORPHIC CONCEPTS

Origin of Karst Landscapes

Surface karst features, such as sinkholes, sinking streams, and large springs, owe their existence to the development of underground solution conduits (caves). When limestone is exposed in relief in a humid climate, groundwater selectively enlarges interconnected fractures, partings, and pores by solution, 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, the way groundwater typically does toward a surface valley. Some conduits serve as diversion routes for perched surface streams and may eventually 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 caves, see White (1988), Ford and Williams (1989), and Palmer (1991).

Cave Morphology

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. They are high and narrow, with sinuous bends controlled mainly by structural irregularities. 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 few 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 or low-angle faults are prominent a braided (anastomotic) pattern of intersecting tubes is formed around the constriction.

Rates of Karst Development

When the bedrock is first exposed to circulating groundwater, the flow within any single opening is minuscule and becomes nearly saturated after only a few meters. Average solution rates are extremely low. With time, a few select flow paths enlarge enough that water is able to retain much of its solutional aggressiveness over the entire distance. A significant geomorphic threshold is crossed. From that point on, the conduit diameters enlarge very rapidly, about 0.01-0.1 cm/yr under ideal conditions. They grow to cave size (i.e., large enough for humans to traverse) in a few thousand or tens of thousands of years. Surrounding groundwater tends to be drawn toward the lower heads that prevail in these few conduits. Feeder sinkholes develop rapidly, funneling progressively more water into these select conduits and robbing the lesser openings of much of their flow. Large conduits grow to form caves while the smaller ones languish with low and generally diminishing solution rates. Most caves therefore consist of a few discrete conduits, rather than a sponge-like array of pockets.

The time required for caves to reach the solutional threshold depends strongly on the initial fissure width and discharge, and less strongly upon hydraulic gradient, flow distance, carbon dioxide partial pressure, and temperature Palmer (1991). The initial stages of cave develpment typically persist for 10^4 - 10^5 years (Figure 3). This probably represents at least half of the evolutionary history of a typical cave and of its related karst topography. In caves fed by aggressive floodwaters, however, bypass routes around constrictions have such steep hydraulic gradients and short lengths that rapid solution can be achieved right from the start along many alternate paths, which results in the diversion mazes described earlier.

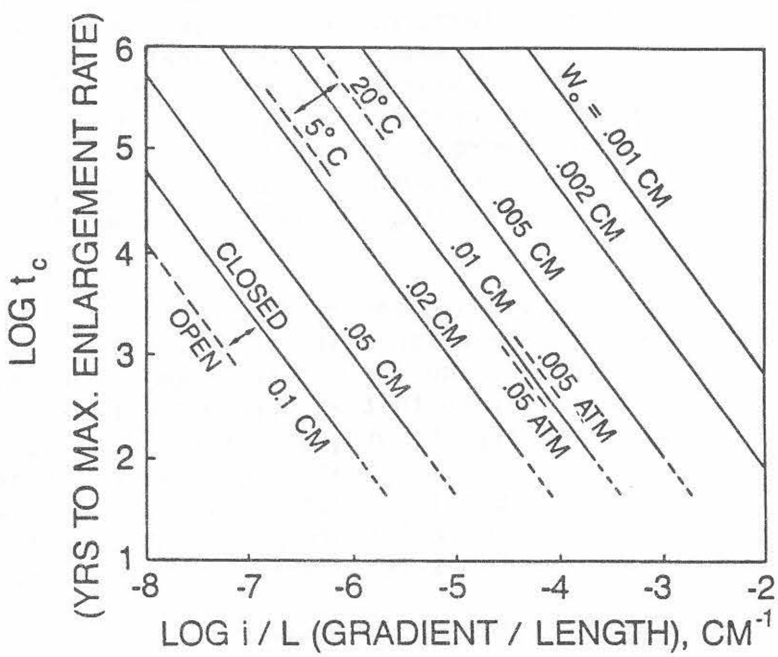

Figure 3: Time (t_{max}) required for a fissure in limestone to reach its maximum rate of enlargement (Palmer, 1991). L = length of fissure (cm), i = hydraulic gradient (dimensionless), W_0 = initial fissure width. The effects of difference in temperature, carbon dioxide partial pressure, and open vs. closed system with respect to carbon dioxide are shown.

Pleistocene Glaciation in New York

The history and effects of Pleistocene glaciation in the field-trip area are well documented elsewhere (e.g. LaFleur, 1969; DeSimone and LaFleur, 1985; Dineen, 1986, 1987), so only a brief summary is given here. The rich variety of glacial landforms and deposits in New York appear to date mainly from the Wisconsinan glaciation. Virtually all traces of earlier glacial events have been effaced or buried. Rare multiple tills in other areas (unpublished data from LaFleur, reported by Rubin, 1991a) suggest earlier ice advances, but their history is uncertain.

The Wisconsinan ice reached its maximum extent about 22,000 years b.p., with a thickness of about 1.5 km (Dineen, personal communication, reported by Rubin, 1991a). 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 1800 to 750 ft (550-230 m). Clays deposited in the lake are found in caves and on the surface throughout much of the lower Schoharie basin. At the surface these are found at elevations as high as 1100 ft (335 m), and in caves they range from 867 to 1130 ft (264-344 m). Along Fox Creek these clays were once used as a source of the bricks seen in many of the local houses.

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 subglacial and proglacial meltwater. There are fine examples of each in the field-trip area, but several questions remain. For example, what is the exact history of glacial and karst events? Did groundwater flow and limestone solution stagnate beneath the ice sheets when they were at their maximum extent? Are speleothems (cave deposits such as calcite) only of interglacial age?

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 times 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 cyles 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.

DESCRIPTION OF FIELD-TRIP STOPS

Most of the stops described here are on private property, and permission for access must be obtained from the owners or managers indicated. With few exceptions, caves in the area are not open to the public.

Please do not collect rock or mineral samples during the field trip. The instructional and scientific value of a site can be diminished greatly by aimless collecting. Specimens obtained without a clear research design are out of context and usually end up scattered and lost.

Cobleskill Plateau

Stop 1 - View of the Cobleskill Plateau

The Cobleskill Plateau (Figures 4 and 5) contains the largest karst drainage systems in the Northeast, of which the tourist caves Howe Caverns and Secret Caverns are part. A general view of the plateau is seen at Stop 1. The low, broad plateau consists mainly of limestones 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 SSW toward the entrenched valley of Cobleskill Creek, into which most of the karst systems drain. The bedrock floor of the valley is buried beneath as much as 30 m of late Quaternary glacial/alluvial sediment. The creek now follows a route slightly different from its deep-stage pattern, and in places it has migrated laterally onto exposed bedrock that once flanked the valley. Where the creek crosses the Coeymans Limestone (Helderberg Group), much or all of its flow is lost to underground solution conduits and emerges nearly a kilometer downstream. The present course of the creek is postglacial, so the diversion conduits have apparently developed within the past 14,000-15,000 years. The conduits are well adjusted to the present flow pattern, so it is doubtful that they are reactivated older features.

Stop 2 - Doc Shaul's Spring

Many karst springs are located in the exposed limestones at the eastern end of the plateau around Howe Caverns. Farther west, however, the erosionally truncated limestones have been covered by valley sediment. Doc Shaul's Spring, the main outlet for water in the western part of the plateau, rises upward from the limestone subcrop through a conical pit in the overlying sediment (Figure 5).

The original spring appears to have issued directly from the exposed down-dip edge of the limestone prior to aggradation. The depth to the original spring level has not yet been determined. Divers have found the opening to be nearly choked with logs and sediment. Did the valley sediment accumulate slowly enough that it was continually swept away by the up-

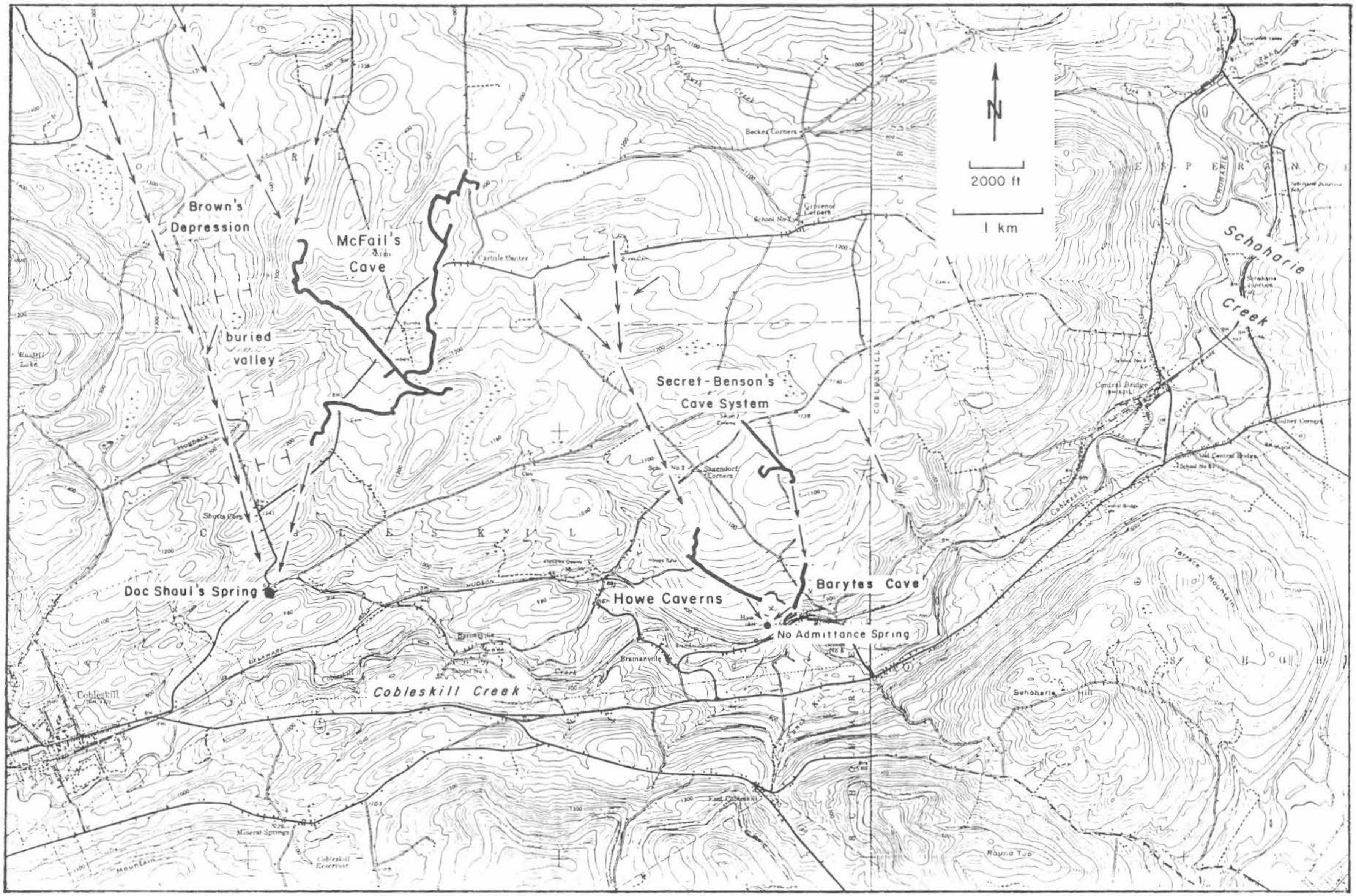

Figure 4: Topographic map of the Cobleskill Plateau. Arrows = dye traces (Mylroie, 1977). Dark lines = caves.

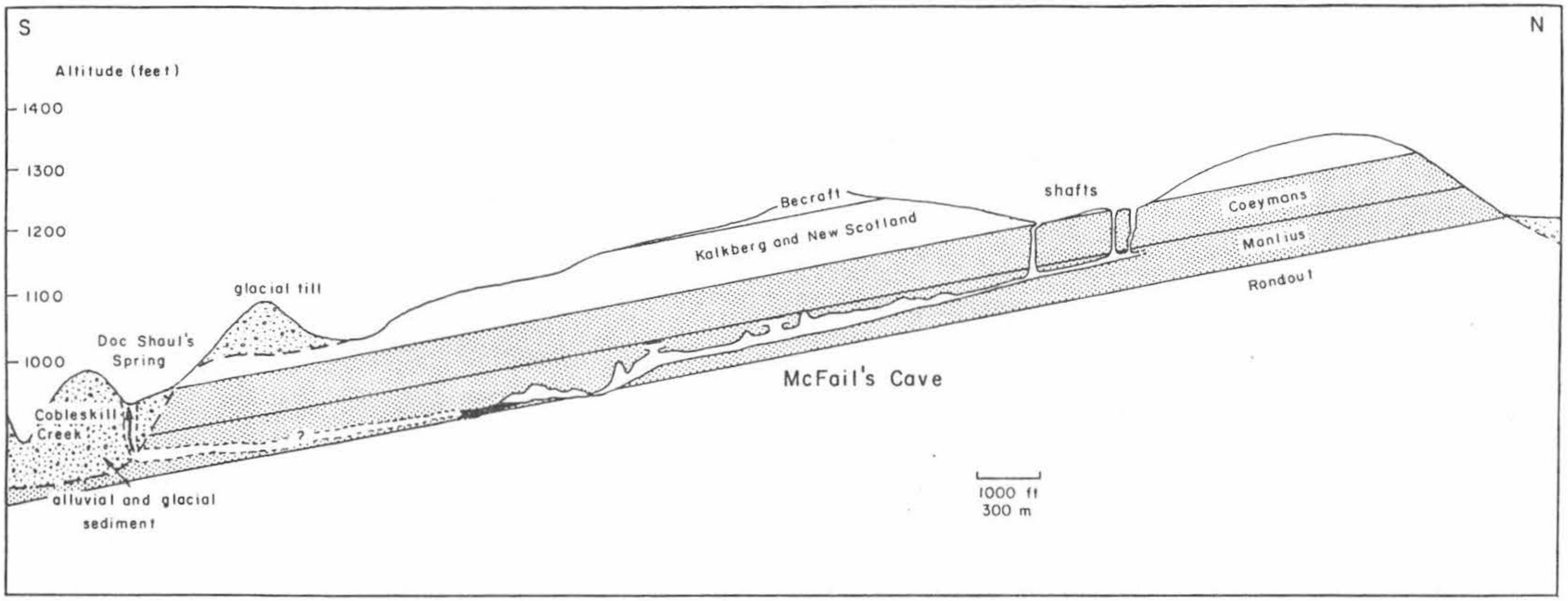

Figure 5: Geologic profile through the Cobleskill Plateau, showing the relationship of the Cobleskill valley and glacial deposits to the pattern of underground drainage.

welling 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? In McFail's Cave, which feeds the spring, calcite deposits show evidence for re-solution as much as 25 m above the present spring level, indicating prolonged or frequent flooding to that height. Saturated sediment has an effective specific gravity of about 1.1 (accounting for buoyancy and porosity), so this pressure head could have balanced the weight of about 23 m of sediment.

Stop 3 - Brown's Depression

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 swamps.

The buried valley, shown in Figure 4, was detected with gravity surveys, refraction seismology, and well logs by Palmer (1976) and with reflection seismology by Mylroie (1977). Its average depth of fill is 60-70 m.

Brown's Depression (Figure 4) appears to be an enormous sinkhole but is in fact an unfilled part of the preglacial valley. 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, in which case the depression was present during the last glacial retreat (Mylroie, 1977).

Stop 4 - Sinkholes and Shafts above McFail's Cave

Sinkholes are clustered in areas of the Cobleskill Plateau where groundwater recharge is most abundant. Stop 4 is located in a relatively low part of the plateau where the lower Kalkberg and Coeymans Limestones are exposed and is surrounded by higher areas of relatively impermeable overlying strata. Deep shafts, sinkholes, and sinking streams are well developed here (Figure 6). Their origin is mainly solutional, although some enlargement has taken place by erosion and subsidence of overburden and bedrock blocks. The effect of prominent vertical joints (with strikes of N 22° E and S 70° E) is clearly shown by the linearity of the shafts. These recharge points feed underlying McFail's Cave, which contains about 10 km of accessible passages. The property is owned by the National Speleological Society (see address in Road Log for access permission). The cave is closed to visitors, but similar features can be seen in Howe Caverns, which is open to the public.

The main part of McFail's Cave consists of a long sinuous canyon passage that descends from the recharge area in the direction of the stratal dip (Figure 7). It intersects a low-gradient tributary tube (Figure 8) that follows roughly along the strike of a low-angle thrust fault of northeasterly dip. It appears that the tube was once the upstream part of the main passage of Howe Caverns to the southeast, but late Pleistocene collapse and sediment fill have blocked the connection. The McFail's Cave water now follows an irregular diversion passage southwestward to Doc Shaul's Spring.

Many of the inactive passages in the cave are blocked by gravel, sand, and clay. Even the active stream passages show evidence of having been nearly filled with sediment at one time. The position and character of the sediments are well out of adjustment with the present flow regime, which suggests that they are related to glaciation. Laminated clays occupy the very lowest levels of stream entrenchment in the main passage. They are found only in caves in the drainage basin of Schoharie Creek (including tributaries Cobleskill Creek and Fox Creek), so they were probably deposited during a phase of glacial Lake Schoharie, when groundwater was rather stagnant. Laminated clays in McFail's Cave range from 970 to 1130 ft (296-344 m) in altitude.

The geomorphic sequence thus includes (1) fluvial entrenchment of the region, exposing the limestones; (2) development of cave passages during a preglacial (or interglacial) time, when the water table stood about 20-25 m higher than today; (3) further fluvial entrenchment, causing diversion of groundwater away from Howe Caverns to Doc Shaul's Spring; (4) filling of large parts of the cave with sediment, probably during glacial advance or retreat, and probably contemporaneous with aggradation of stream valleys; (5) excavation of much of the sediment by cave streams, possibly with some enlargement of passages by glacial meltwater; (6) deposition of lake clays and thin overlying gravels and sands; and (7) postglacial erosion of clays and other sediments. Some of these stages may actually overlap in time or represent sequential phases of the same event. Does the intervention of stage 5 between stages 4 and 6 indicate multiple glaciation, or simply different phases of a single glaciation? The close relationship of the sinkholes and cave to the present topography shows that the basic morphology of the plateau is fairly ancient. The cave is adjusted to the bedrock configuration, including the deep stage of fluvial erosion, whereas the glacial features are simply superimposed.

Despite the great concentration of groundwater recharge in this area, during most of the year the water entering the cave is slightly supersaturated with dissolved carbonate from the soil and the upper part of the bedrock. Degassing of carbon dioxide from the water in the cave reduces the ability of the water to hold dissolved carbonates. It is therefore unable to enlarge the cave by solution at these times, although small amounts of abrasional enlargement may take place. Only during periods of heavy overland flow does solutionally aggressive water enter the sinkholes. Most of the present cave enlargement takes place at these times. During the early stages of cave development, before the cave became aerated, groundwater was able to penetrate through the entire plateau before becoming saturated,

Figure 6: McFail's Hole, formerly a shaft entrance to McFail's Cave, now blocked by collapse material (Photos by A. Palmer).

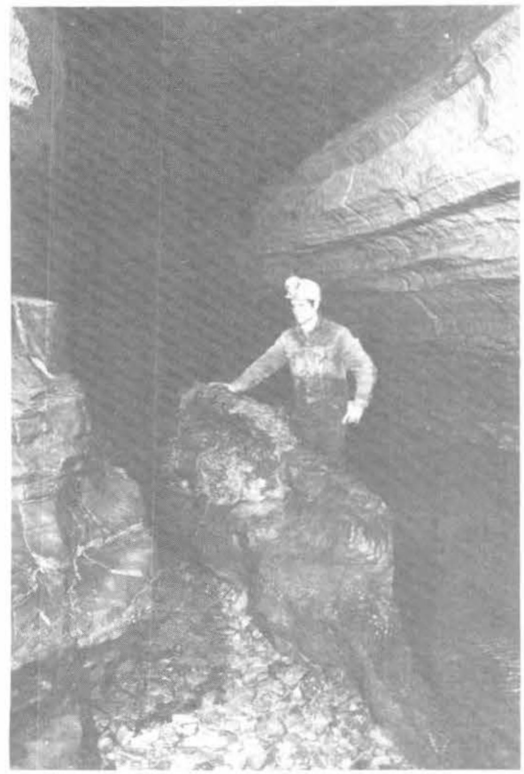

Figure 7: Down-dip vadose canyon passage in McFail's Cave. Rock pedestals in this area show evidence for two stages of down-cutting interrupted by a sediment fill stage. Remnants of cobble fill and glacial lake clay are visible in the foreground.

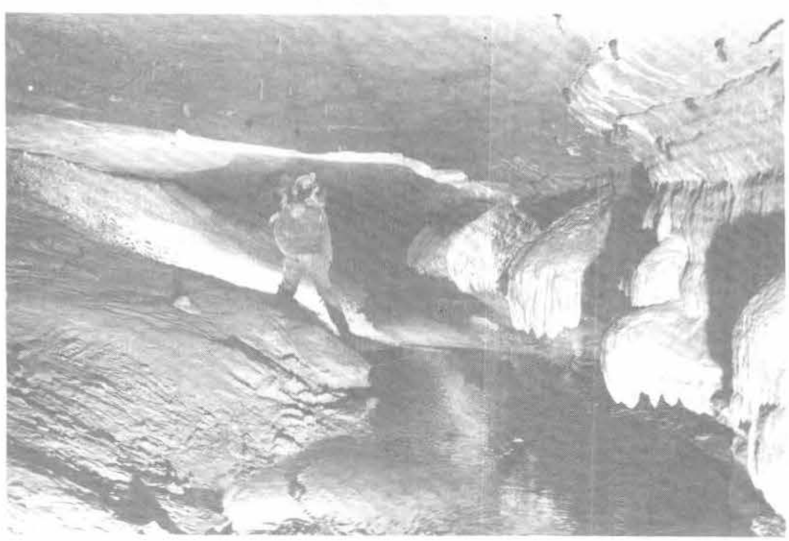

Figure 8: Strike-oriented tubular passage in McFail's Cave. Travertine has accumulated where water seeps into the passage along a gently dipping thrust fault.