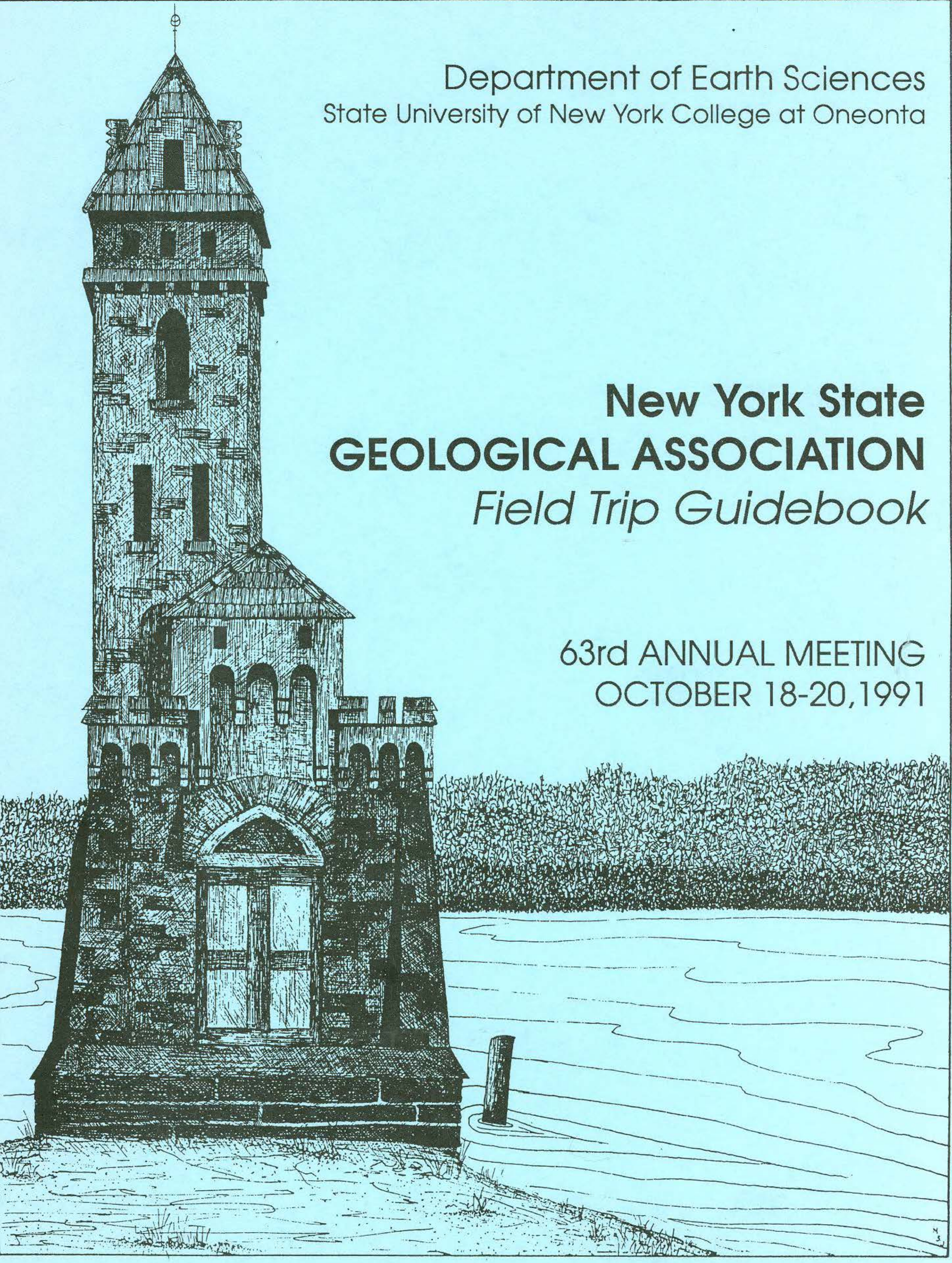
Department of Earth Sciences
State University of New York College at Oneonta

New York State GEOLOGICAL ASSOCIATION
Field Trip Guidebook

63rd ANNUAL MEETING
OCTOBER 18-20, 1991



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James R. Ebert, Editor

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Cover: Kingfisher Tower, Otsego Lake

DEDICATION

This Guidebook is affectionately dedicated to all those that share their knowledge of the Earth with students in the field. In this age of full-color textbooks, interactive software packages and exquisitely-produced videotapes, these teachers, from kindergarten through graduate programs, know the excitement that is instilled when students experience the Earth firsthand.

> *Geologists, in their all but closed conversation, inhabit scenes that no one ever saw, scenes of global sweep, gone and gone again, including seas, mountains, rivers, forests, and archipelagoes of aching beauty rising in volcanic violence to settle down quietly and then forever disappear - almost disappear.*
>
> John McPhee

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PREFACE AND ACKNOWLEDGEMENTS

We are very pleased to welcome you to the 63rd Annual Meeting of the New York State Geological Association and to the State University of New York College at Oneonta. Seven of this year's trip leaders were also contributors in 1977, when this department last hosted NYSGA. In the intervening fourteen years, tremendous advances have been made in our understanding of the geology of central New York.

From precision geochronology of precambrian rocks to processes of deglaciation and interaction of glacial and karst processes, geological investigation in the heart of New York is alive and well. Since 1977, we have witnessed the burgeoning of "sequence stratigraphy", the emergence of hydrogeology as a distinct subdiscipline and the expansion of process-oriented sedimentology. These changes are clearly represented by the spectrum of field trips in this guidebook. Sequence stratigraphy is applied to Lower Devonian Limestones, Middle Devonian clastics, and Quaternary sediments in Finger Lakes' basins. Hydrological studies of caves and glacial sediments are described. Physical processes of sedimentation in Devonian shelf, shore, and fluvial environments are also examined.

We hope that you enjoy the field trips and other activities this weekend and that you find them stimulating and enlightening.

The core of an NYSGA meeting is obviously the slate of field trips. The contributions of all the trip leaders (especially those that submitted manuscripts early!) are gratefully acknowledged. Without their efforts, the Association could not exist.

For all endeavors such as this, there is usually an unsung hero who is largely responsible for bringing the event to fruition. For typing the announcements, executing editorial changes, generally greasing the institutional wheels and reassuring a sometimes frazzled editor and frantic president, Moira Beach is our unsung hero.

We would also like to acknowledge the herculean efforts of Wayne Byam and his staff in the college print shop, Dorothy Gill for the cover layout, Bill Harman for supplying the sketch for the cover, Charlie Winters for shooting negatives for some figures and the administration of the college for their support and encouragement.

Finally, we thank the students of the Department of Earth Sciences, especially the Geology Club under the leadership of Heather Finlayson and Kelly Milunich for their logistical and clerical assistance.

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SUBSURFACE GEOLOGY OF THE FINGER LAKES REGION

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INTRODUCTION

The Finger Lakes of central New York State (Fig. 1) have long been recognized as the product of continental glaciation. In 1868 Louis Agassiz spoke of the "glacial heritage" of the region (Coates, 1968) and since then numerous glacial geomorphic investigations have been undertaken (Mullins et al., 1989). However, most previous studies dealt with surficial features and because of this relatively little was known about the subsurface Quaternary geology of these world renown lakes.

That the Finger Lakes are deeply scoured and infilled by thick sediment sequences was known from bathymetric surveys of the lakes completed in the late 1800's (Bloomfield, 1978), a drill record of 1,080' (329 m) of unconsolidated sediment at Watkins Glen (Tarr, 1904), and the publication of a line drawing of one seismic reflection profile from Seneca Lake (Woodrow et al., 1969). This paucity of subsurface geologic data has been a major void in our knowledge and understanding of the geologic history of the Finger Lakes. Many questions remained unresolved including the most fundamental -- "When and by what processes were the great Finger Lakes troughs eroded?" (Bloom, 1984, p. 61). Based on the one seismic reflection profile from Seneca Lake, Bloom (1984) noted that bedrock beneath the lake floor was more V-shaped like that of a river valley rather than the expected U-shaped glacial trough, and that there had been multiple erosional and depositional events. He concluded that more comprehensive seismic reflection surveys of the Finger Lakes "might reveal a barely suspected missing chapter in the history of the Finger Lakes region" (Bloom, 1984, p. 61).

During the summer of 1984 we initiated a seismic reflection investigation of the Finger Lakes with a pilot study of the northern half of Seneca Lake (Stephens, 1986) supported by a "starter" grant from Syracuse University. Preliminary results were very encouraging and revealed a detailed stratigraphy of a thick sediment fill. These preliminary data formed the basis of a National Science Foundation grant for a comprehensive seismic reflection study of all eleven Finger Lakes which was carried out during the summers of 1986 and 1987 (Mullins and Hinchey, 1989). The scientific objectives of this study were three-fold: (1) the provincial question of the origin and evolution of the Finger Lakes; (2) evaluation of the processes of deglaciation along the southern margin of the Laurentide ice sheet; and (3) to use the thick sediment sequence beneath the lakes as a record of continental environmental (paleo-limnology/paleoclimatology) change.

In order to fully realize these scientific objectives it was also necessary to extend our lake-based seismic surveys to adjacent on-land areas

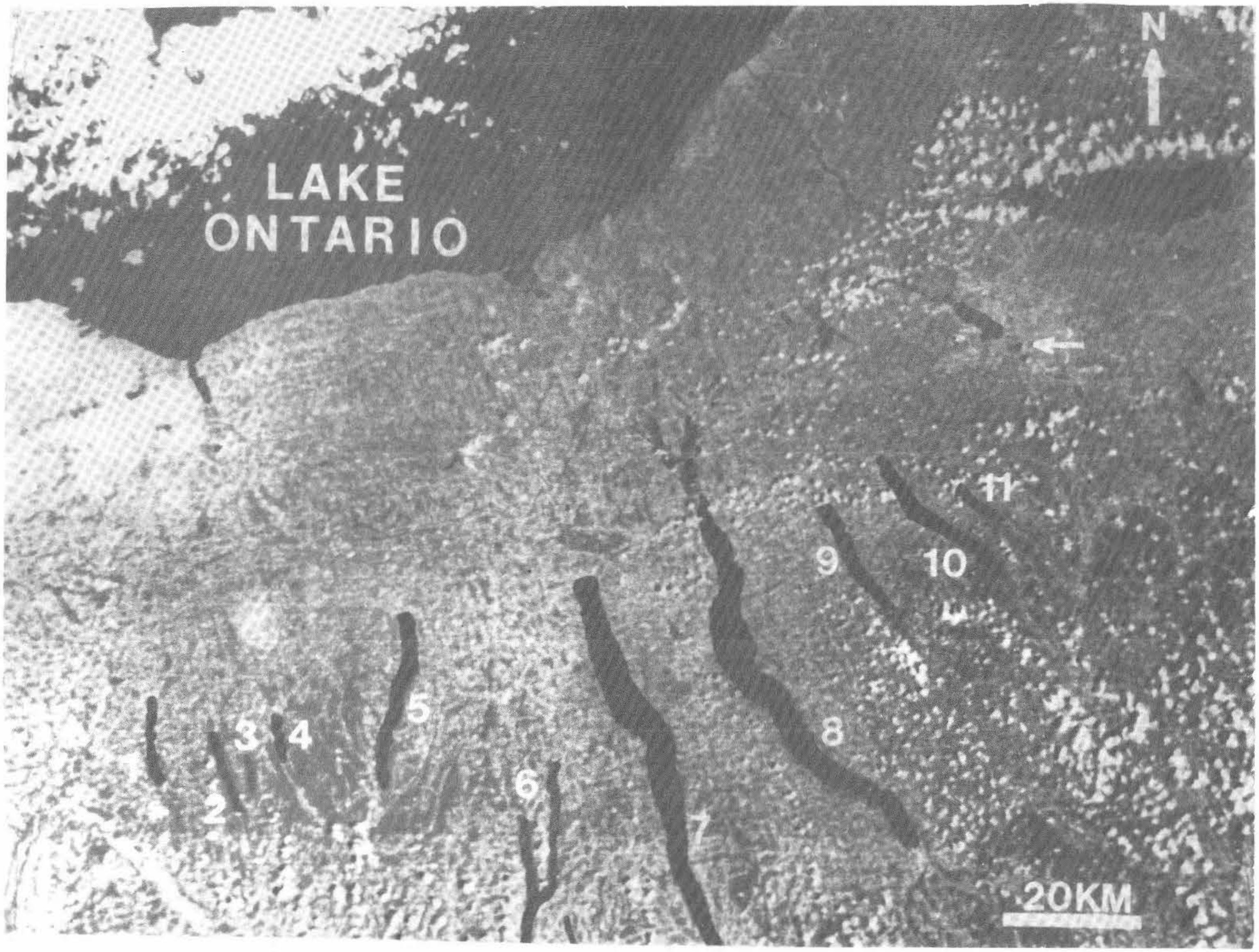


Fig. 1 - Satellite photograph of the Finger Lakes region. Numbers correspond to lakes: (1) Conesus; (2) Hemlock; (3) Canadice; (4) Honeoye; (5) Canandaigua; (6) Keuka; (7) Seneca; (8) Cayuga; (9) Owasco; (10) Skaneateles; (11) Otisco. Arrow points to city of Syracuse.

and to "ground truth" these geophysical data with subsurface geologic samples. Phase II of our investigation was also supported by NSF, and on-land multi-channel seismic reflection profiles were obtained during the summers of 1988 and 1990. In July 1990 we also recovered a 120 m (400') drillcore from the dry valley south of Canandaigua Lake, north of the village of Naples, N.Y. These new drillcore samples, which are presently being dated and analyzed, compliment 26 piston cores (up to 5 m long) collected from Seneca and Cayuga Lakes. Previous subsurface geologic information on Quaternary deposits beneath the Finger Lakes was limited to stratigraphic descriptions of water wells at Ithaca (Tarr, 1904) and short cores recovered from southern Cayuga Lake (Ludlam, 1967) and northern Seneca Lake (Woodrow et al., 1969).

The purpose of this field trip is to provide a new and unique perspective on the Quaternary geology of the Finger Lakes -- the shallow subsurface. Following a brief tour of surficial features, we will embark on a comprehensive overview of the subsurface Quaternary geology of the eastern Finger Lakes (Skaneateles, Owasco, Cayuga, Seneca, Keuka, and Canandaigua) as well as Tully Valley (a "dry Finger Lake") and Montezuma wetlands (a glacial meltwater system). We have planned 12 stops over a 1½ day period to examine and discuss seismic reflection, gravity, and drillcore data at acquisition sites. This non-traditional field trip has no outcrop stops so leave your shovels but bring your imagination!

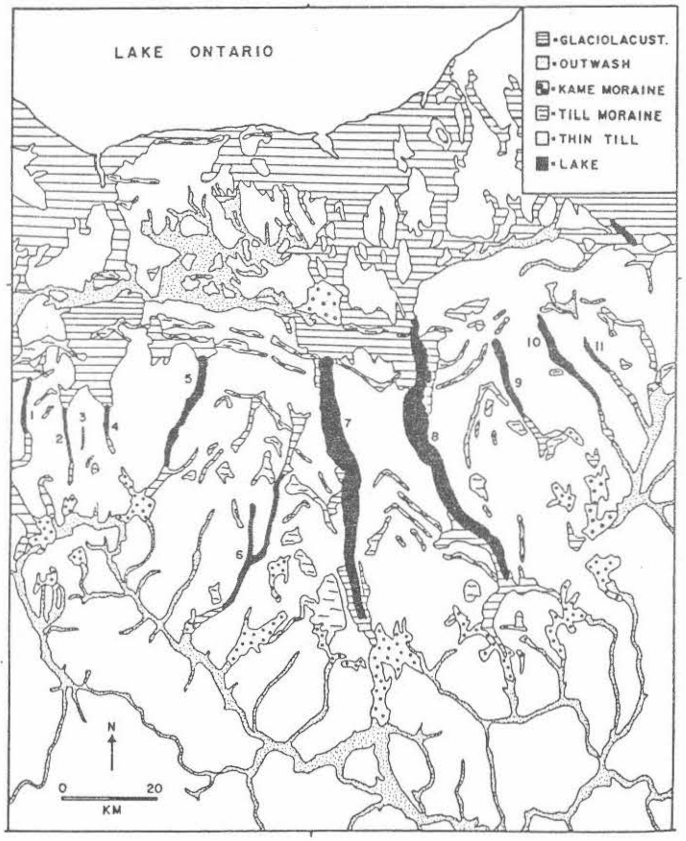
SURFICIAL GEOLOGY

The surficial geology of the Finger Lakes region has recently been compiled by Muller and Cadwell (1986; Fig. 2). Much of the upland area in the Finger Lakes region is covered by relatively thin till. However, a series of low-relief, chevroned, recessional till moraines also occur on the uplands in the central and southern regions of the lakes. At the north end of the lakes, recessional till moraines become more continuous in an east-west direction (Fig. 2).

North of the Finger Lakes is an extensive field of drumlins and meltwater channels which extend to Lake Ontario. Glaciolacustrine and swamp deposits selectively infill north-south oriented channels whereas an east-west system of channels is largely filled by outwash.

South of the lakes, and confined to the valleys, are well-developed kame moraines, collectively known as the Valley Heads moraine. These moraines are not connected over the uplands and consist principally of water-laid stratified drift. Based on regional correlations, Fullerton (1986) suggests that the Valley Heads moraines were deposited between 12.95 and 14.1 ka. If correct, Valley Heads deposition correlates with a pulse of rapid melting of the Laurentide ice sheet centered at 13.5 ka as recognized in the oxygen isotope record of North Atlantic and Gulf of Mexico deep-sea sediments (Ruddiman, 1987). To the south of the Valley Heads is a well-developed, arcuate system of outwash-filled valleys that likely served as meltwater channels during deglaciation.

There are relatively few radiocarbon dates available from the Finger Lakes region. Many of those reported (summarized by Muller and Cadwell, 1986) are "dead" dates >35 ka suggesting pre-late Wisconsin glacial deposits. Based on these dates Bloom (1986) has suggested that the valleys around Ithaca, and



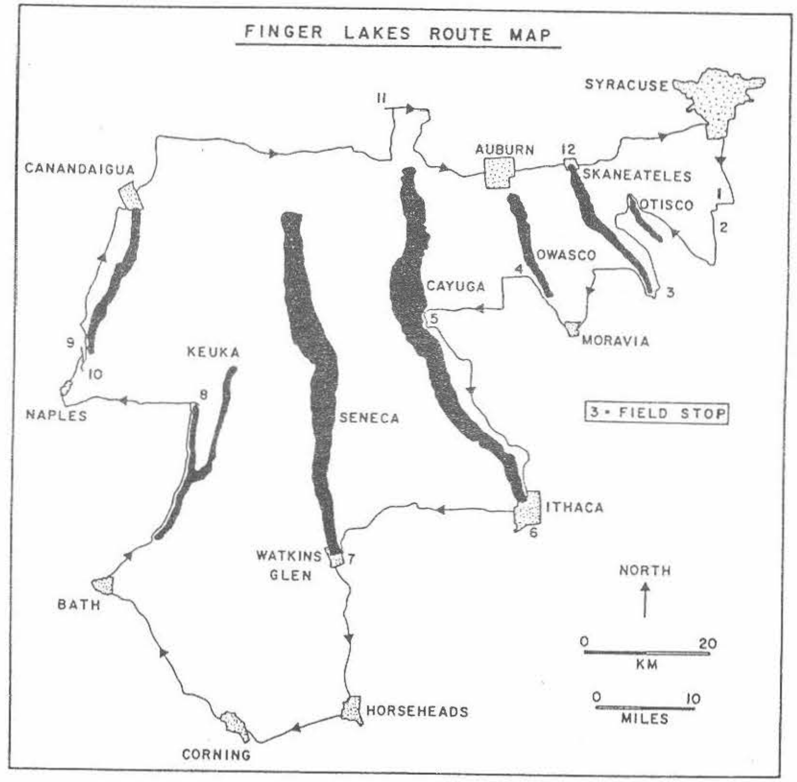


Fig. 3 - Field trip route map with stop locations numbered with text.

Fig. 2 - Surficial geologic map of the Finger Lakes region. Lakes are numbered as in Figure 1. Note Valley Heads ("kame moraine") at south end of lakes. Chevroned till moraines on uplands between lakes; outwash channels to the south. Simplified from Muller and Cadwell (1986).

presumably the other Finger Lakes valleys, were excavated "prior to two ice ages ago" (p. 263), well before the late Wisconsin.

FIELD TRIP STOPS

This trip will depart from the Heroy Geology Laboratory on the campus of Syracuse University. From Syracuse we will head south to Tully Valley and then across the southern margin of the eastern Finger Lakes to Keuka Lake with an overnight stay in Bath, N.Y. (Fig. 3). From Bath we will head northwest to Naples, N.Y. and Canandaigua Lake, then north and east to Montezuma Wetlands. A final stop for wrap-up discussion will be made at the north end of Skaneateles Lake before returning to Syracuse in the early afternoon (Fig. 3). The purpose, significance, and relevant data for each stop will be briefly described followed by a complete road log.

STOP 1: OVERVIEW OF A "DRY" FINGER LAKE

This brief stop at the dead end of Amidon Road provides (weather permitting) an outstanding south-directed view of Tully Valley, a "dry" Finger Lake valley. Clearly visible are the steep glacially-eroded bedrock walls of the valley and its relatively flat, sediment-filled floor which produced an overall U-shape profile. To the south is the Valley Heads moraine at Tully which completely fills the valley but does not extend over adjacent upland divides. The primary purpose of this stop is to give you a view of a "Finger Lake" without the water!

During late Wisconsin deglaciation glacial meltwaters ponded in this valley to form a proglacial lake more than 183 m (600') deep (Hand and Muller, 1972). The valley floor has an elevation today of about 152 m (500') and Bare Mountain on the west side of the valley crests at about 457 m (1500') which means that the valley was filled to about 60% capacity by this proglacial lake or about to the level we are standing (351 m, 1150').

As the ice receded to the north up contiguous Onondaga Trough east-west outlets were progressively exposed and the proglacial meltwaters in Tully Valley began to drain into the Mohawk through the Syracuse meltwater channels (Hand and Muller, 1972). Hand (1978) suggests that the outflow of water from Tully Valley occurred in a series of catastrophic flood events which ultimately led to the "dry Finger Lake" valley we see today. Since the draining of Tully Valley some 12 ka ago, post-glacial alluvium has washed into the valley, with the large alluvial fan at the mouth of the Rattlesnake Gulf being the most prominent.

From the vantage point of this overview stop we will drive down into this "dry Finger Lake" to examine subsurface data from the floor of Tully Valley.

STOP 2: THE FLOOR OF TULLY VALLEY

At this stop along Otisco Road you are now standing at the bottom of what once was a Finger Lake as deep as modern day Seneca Lake. How far below the valley floor does bedrock extend? How thick are the sediments beneath the valley floor and what is their geologic nature? To address these simple questions we have collected preliminary multichannel seismic reflection and gravity data which have been integrated with available drillhole information.

A gravity transect across Webster Road, 2.5 km north of our current location, suggests a total sediment fill of about 134 m (440') with bedrock eroded to within 30 m (100') of sealevel (Fig. 4). Sediment thickness of at least 117 m (385') has been confirmed there by a drillhole that did not reach bedrock (Fig. 4).

A multichannel (6-fold) weight-drop seismic reflection profile shot across Otisco Road reveals a high-amplitude reflector interpreted as bedrock at a maximum of 0.13 sec. of two-way travel time (Fig. 5). Calculation of true depth to bedrock is dependent upon the compressional wave velocity of overlying sediment. Based on wide-angle reflection experiments conducted in the Finger Lakes (Mullins and Hinchey, 1989) as well as seismic refraction data from southern Tully Valley (Faltyn, 1957) unconsolidated sediments beneath the Finger Lake valleys have a P-wave velocity in the range of 1.5-2.2 km/sec. Using this range of velocities, maximum subsurface depth to bedrock beneath Otisco Road is on the order of 98-130 m (321-4262'). If one accepts the higher range of velocities (2.0 km/sec.) there is good agreement between the seismic (130 m) and gravity (134 m) data. Both sets of data show a relatively flat bedrock floor beneath the axial portion of Tully Valley. Total bedrock erosion, as measured from the top of adjacent divides to bedrock beneath Tully Valley, has been on the order of 540 m (1770').

The seismic reflection data also suggest that there are three major stratigraphic sequences beneath Tully Valley (Fig. 5). The youngest sequence is characterized by discontinuous, high-frequency reflections and thickens to the west toward the alluvial fan at the mouth of Rattlesnake Gulf where it is as much as 60-80 m (197-262') thick. We interpret this sequence as post-glacial alluvial in-fill. The middle stratigraphic unit is largely "transparent" or reflection-free indicating massive sediment which we interpret as glaciolacustrine muds, that are as much as 53-70 m (174-230') thick. The top of this middle unit is undulatory which may be a consequence of loading by the overlying alluvial fill. The lowermost stratigraphic sequence beneath Tully Valley is also transparent and is characterized by a discontinuous, high-amplitude reflection at its top. This oldest sequence is relatively thin (19-25 m; 62-82') and is interpreted as a coarse-grained facies.

Our inferences on depth to bedrock and stratigraphic nature of the sediment fill beneath Tully Valley based on geophysical data are supported by available drillhole data. Getchell (1983) reported the gross stratigraphy of nine drillholes located in the southern end of Tully Valley (Fig. 6). Eight of the nine wells penetrated three major stratigraphic units: (1) an upper gravel/clayey gravel unit up to 24 m (80') thick; (2) a middle clay sequence 99-134 m (325-440') thick; and, (3) a basal gravel/sandy gravel unit at least 17 m (55') thick. The one well (WL-6) that did not penetrate the basal gravel unit reached shale bedrock at a subsurface depth of 142 m (465') which is at an elevation of 72 m (235') above sea level.

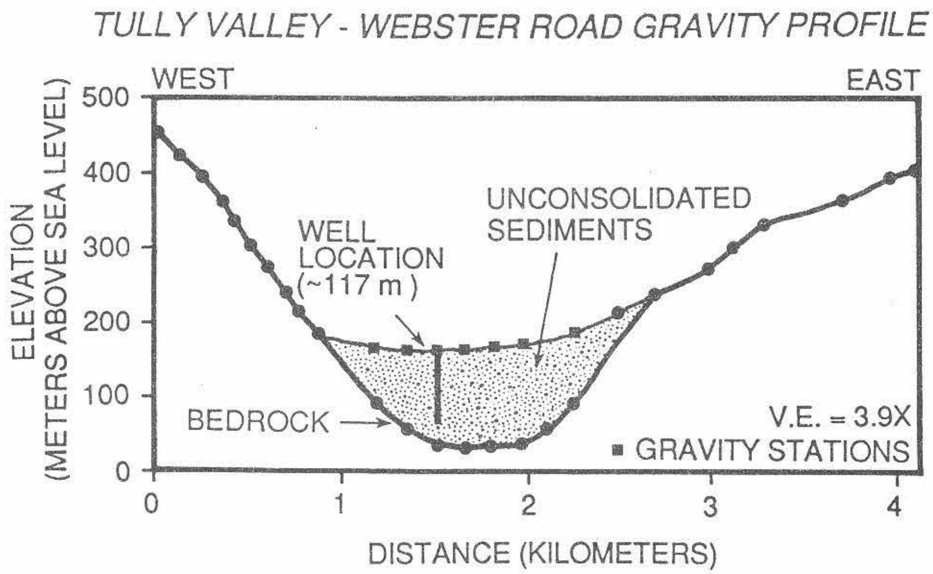


Fig. 4 - Cross-section of bedrock morphology and sediment fill of Tully Valley along Webster Road based on gravity data. Upland data points are based on topography.

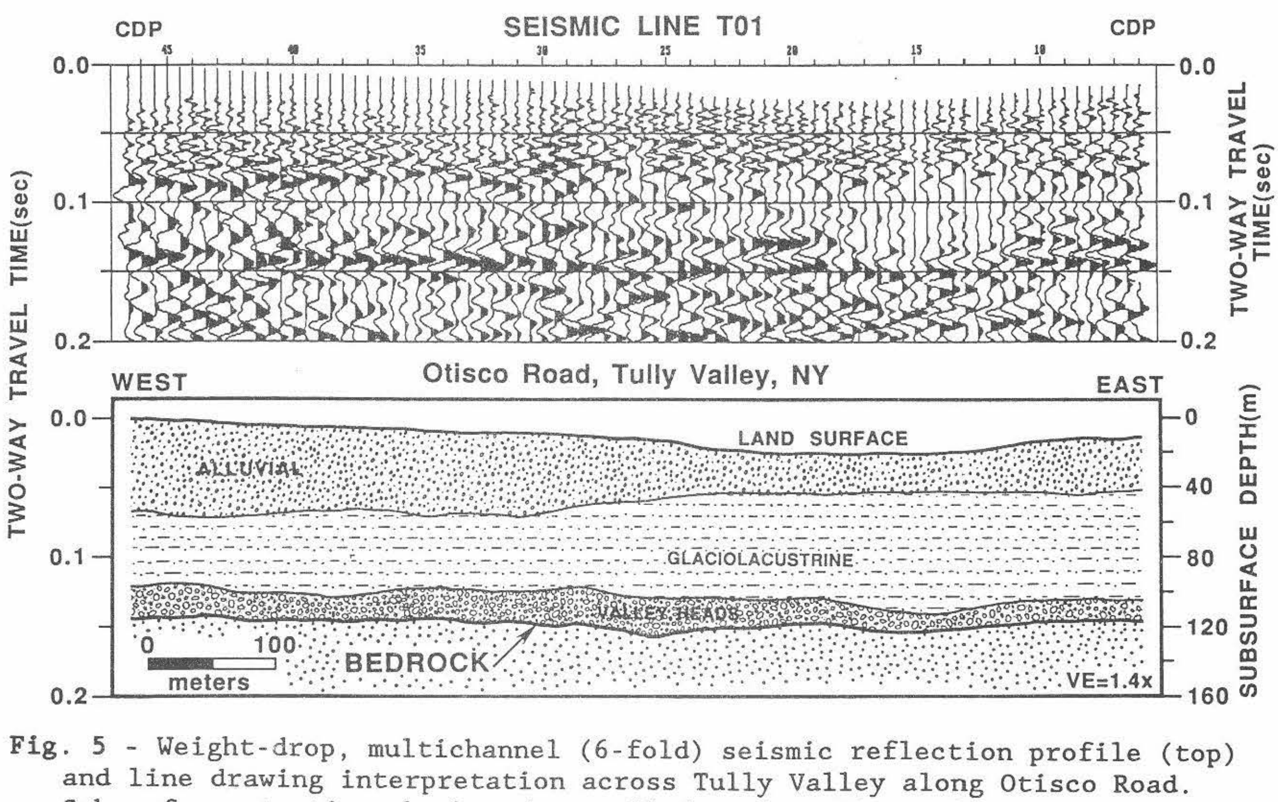


Fig. 5 - Weight-drop, multichannel (6-fold) seismic reflection profile (top) and line drawing interpretation across Tully Valley along Otisco Road. Subsurface stratigraphy based on well data (see Figure 6). Subsurface depth scale assumes a P-wave velocity of 1.6 km/sec.

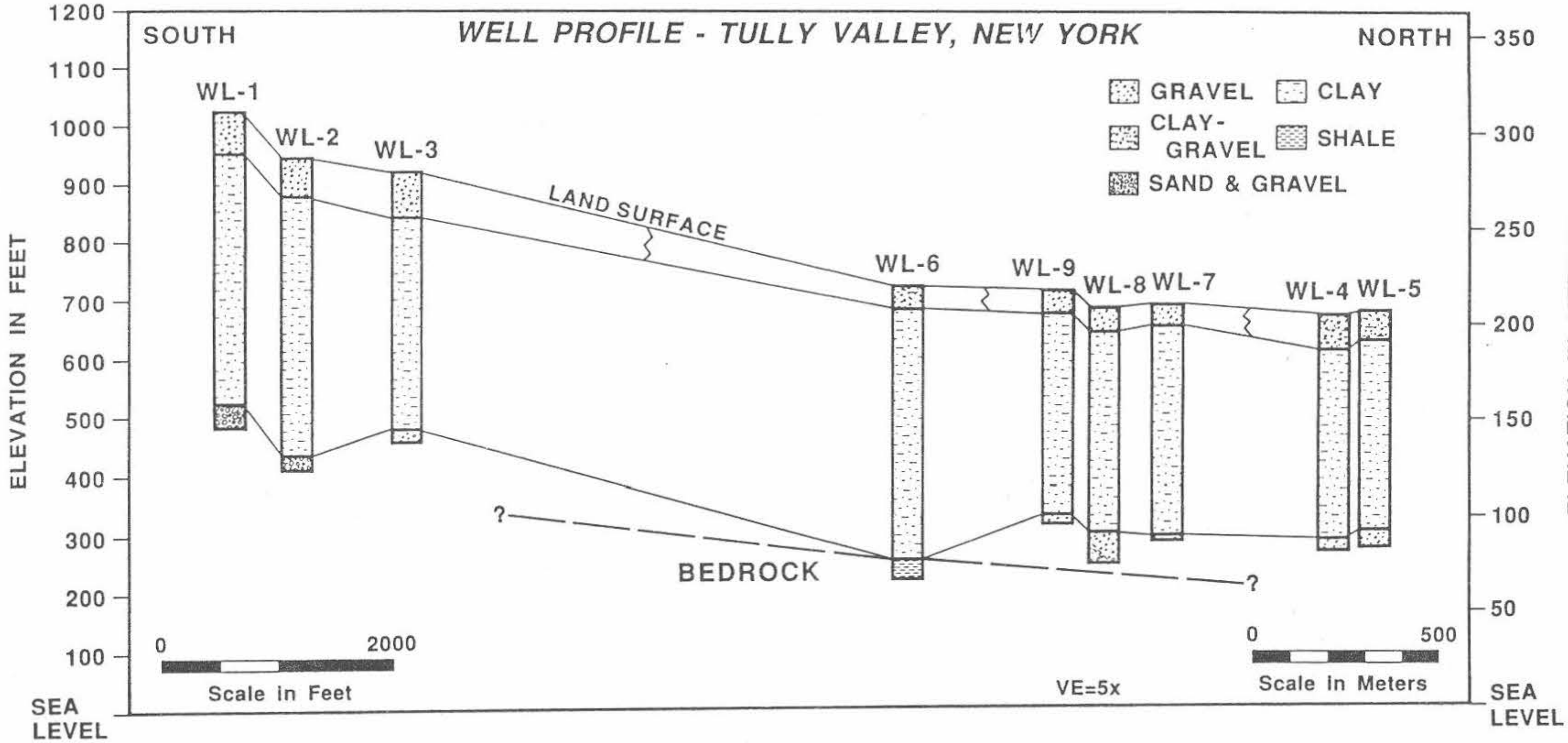


Fig. 6 - Schematic descriptions of south-north transect of wells in southern Tully Valley, based on reports in Getchell (1983). Wells are in their proper relative positions both laterally and with respect to elevation.

Combined, the subsurface geophysical and drillhole data reveal a deeply scoured trough beneath Tully Valley with bedrock rising to the south, which is consistent with reflection data from the Finger Lakes (Mullins and Hinchey, 1989). The data also indicate a relatively simple infill stratigraphy of washed subglacial (?) sands and gravel at the base overlain by glaciolacustrine muds which in turn are capped by coarse-grained alluvial gravels and clayey gravels. Although there is no chronostratigraphic control available, the gross lithostratigraphy beneath Tully Valley argues for a single overall infill sequence.

As we leave Otisco Road we will drive south along Tully Valley and then up the Valley Heads moraine which crests at an elevation of about 366 m (1200') which is some 213 m (700') above the floor of the adjacent valley. Gravel pit exposures here reveal washed sands and gravel that are crudely stratified. The Valley Heads deposits argue for an extended ice marginal environment here with large volumes of coarse-grained debris, both local and exotic, pumped through Tully Valley about 13-14 ka. The thickness of sediment beneath the Valley Heads moraine at Tully and the adjacent outwash plain has yet to be resolved, but is certainly an important question. Durham (1958) suggested that the position of the Valley Heads moraine was controlled by a pre-glacial drainage divide based on a well record at Little York which encountered bedrock at a subsurface depth of only 67 m (220'). However, it is not known whether or not this well, at an elevation of 354 m (1160'), was located along the valley axis. Preliminary gravity data obtained along Little York Road by Steve Wanzer, suggest that bedrock may be considerably deeper than that suggested by Durham (1958). This topic will be part of Steve's master's thesis, and hopefully he will have more definitive results by the time of our field trip.

From the Valley Heads moraine at Tully we will head northwest and drive around the north end of Otisco Lake (the easternmost Finger Lake), and then south to our next stop at the south end of Skaneateles Lake. When leaving the Valley Heads we will be able to see its contact with the bedrock wall of Tully Valley and observe that the moraine does not extend up and over the divide. At the northeast corner of Otisco Lake we will also pass by a hanging delta at Amber some 61 m (200') above the modern lake level of Otisco Lake, attesting to higher glacial lake levels.

Otisco Lake today is relatively shallow (maximum depth of 21 m or 69') and eutrophic which greatly limited our ability to collect seismic reflection data due to gas-saturated sediment (Hinchey, 1986). However, "windows" of seismic data indicate at least 83 m of sediment beneath Otisco Lake (Hinchey, 1986). As we drive around the north end of Otisco Lake we will (weather permitting) have an excellent view down the axis of the lake and Otisco Valley. The steep, U-shaped valley wall on the west side of the valley has a slope of about 35°. From Otisco Lake we will drive over the divide to Skaneateles Lake. Bedrock outcrops along the road attest to the relative paucity of glacial erosion and deposition on the uplands compared to the adjacent valleys.

STOP 3: SOUTH END OF SKANEATELES LAKE

This stop, high above the lake at the Onondaga-Cortland county line, provides a spectacular north-view of one of the most scenic Finger Lakes --

Skaneateles. We are standing at an elevation of 506 m (1660') which is slightly lower than the surrounding top of the plateau at about 579 m (1900'), but well above lake level which is at 263 m (863'). As you look to the north, it is quite apparent that the southern end of Skaneateles Lake is much more deeply incised than it is to the north. Also visible are three "points" along the west shore of the lake which represent post-glacial stream deltas that have built into Skaneateles Lake as higher glacial lake levels receded. The overall morphology of the Skaneateles Lake basin is similar to that observed in Tully Valley except that the valley is partially filled with water.

Maximum water depth in Skaneateles Lake is 90 m (295') which occurs at about the north-south midpoint of the lake. Seismic reflection data indicate that bedrock is relatively shallow beneath the north end of the lake but deepens toward the central portion of the lake where maximum erosion down to about sealevel has produced a closed rock basin (Fig. 7). Bedrock gradually rises toward the south end of the lake producing an overall spoon-shaped longitudinal profile which is typical of the Finger Lakes (Fig. 7; Mullins and Hinchey, 1989). Crossing lines indicate broad, U-shaped bedrock profiles at the north end of the lake which become progressively more incised and V-shaped toward the south (Fig. 8), which is also typical of the Finger Lakes (Mullins and Hinchey, 1989). Total erosion of bedrock, from the top of the plateau to the maximum extent of bedrock beneath the lake, has been on the order of 580 m (1900').

Maximum sediment thickness beneath Skaneateles Lake is on the order of 170 m (Mullins and Hinchey, 1989) with most of the sediment in the southern half of the lake basin (Figs. 7 and 8). The most regionally significant feature of the stratigraphy beneath Skaneateles is onlap at the south end of the lake of transparent to highly reflective sediments onto a southward thickening wedge characterized by a hummocky chaotic seismic facies (Fig. 9). The younger package of sediment is interpreted as largely northerly-derived glaciolacustrine deposits. The older chaotic wedge that thickens to the south, projects to on-land outcrops of Valley Heads. If Fullerton's (1986) correlation of Valley Heads at 13-14 ka is correct, this onlap at the south end of Skaneateles Lake implies that the sediment fill beneath the lake is all less than 14 ka, or of late Wisconsin to Holocene age. If older glacial sequences were deposited in Skaneateles valley they have been effectively removed by late Wisconsin glaciation.

Using this onlap relationship (14 ka) and a maximum sediment thickness of 170 m for Skaneateles Lake, we calculate an <u>average</u> rate of sediment accumulation of 12.1 m/1000 years. However, drilling south of Canandaigua Lake (see Stop #10) suggests that the post-glacial section is quite thin relative to the total thickness of sediment-fill which is consistent with our seismic stratigraphic interpretations. If 90% of the sediment fill beneath Skaneateles Lake was deposited over a 2 ka period between 12 and 14 ka, accumulation rates may have been as high as 77 m/1000 years. If correct, these calculations suggest that large volumes of sediment were rapidly deposited in the Skaneateles Lake basin during deglaciation.

A second stratigraphic feature of significance occurs at about the north-south midpoint of Skaneateles Lake. Here, a series of stacked, transparent wedges, each up to 15 m thick, occur (Fig. 10). These wedges pinch-out to the south over a horizontal distance of 3-4 km and are interbedded with highly

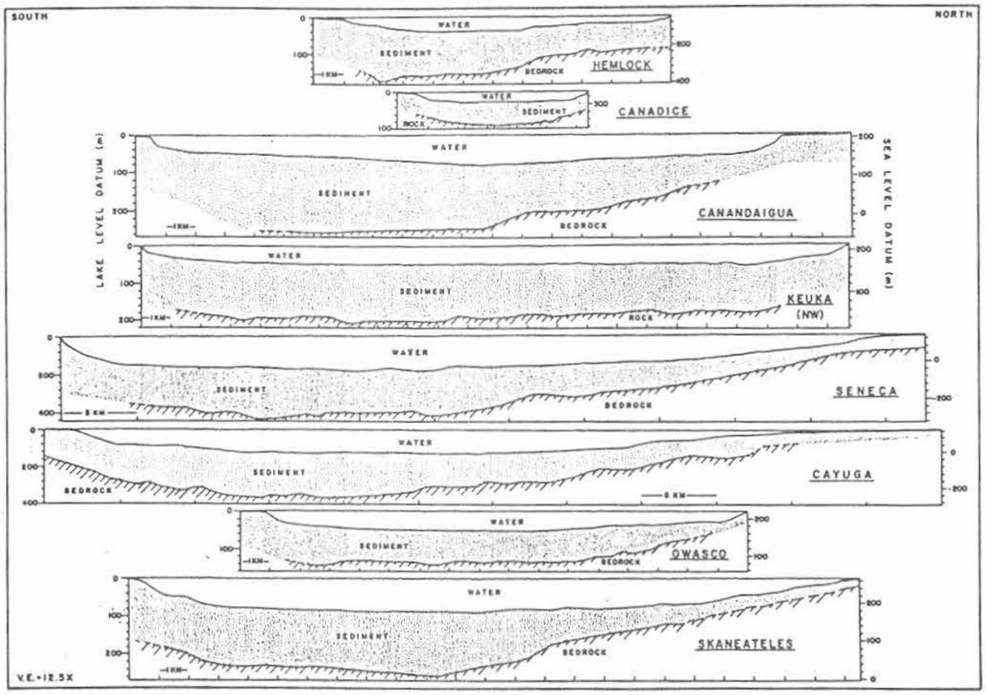


Fig. 7 - Schematic, longitudinal (N-S) bedrock and sediment-fill profiles of the Finger Lakes relative to both sea-level (right) and lake-level (left). Profiles based on reflection data collected from each lake. Scales for Seneca and Cayuga differ from other lakes; from Mullins and Hinchey (1989).

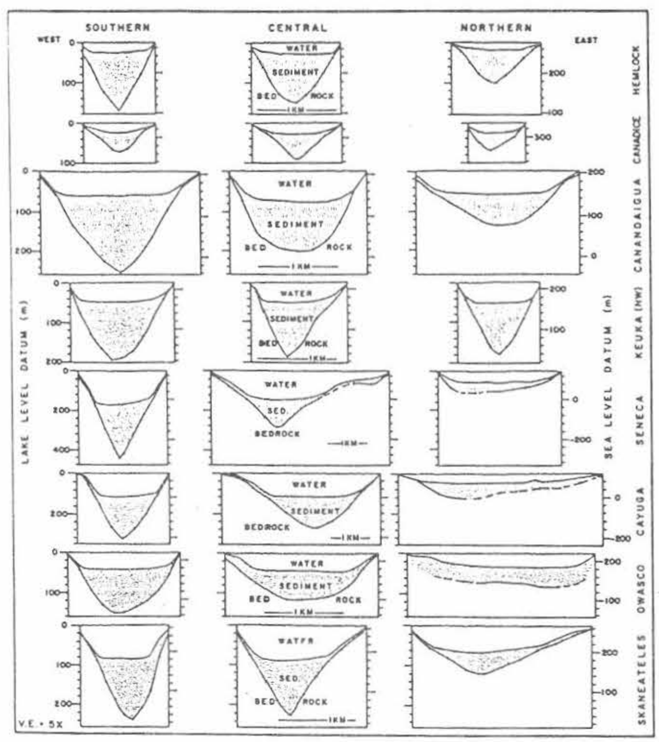


Fig. 8 - Schematic, transverse (E-W) profiles of bedrock and sediment-fill for the northern, central, and southern portions of the Finger Lakes relative to both sea-level (right) and lake level (left). Separate scales for Seneca and Cayuga Lakes; from Mullins and Hinchey (1989).

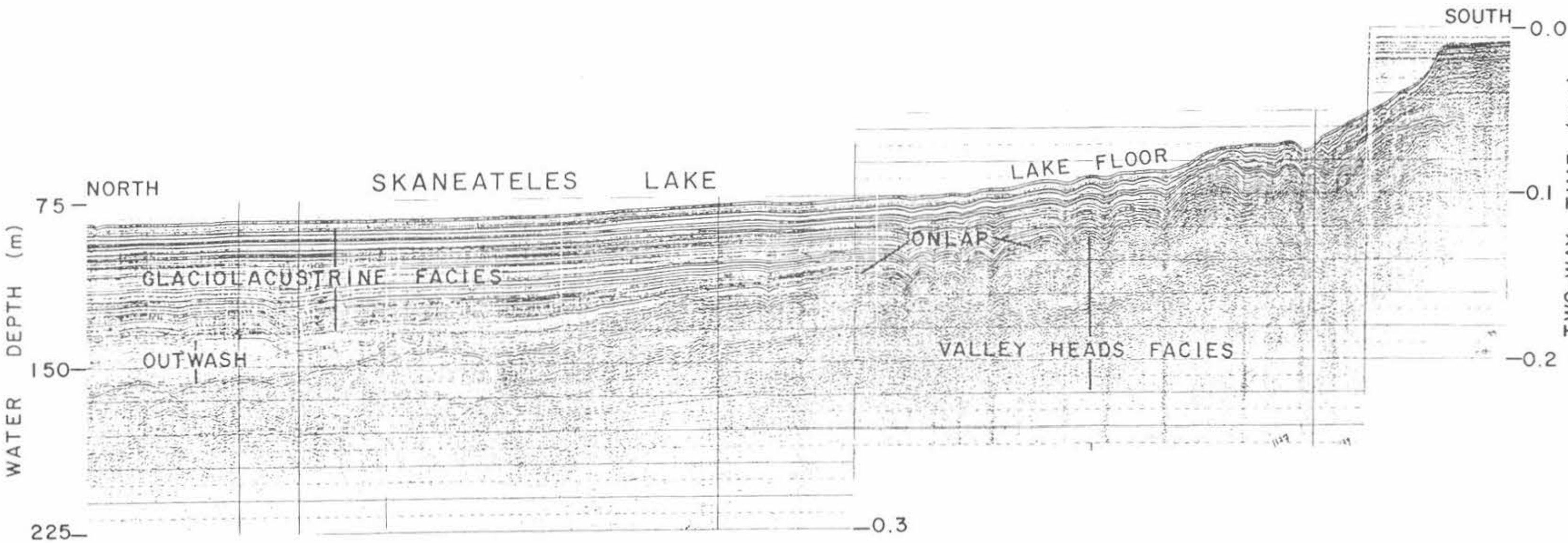


Fig. 9 - Southern portion of axial reflection profile from Skaneateles Lake illustrating onlap of glacio-lacustrine facies onto chaotic Valley Heads facies which thickens to the south. Water depth scale assumes a velocity of 1.5 km/sec.

reflective units interpreted as glaciolacustrine deposits (Fig. 10). Longitudinal (N-S) profiles (Fig. 10) indicate irregular tops with large diffractions. The seismically transparent (reflection-free) nature of these wedges indicates that they consist of massive sediment.

One interpretation is that these wedges represent "inflow events" deposited rapidly, issuing perhaps from a subglacial conduit; in essence a rapid pulse of massive debris. An alternative interpretation is that these wedges represent "till-tongues" similar to those described by King et al. (1991). Unfortunately, no samples from the wedges are available to resolve which (if either) interpretation is correct. However, Ed Hinchey promises to be on hand during the field trip to defend our view that these wedges represent sudden inflow events.

Regardless of the interpretation of these wedges, they suggest the presence of an ice margin that was either grounded or pinned here, where there is a break in the bedrock slope. North of these wedges total sediment thickness thins appreciably, consisting of a thin, highly-reflective (glaciolacustrine?) unit overlying a thin, chaotic (subglacial?) sequence. These stratigraphic relationships suggest that the ice margin in Skaneateles valley retreated rapidly (by calving?) to the north from its Valley Heads position to the lake's midpoint where it stabilized and issued large volumes of sediment into a proglacial lake.

From our overview of Skaneateles Lake we will drive down into the valley and around the south end of the lake. This will give you a good opportunity to get a "feel" for how deeply (and steeply) incised the south end of Skaneateles Lake is into bedrock of the surrounding Appalachian Plateau. As we cross the valley floor at the south end of Skaneateles Lake keep in mind that bedrock extends about 150 m (~500') beneath the surface. We will then drive up the west wall of Skaneateles Lake, over the divide to Owasco valley and through the village of Moravia to our next stop overlooking Owasco Lake at Ensnore.

STOP 4: OWASCO LAKE

This stop near the intersection of Ensnore Road and State Route 38 provides a good north-oriented overview of Owasco Lake. Elevation here is 323 m (1058') which is about half way between lake level at 217 m (711') and the crest of the upland divide at 402 m (1320') just to our west (Fig. 14).

Maximum water depth beneath Owasco Lake is 52 m (170'). Like the other Finger Lakes, it has steep walls and a relatively flat floor. Bedrock beneath the lake extends down to an elevation of ~80 m (262') above sea-level. Total erosion, from the top of the divide here at Ensnore to the bottom of bedrock beneath the lake, has been on the order of 320 m (1050').

Total sediment thickness beneath Owasco Lake is on the order of 100 m (328'; Figs. 7 and 8). The stratigraphy beneath the lake is particularly well-defined by our seismic reflection data. A basal, seismically chaotic sequence which thickens to the south, is overlain by a highly reflective unit (Fig. 12). We infer a single coarse-grained, subglacial sequence overlain by proglacial and post-glacial lacustrine deposits.

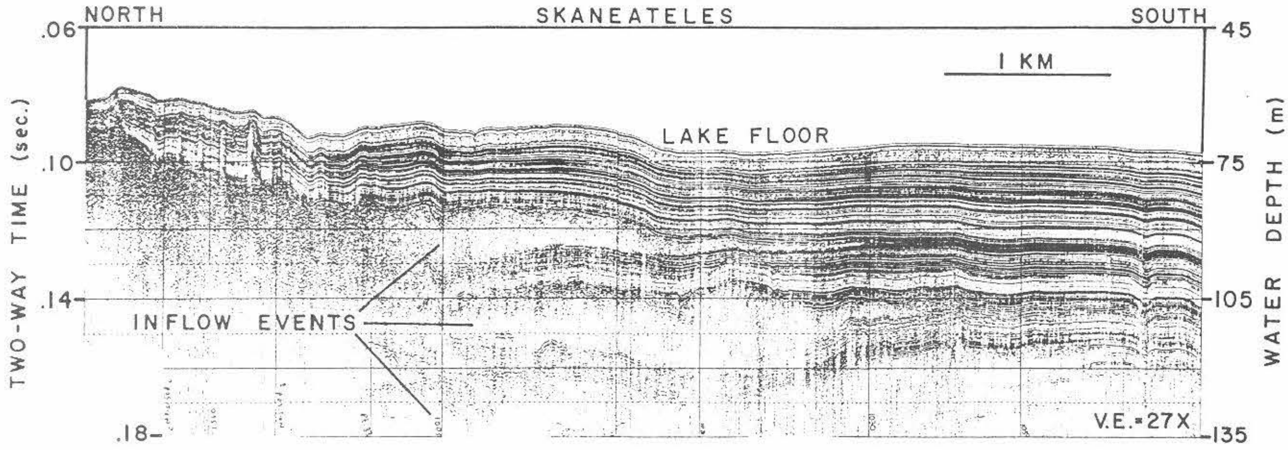


Fig. 10 - Central portion of axial reflection profile from Skaneateles Lake illustrating transparent wedges of sediment that are interpreted here as "inflow events" from a subglacial conduit into a proglacial lake.

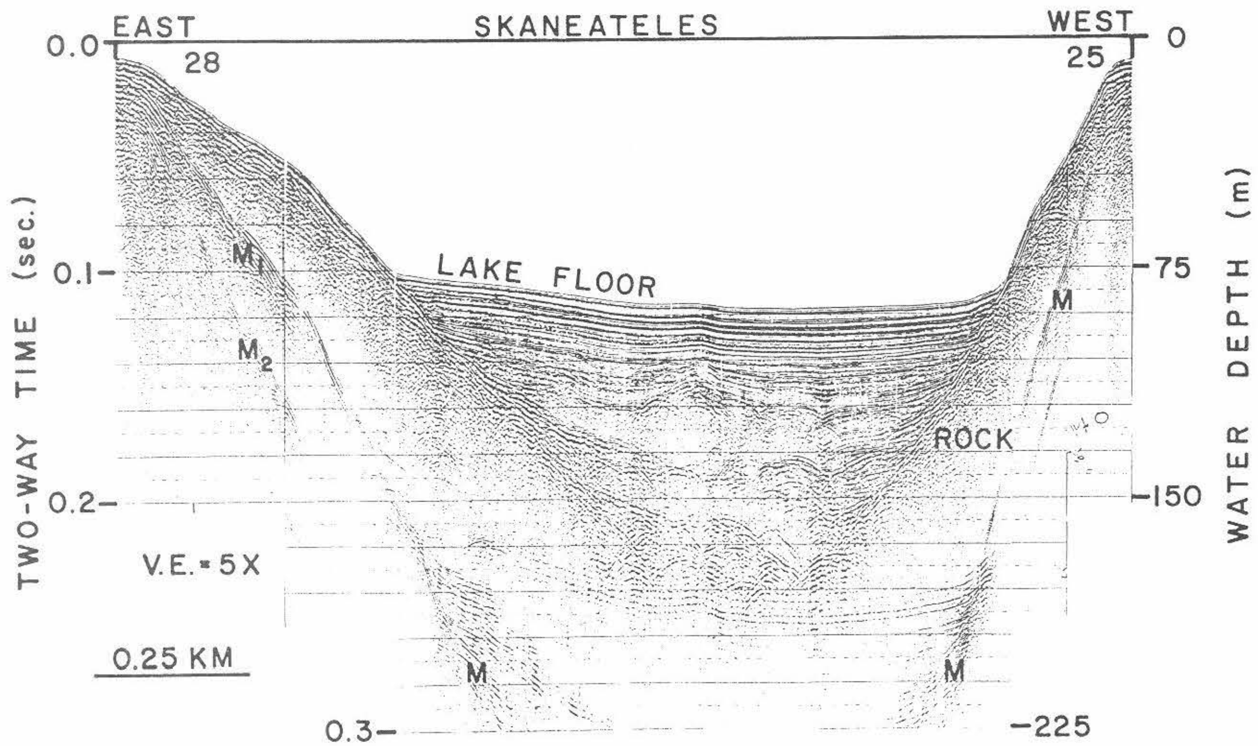


Fig. 11 - Transverse reflection profile from central Skaneateles Lake illustrating stacked wedges of debris with large diffractions at their tops. Note bedrock reflection and overlying highly-reflective sequence; M = multiple reflection.

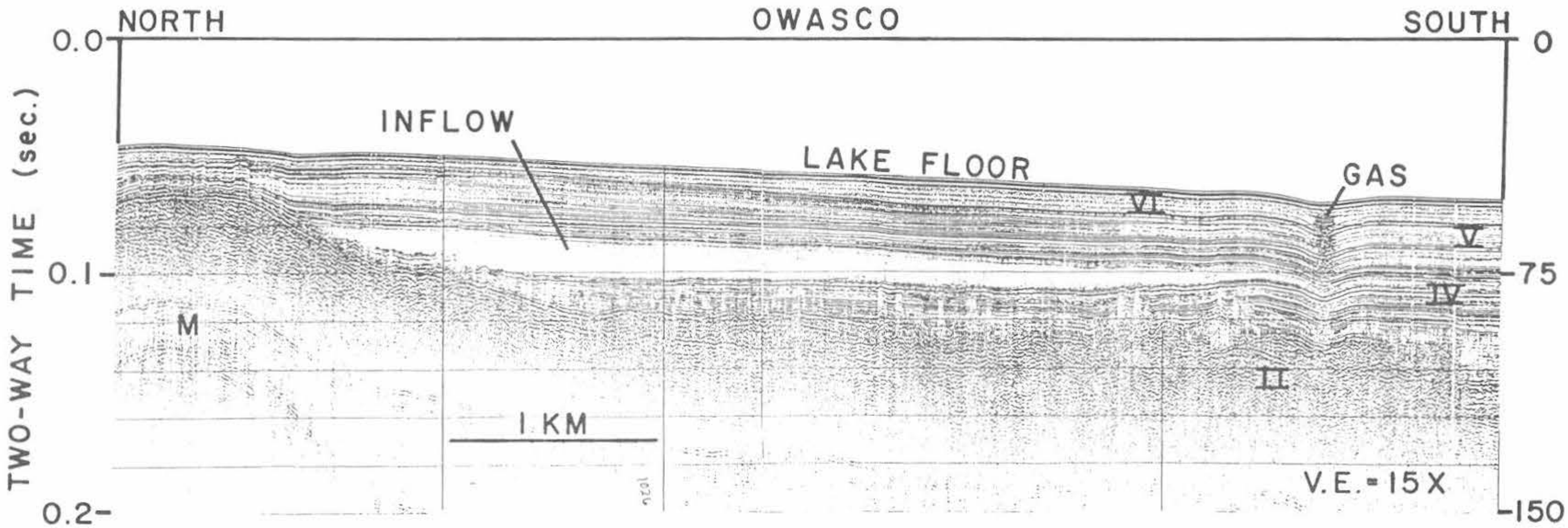


Fig. 12 - Central portion of axial reflection profile from Owasco Lake illustrating transparent sediment wedge interpreted as an inflow deposit. Roman numerals refer to depositional sequences; M = multiple.

In the northern third of Owasco Lake, an axial ridge is buried beneath the highly reflective lacustrine sequence (Fig. 13). This ridge, with up to 20 m (66') of relief, is sinuous and can be traced over a lateral distance of at least 5 km. We interpret this buried ridge as an esker which implies the past presence of a large subglacial meltwater conduit along the axis of northern Owasco Lake. Such an ice tunnel may have been responsible for the supply of coarse subglacial debris to the basal chaotic sequence beneath Owasco.

Another significant stratigraphic feature of the sediment-fill beneath Owasco Lake is a remarkably transparent (reflection-free) wedge of sediment in the northern half of the basin (Fig. 12). The transparent nature of this deposit indicates massive sediment which is interbedded with highly reflective glaciolacustrine sequences. Underlying units reveal only slight erosion and the top of this transparent unit is smooth. The wedge is up to 15 m (50') thick in the north and pinches out to zero thickness near the north-south midpoint of the lake. Transverse reflection profiles (Fig. 14) indicate that the wedge onlaps both the east and west bedrock walls of the lake. Although no subsurface samples are available, we interpret this transparent wedge as a subaqueous outwash fan of massive sand or silt, perhaps analogous to the deposits described by Rust and Romanelli (1973) near Ottawa. If this wedge represents in essence an "inflow event," it further argues in favor of the role of subglacial meltwater in the transport of debris to the southern margin of the Laurentide Ice Sheet, similar to the suggestion by Gustavson and Boothroyd (1987).

From our overlook of Owasco Lake we will head west over the divide toward Cayuga Lake. Just prior to entering the village of Poplar Ridge, we will cross one of the "chevron" till moraines mapped by Muller and Cadwell (1986). Radiocarbon analysis of wood overlying the south end of this till moraine yielded a date of 11,410 ±410 years (Muller and Cadwell, 1986).

STOP 5: CAYUGA LAKE

This stop is located at Long Point State Park at about the midpoint (east shore) of Cayuga Lake, just south of the village of Aurora. Cayuga Lake is one of the two largest Finger Lakes extending north-south for a distance of about 60 km with a maximum width of about 5 km. It has a maximum water depth of 132 m (433') located about two-thirds of the length of the lake form its north end. At this point of maximum water depth, the lake bottom is 15 m (51') <u>below</u> sea-level. Off Long Point maximum water depth is 100 m (361'). Lake level is at 116 m (382') which is the lowest elevation of all the Finger Lakes.

We have collected more than 150 km of high-resolution uniboom seismic reflection data from Cayuga Lake which have been correlated with 11 piston cores (Fig. 15). Bedrock beneath Cayuga Lake has been eroded as much as 358 m (1174') below lake level, which means that bedrock extends as much as 242 m (794') <u>below</u> sea-level!

A north-south schematic (Fig. 16) illustrates the overall "spoon-shaped" longitudinal profile of Cayuga Lake. Note that the maximum extent of erosion extends down to the Onondaga Limestone and appears to follow its southward dip

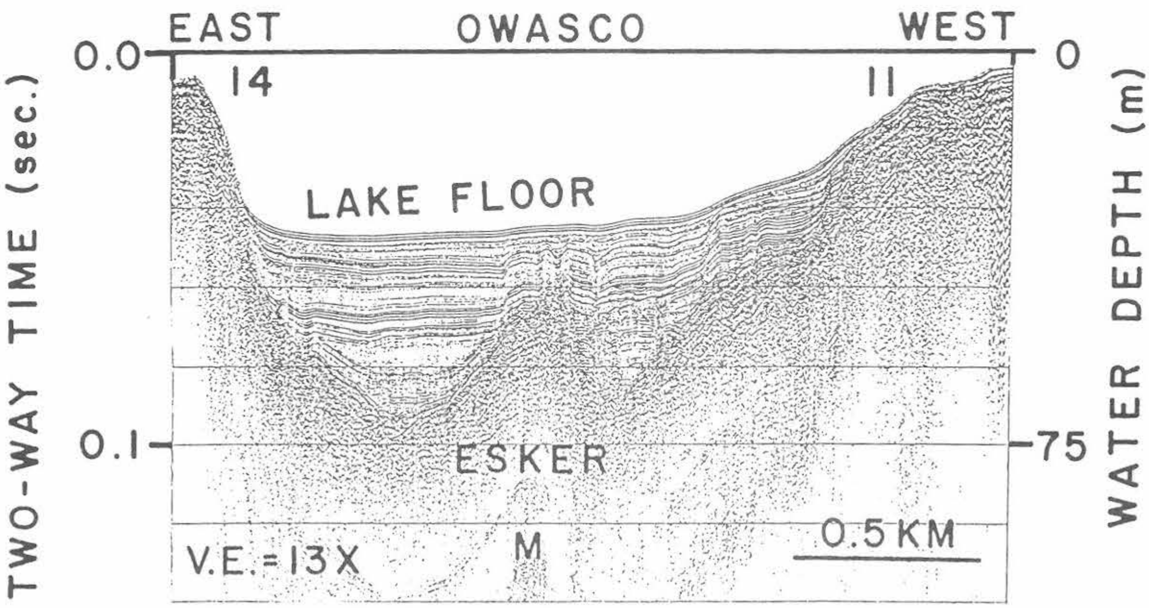


Fig. 13 - Transverse reflection profile across the northern portion of Owasco Lake illustrating buried, subsurface ridge interpreted here as an esker. Note small compactional "faults" at its crest; M = multiple.

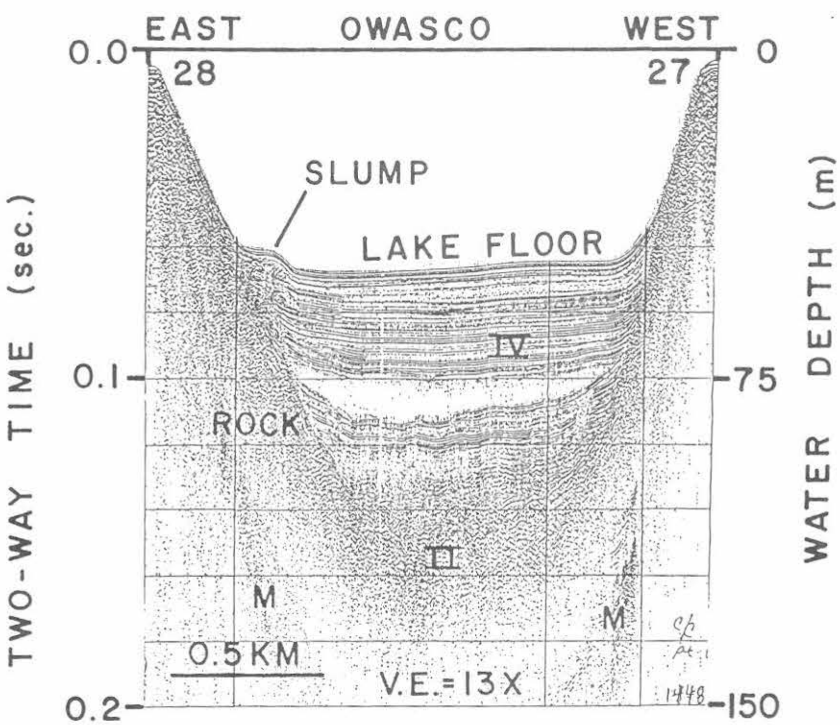


Fig. 14 - Transverse reflection profile across the central portion of Owasco Lake illustrating (in cross-section) transparent sediment wedge. Note onlap onto bedrock walls and small surficial slump on lake floor.

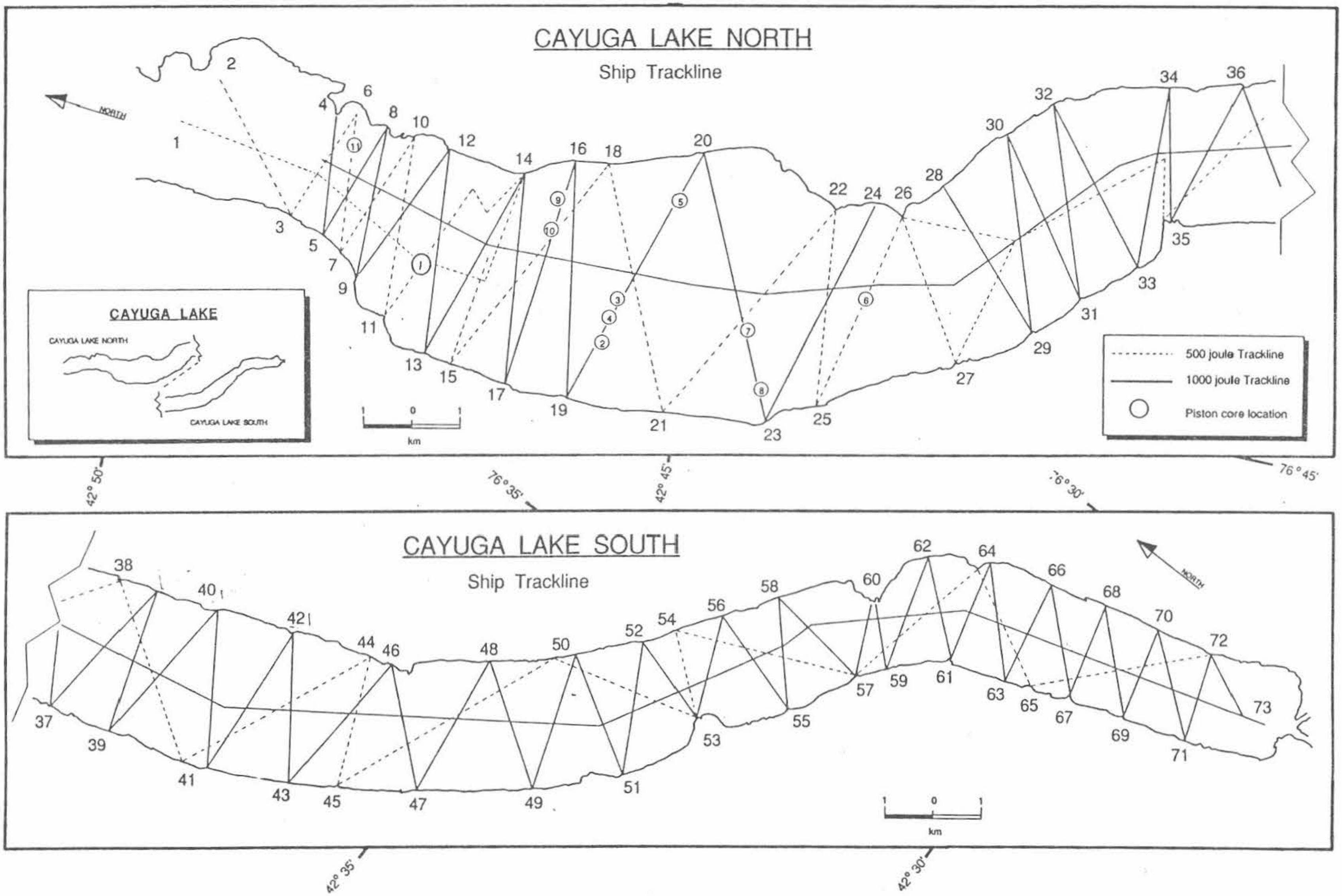


Fig. 15 - Geophysical trackline map for Cayuga Lake. Data spacing is typical of that collected for all Finger Lakes. Also shown are locations of piston cores.

before rising up through Hamilton Group shales at the south end of the lake. East-west schematics (Fig. 17) illustrate the broad, shallow form of the bedrock basin at the north end of the lake which becomes more narrow, and deeply incised to the south.

Maximum sediment fill beneath Cayuga Lake is 226 m (741') which occurs in the southern half of the lake basin. We have divided this fill into six seismically-defined depositional sequences: (I) The oldest sequence occurs only in the axial southern half of the lake basin where it is as much as 100 m (328') thick. This unit is characterized by a chaotic seismic facies and an irregular, hummocky upper surface which projects to Valley Heads outcrops south of the lake; (II) Sequence II is present throughout the lake basin and is up to 135 m (443') thick. It is characterized by a distinct north to south change in acoustic facies from chaotic to transparent (Figs. 18 and 19). A piston core (#10) located at the crest of an esker-like ridge along the top of sequence II (Fig. 18) recovered less than a meter of fine-grained rhythmites with dropstones before bottoming in clast-supported gravels (Fig. 20); (III) This sequence occurs throughout the lake basin as a transparent (reflection-free) unit up to 59 m (194') thick. At the north end of the lake this sequence is diapiric where it intrudes vertically up to 20 m (66'). Although unsampled, the massive, diapiric nature of this sequence suggests rapidly deposited muds; (IV) Sequence IV is a package of high-frequency, parallel reflections (Fig. 18) in which individual reflectors can be traced for 10's km along the length of the lake basin. It has a rather uniform thickness of about 30 m (98') except at the north end where it is more than 60 m (197') thick. Piston cores from lake floor outcrops of this sequence, recovered fine-grained, centimeter-scale rhythmites with occasional dropstones that were deposited in a proglacial lake; (V) This sequence records a "turning point" in the history of Cayuga Lake. It is thickest (>30 m; 98') at the south end of the lake basin as well as off major "points" such as Myers Point and Sheldrake Point; and, it is not present at the north end of the lake. This sequence heralds the beginning of northward drainage and lateral sediment input to Cayuga Lake that likely accompanied the lowering of proglacial lake levels and erosion of regional, upland gorges as ice withdrew from Cayuga Valley (Fairchild, 1934a); (VI) Sequence VI is a low-amplitude to transparent seismic unit that is variable in thickness (8-19 m; 26-62') throughout the lake basin. Piston cores have recovered fine-grained organic-rich sediments from this sequence (Fig. 20), some of which are "banded" and were perhaps deposited by turbidity currents (Ludlam, 1967).

After lunch at Long Point State Park we will head south to Buttermilk Falls State Park near the south end of Cayuga Lake at Ithaca.

STOP 6: SOUTH END OF CAYUGA LAKE

The purpose of this stop at scenic Buttermilk Falls State Park is to examine and discuss land-based, seismic reflection data (acquired from Cayuga valley across from the Park) that have been integrated with results from 13 wells reported by Tarr (1904). The bedrock reflection extends as much as 0.19 sec. of two-way travel time beneath the valley (Fig. 21). Using a range of interval velocities from 1.5 to 2.0 km/sec., maximum depth to bedrock here is 143 m (467') to 190 m (623'). Tarr (1904) reported a salt well along the western margin of Cayuga valley at Ithaca, that encountered bedrock at 131 m

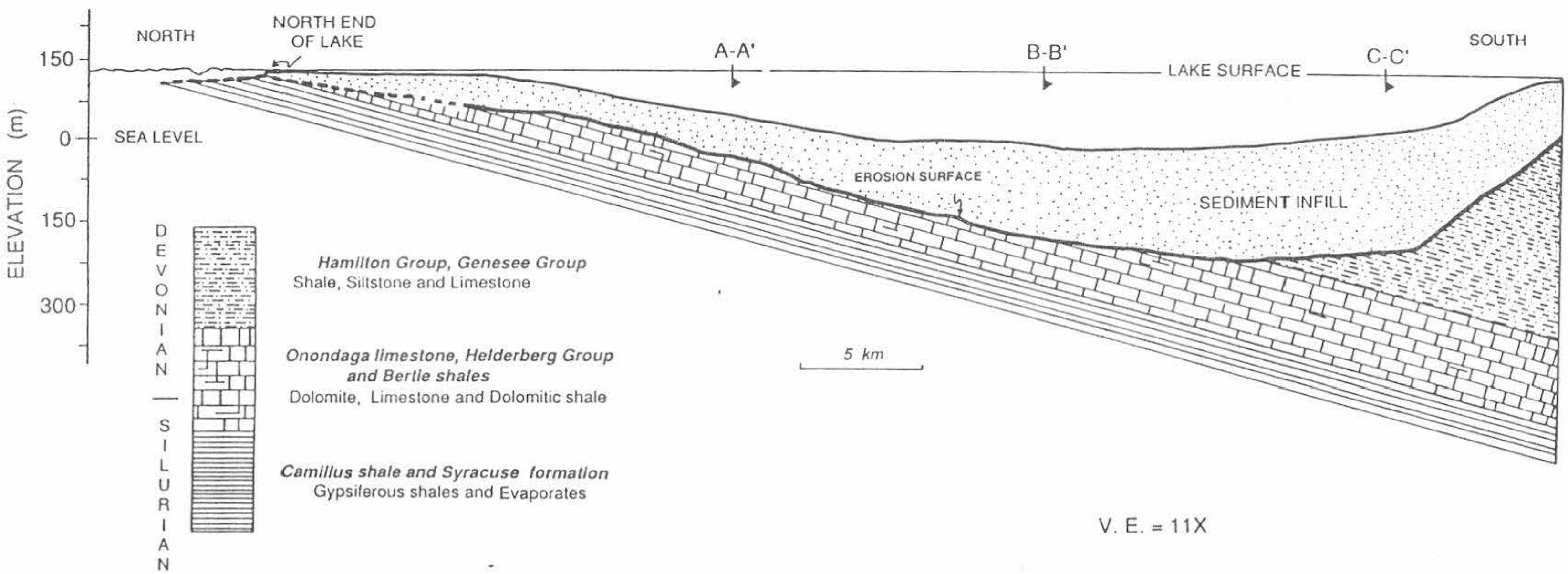


Fig. 16 - Schematic axial (N-S) profile of bedrock geology and sediment-fill for Cayuga Lake. Crossing profiles A-A′, B-B′, and C-C′ illustrated in Figure 17.

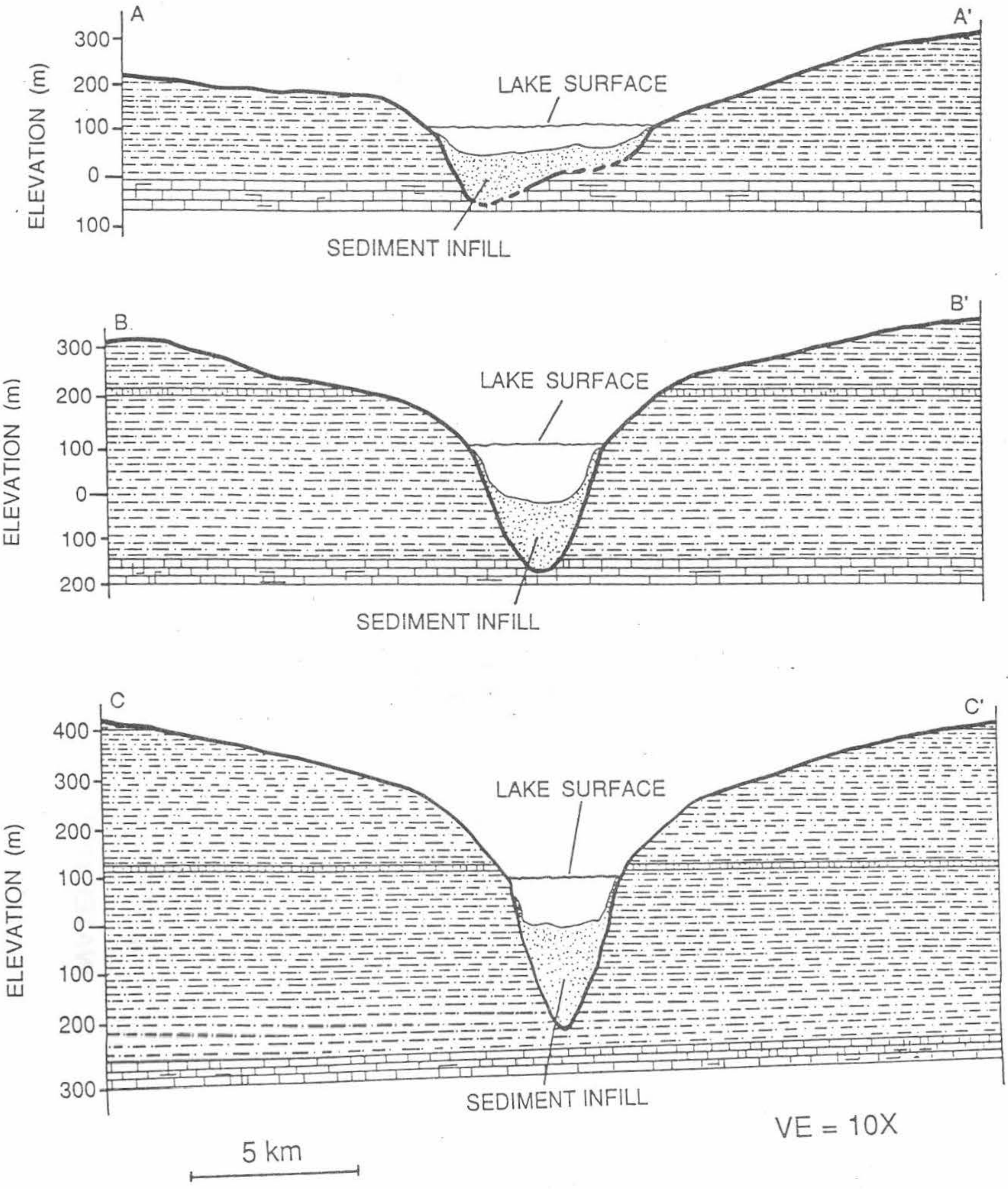


Fig. 17 - Schematic transverse (E-W) profiles of bedrock and sediment-fill for northern (top), central, and southern (bottom) Cayuga Lake. See Figure 16 for locations and legend.

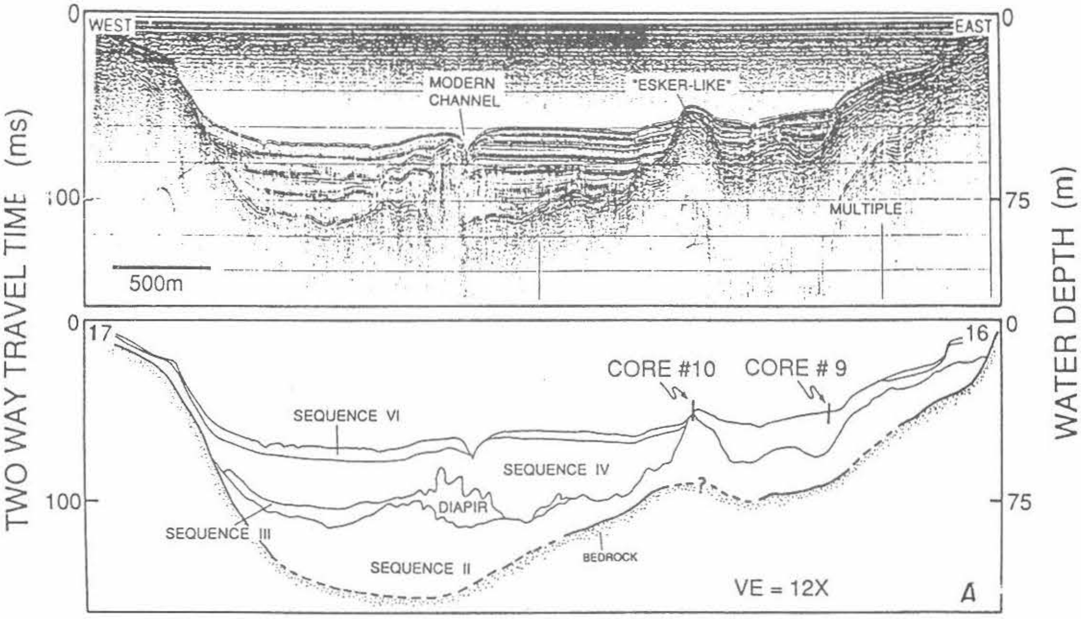


Fig. 18 - Transverse seismic reflection profile (top) and line-drawing interpretation (bottom) across the northern portion (16-17, Fig. 15) of Cayuga Lake. Note depositional sequences and location of piston cores.

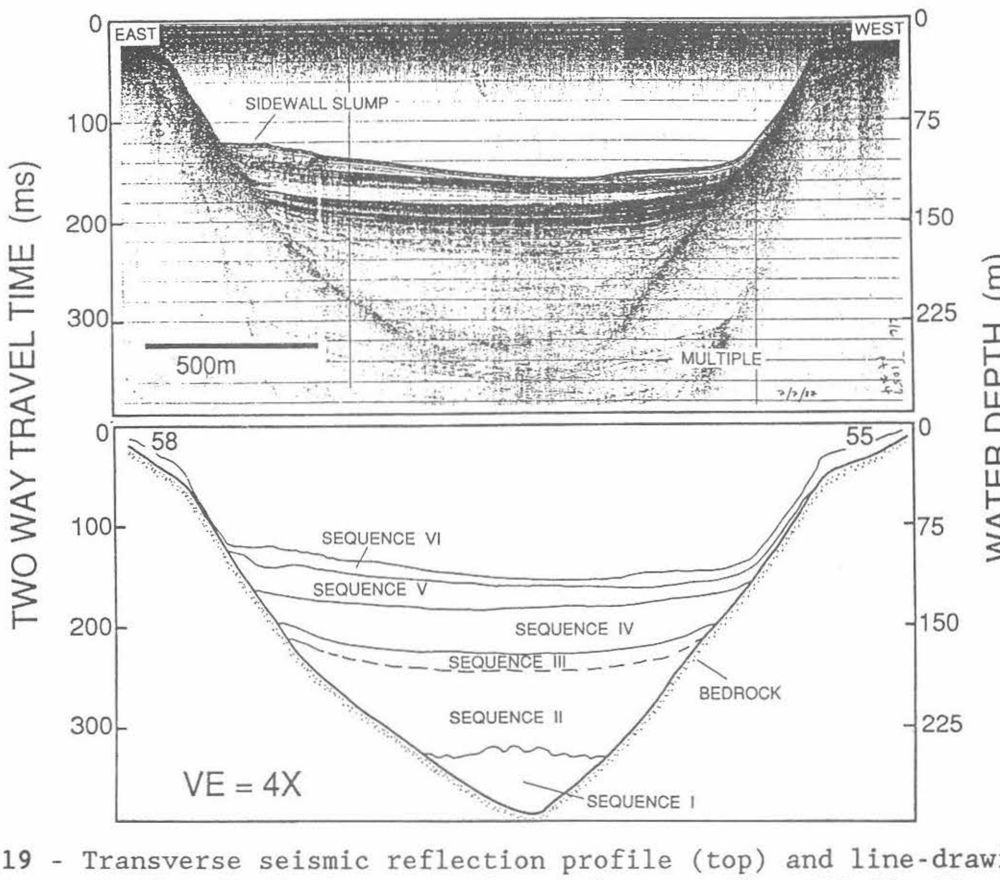


Fig. 19 - Transverse seismic reflection profile (top) and line-drawing interpretation (bottom) across the southern portion (55-58, Fig. 15) of Cayuga Lake.

which provides a minimum value for depth to bedrock because of its western location. Maximum sediment thickness beneath Ithaca is probably on the order of 166 m (545').

Our land-based seismic reflection profile (which is of lower resolution than our lake data) also reveals four major stratigraphic units beneath the valley: (1) an upper transparent (reflection-free) unit up to 20 m (66') thick (assuming an interval velocity of 1.5 km/sec.); (2) a high-frequency unit about 19 m (62') thick; (3) a sequence, bound by high-amplitude reflections, up to 28 m (92') thick (assuming a velocity of 1.6 km/sec.), and, (4) a basal unit characterized by discontinuous high-amplitude reflections up to 85 m (279') thick (assuming a velocity of 1.7 km/sec.).

The stratigraphy of Tarr's (1904) 13 wells, all located along the western margin of the valley, is schematically illustrated in Figure 22. The upper layer in these wells is predominantly fine-grained massive clay (up to 18 m or 60' thick) with fragments of mollusks, plants, and wood, including logs. Beneath these organic-rich clays is a series of sand and gravel layers 6-21 m (20-70') thick. According to Tarr (1904), the sands are well-washed and the gravels well-rounded. Plant fragments, mollusks and logs also occur in these coarser beds.

In all wells, these coarser deposits are underlain by "a great thickness of clay" devoid of mollusk fragments (Tarr, 1904). There are also minor coarser units within this clay layer including scratched, angular pebbles. The base of this clay layer is irregular, occurring at subsurface depths between 61 m (200') and 85 m (280') in the 13 wells (Tarr, 1904).

Beneath this clay layer is a heterogenous, coarse-grained unit from which artesian water flowed (up to 300,000 gallons/day!) in many of the wells. The total thickness of this basal coarse-grained unit is not known as 11 of the 13 wells bottomed in it. Tarr (1904) describes this coarse basal unit as consisting of variable sediment ranging from washed sand and gravel, to till, to "quicksand."

Tarr (1904) interpreted all these subsurface sequences as late Wisconsinian -- "Neither here nor in the other well that reached rock, nor, in fact, in any of the wells, was any older drift encountered. All the materials are such as might have been brought by the last ice advance, or deposited since the ice-sheet melted away." His view of the heterogenous basal coarse layer was that it is morainic; a subsurface extension of the nearby Valley Heads. We concur with Tarr's interpretation and our seismic data suggest that this basal Valley Heads facies extends to bedrock and is more than 100 m (328') thick. We correlate this unit with seismic sequence I beneath Cayuga Lake.

The middle clay layer with scratched angular stones, is interpreted by both Tarr and us as a proglacial lacustrine sequence, perhaps with dropstones which would suggest the presence of icebergs. We correlate this proglacial clay layer with seismic sequence IV beneath the lake. The overlying washed sands and gravels appear to be a fluvial/alluvial or shallow lacustrine facies, that likely records a drop in lake level and drainage reversal as ice receded from the north end of Cayuga Lake (seismic sequence V; Mullins and Hinchey, 1989). The succeeding organic-rich clays at the top of the section

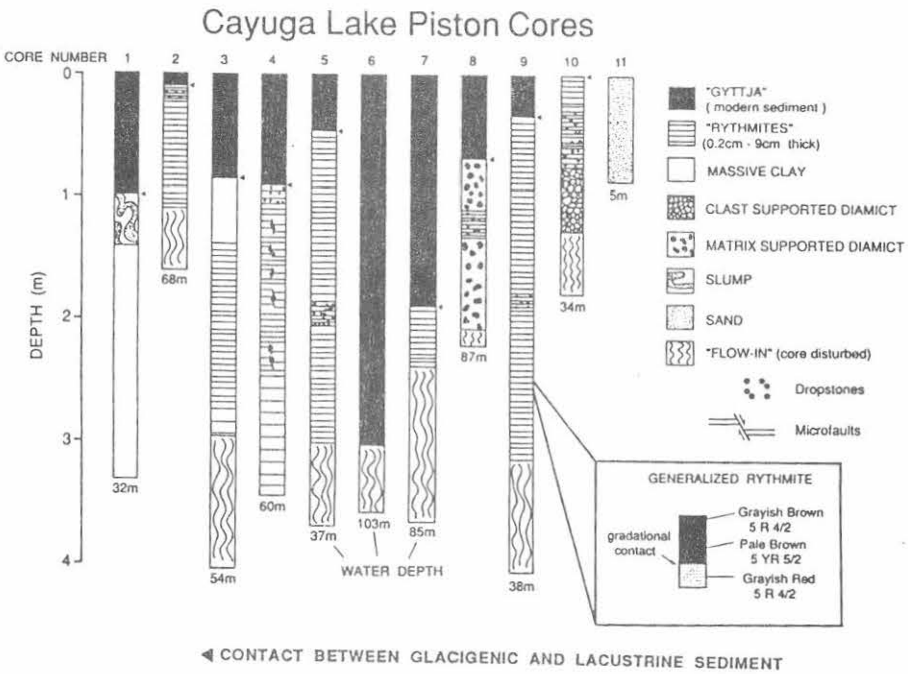


Fig. 20 - Schematic description of piston cores recovered from northern Cayuga Lake; see Figure 15 for locations.

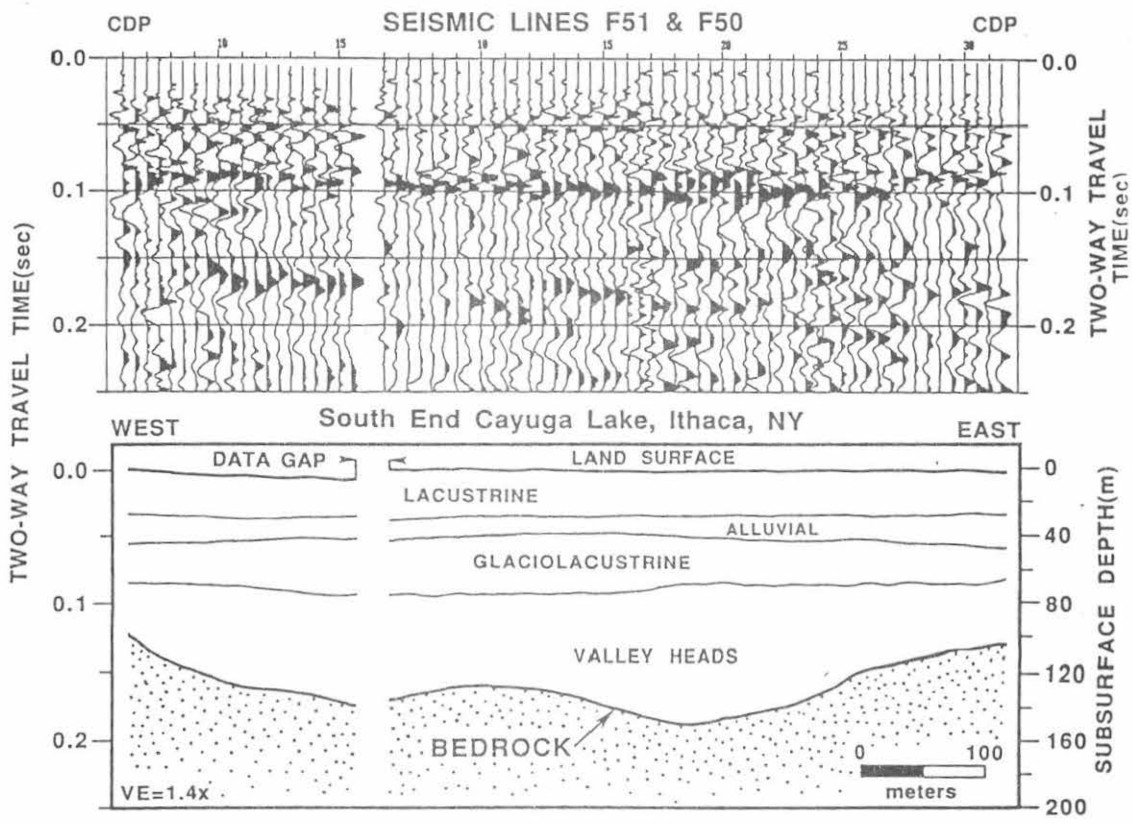


Fig. 21 - Weight-drop reflection profile (top) and line-drawing interpretation (bottom) across Cayuga valley south of Ithaca across from Buttermilk Falls State Park. Numbers across top indicate CDP shot points, subsurface depth scale assumes a velocity of 1.6 km/sec.

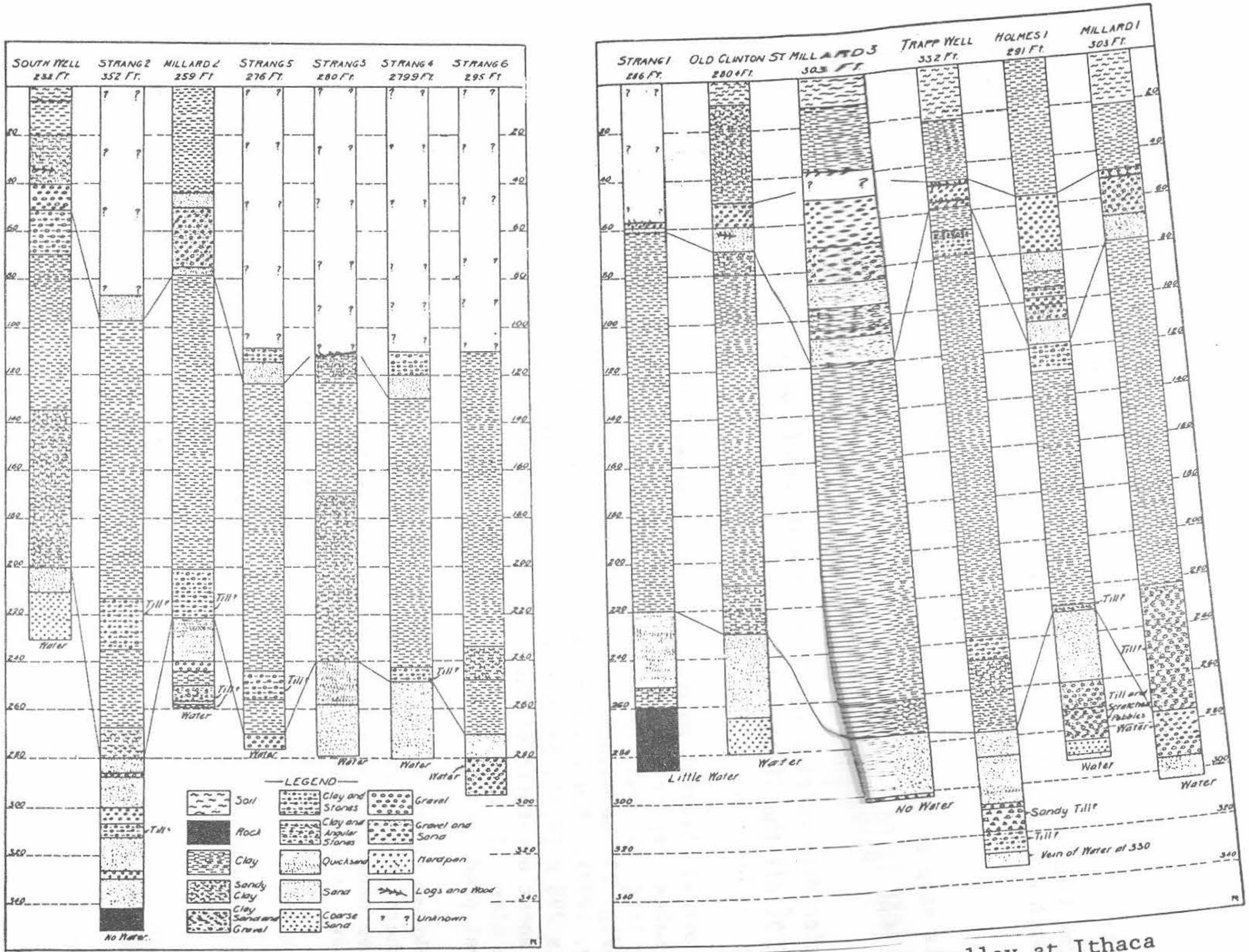


Fig. 22 - Schematic descriptions of well records from Cayuga valley at Ithaca arranged along north-south (left) and east-west (right) transects. Taken from Tarr (1904).

are likely a post-glacial lacustrine facies that records the flooding of the south end of Cayuga valley in response to differential isostatic rebound at the north end of the lake (Tarr, 1904). Since this flooding event, only a thin (8-12 m; 26-39') unit of organic-rich muds has accumulated in Cayuga Lake (seismic sequence VI).

Based on these correlations between our lacustrine seismic reflection data, land-based profile, and Tarr's well data, we have been able to "groundtruth" four of the six seismic stratigraphic sequences beneath Cayuga Lake (sequences I, IV, V, and VI). Sequences II and III pinchout at the south end of Cayuga Lake and do not extend to Tarr's wells.

From Ithaca we will head west across the upland divide to Watkins Glen at the south end of Seneca Lake.

STOP 7: SENECA LAKE

This stop at Warren Clute Memorial (Lakeside) Park near the south end of Seneca Lake is designed to examine both our lake-based and land-based seismic reflection data coupled with available drillhole data. Seneca Lake is the deepest of all the Finger Lakes having a maximum water depth of 186 m (610') (in the southern half of the lake) which is 50 m (165') below sea level.

The overall bedrock morphology beneath Seneca Lake is similar to the other Finger Lakes: spoon-shaped and deepening to the south in longitudinal (N-S) profile (Fig. 7); and, in transverse profile (E-W), broad and shallow in the north (Fig. 23) becoming more deeply incised to the south (Fig. 8; Mullins and Hinchey, 1989). Bedrock extends as much as 434 m (1423') below lake level which is 298 m (978') below sea level! This maximum depth of bedrock erosion occurs about one-third of the distance to the north end of the lake from Watkins Glen. It is only slightly deeper than the 257 m (842') depth below sea level predicted by Fairchild (1934b).

Thickness of the sediment-fill beneath Seneca Lake increases from north to south where it reaches a maximum of 270 m (886'). Seismic stratigraphic sequences beneath Seneca Lake are similar to those beneath Cayuga Lake except that there is a thicker transparent unit (sequence III). However, stratigraphic sequences could not be traced completely to the south end of the lake due to a highly irregular lake floor near the south end.

Similar to Owasco Lake an apparent esker is buried in the subsurface beneath Seneca Lake. A ridge, with up to 35 m (115') of relief, occurs at the base of the axial thalweg of northern Seneca Lake (Fig. 24). This ridge can be traced from line to line for several kilometers. We interpret this sinuous ridge as an esker which, if correct, further implies the presence of subglacial meltwater conduits during early stages of erosion and infill of the Finger Lakes.

A "half-valley" (east side) multichannel profile collected on-land between Watkins Glen and Montour Falls, further documents large scale erosion and thick sediment fill beneath Seneca valley (Fig. 25). The bedrock reflection here is quite distinct and extends at least 0.28 seconds of two-way travel time into the subsurface. Again, using a range of P-wave velocities of 1.5-2.0 km/sec., bedrock extends as much as 210-280 m (689-918') beneath the

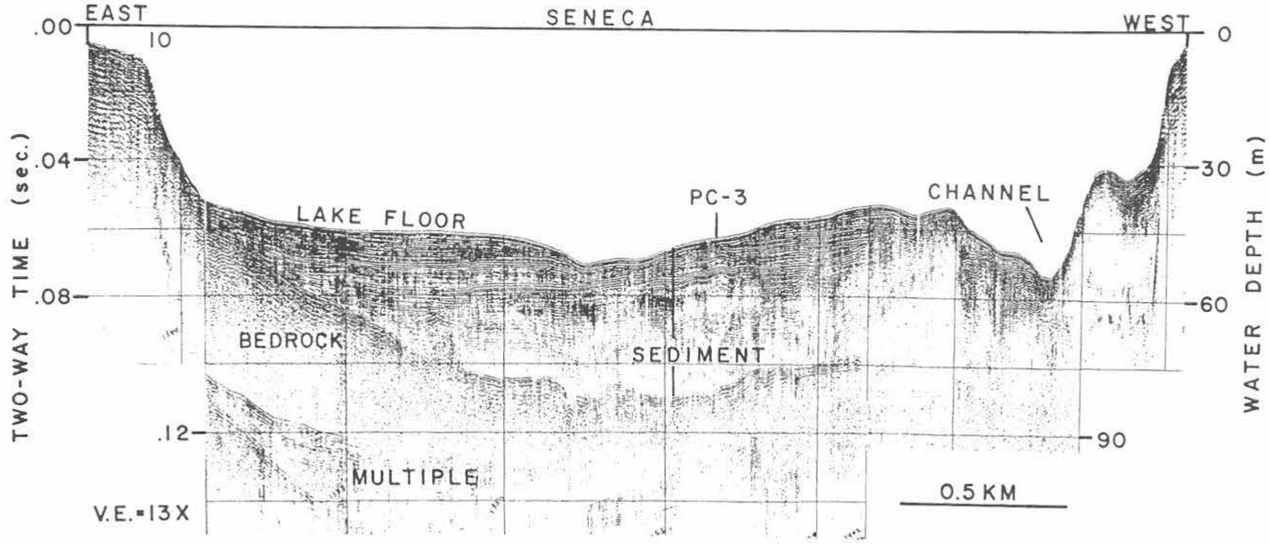


Fig. 23 - Transverse reflection profile from northern Seneca Lake illustrating broad U-shaped bedrock surface and thin (~30 m; 98′) sediment-fill. Piston core (PC)-3 recovered glaciolacustrine rhythmites.

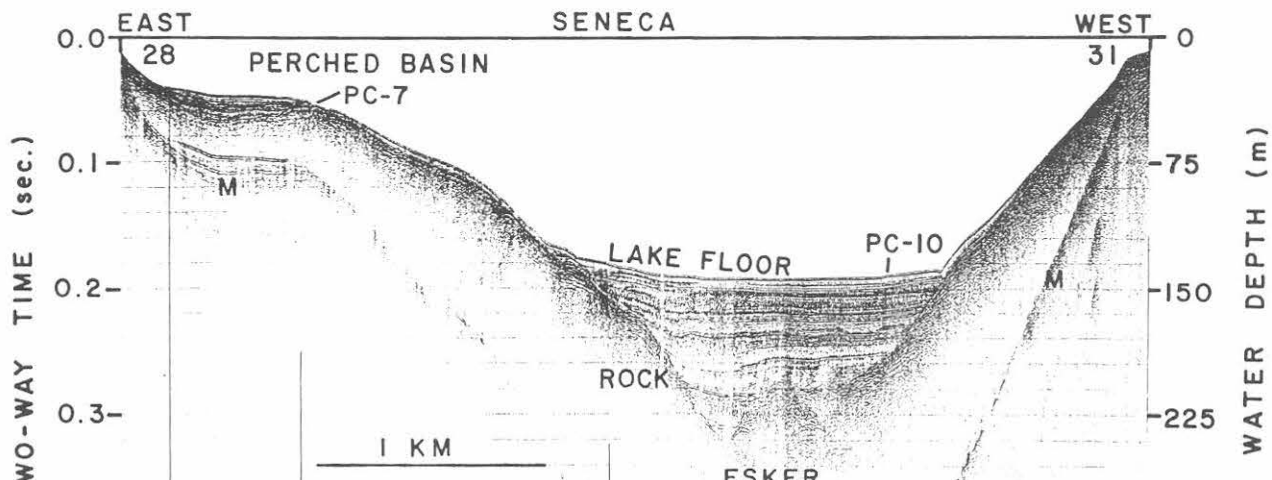


Fig. 24 - Transverse seismic reflection profile from central Seneca Lake illustrating asymmetric bedrock surface, ~90 (295′) of sediment-fill, and buried ridge in axial thalweg interpreted as an esker; M = multiple.

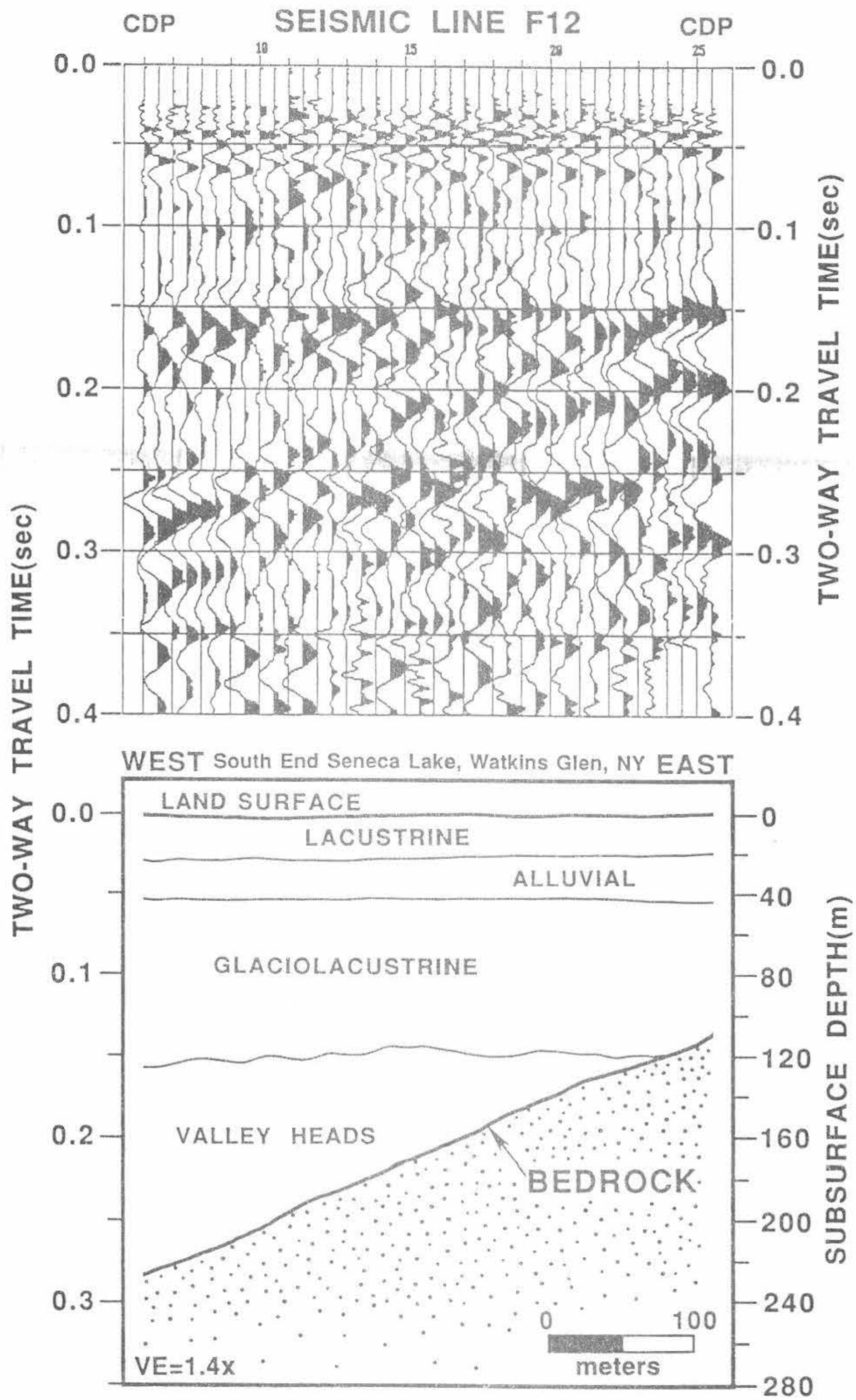


Fig. 25 - Weight-drop, multichannel reflection profile collected from the eastern half of dry valley south of Seneca Lake between Watkins Glen and Montour Falls (top) and line-drawing interpretation (bottom). Note distinct bedrock and Valley Heads reflections as well as three overlying stratigraphic units. Subsurface depth scale assumes a velocity of 1.6 km/sec.

valley floor. The higher end of these possibilities is very compatible with a well drilled at the power station at the south end of Seneca Lake (just to the west of our stop at Lakeside Park) which reached bedrock at 281 m (922') (Wendy McPherson, USGS Ithaca, personal communication, 1991). Thus, bedrock at the south end of Seneca Lake extends only about 145 m (476') below sea level compared to 298 m (978') below sea level further north, indicating that bedrock is rising to the south.

Our on-land profile reveals four major stratigraphic units above bedrock: (1) an upper transparent sequence about 20 m (66') thick; (2) a unit characterized by high-frequency reflections about 23 m (76') thick; (3) a largely transparent sequence with discontinuous high-amplitude reflections about 81 m (265') thick; and, (4) a basal, seismically chaotic high-amplitude sequence that is about 108 m (354') thick. By analogy with Tarr's (1904) well data from Ithaca as well as our own drill records south of Canandaigua Lake (stop #10) we interpret the basal chaotic sequence as Valley Heads equivalent; the overlying transparent unit as proglacial lake clays; the overlying high-frequency unit as fluvial/alluvial sands and gravels deposited as glacial lake levels dropped; and, the upper transparent unit as post-glacial lake muds deposited when the south end of the valley was flooded by isostatic rebound to the north. Again, it is our interpretation that there is only a single large infill sequence beneath the Finger Lakes.

From Watkins Glen we will head south to Horseheads passing through one of the largest Valley Heads deposits in the Finger Lakes region. Outcrops indicate that the Valley Heads here consist of water-laid, stratified sands and gravel. Near Horseheads the hummocky topography of the Valley Heads will give way to a more smooth outwash plain. From Horseheads we will drive west-northwest through Corning and then on to Bath where we will spend the evening. If it is still light, this drive will take us along a broad continuous valley along the Southern Tier Expressway which likely served as a channel for meltwater flow from the western Finger Lakes into the Susquehanna drainage system.

STOP 8: KEUKA LAKE (DAY TWO)

From our overnight stay in Bath, N.Y. we will head north to Hammondsport and then up along the west shore of Keuka Lake to our next stop at the north end of the west branch of Keuka Lake. Bath is located on an outwash plain just south of Valley Heads fill. Soon after leaving Bath you will be able to see hummocky terrain typical of the Valley Heads. We will then drop down to the dry lake floor on which the village of Hammondsport is located.

Keuka Lake is an anomalous Finger Lake in many regards. Most obvious is its two branches that converge to the south at Bluff Point (Figs. 1 and 2) which strongly suggests a southward directed, preglacial drainage system. Second, there are a number of circular to elliptical closed bathymetric depressions on the floor of both branches of Keuka Lake. Third, there is little north-south variation in the depth to bedrock beneath the northwest branch of the lake (Figs. 7 and 8). And, fourth, Keuka Lake valley is the only non-through valley of the Finger Lakes.

Maximum water depth beneath Keuka Lake is 58 m (190') which occurs in a closed depression near the junction of the Lake's two branches. Bedrock has been eroded as much as 200 m (656') below lake level which is 18 m (59') above sea level. Total sediment thickness is rather uniform beneath the west branch with a maximum of about 160 m (525').

Our axial (longitudinal, N-S) reflection profiles clearly define the east branch of Keuka Lake as a hanging tributary valley to the west branch (Fig. 26). Depth to bedrock beneath the east branch is significantly shallower and there is a distinct bedrock high or "sill" where the east branch joins the more deeply scoured west branch.

Axial profiles from the northwest branch of Keuka Lake indicate that the closed bathymetric depressions on the lake floor are remnants of ice-meltout features (Fig. 27). Reflectors beneath these depressions are displaced downwards suggesting post-depositional collapse and chaotic patterns suggest mass wasting (Fig. 27). These meltout features may be analogous to "dead-ice sinks" described by Fleisher (1986) who noticed that they occur preferentially in non-through valleys. He suggested that stagnant ice conditions develop in non-through valleys because of glacial thinning and detachment at headward divides which are not present in through valleys (Fleisher, 1986). The discovery of ice-meltout features beneath Keuka Lake suggests a component of downwasting here during deglaciation unlike the other Finger Lakes where ice retreat appears to have been more rapid, perhaps by calving.

Despite this evidence for downwasting, the seismic stratigraphy beneath Keuka Lake at its southern end is typical of the other Finger Lakes (Fig. 28). A lower chaotic unit (sequence II) is overlain by a very transparent (reflection-free) sequence (III) which in turn is overlain by a highly reflective unit (IV) derived from the north. Sequence V, which is also highly reflective, thickens to the south and contains southerly-derived mass-movement deposits; and, sequence VI is a relatively low-amplitude facies that probably represents contemporary (post-glacial) lacustrine sedimentation.

From our overlook of the northwest branch of Keuka Lake we will drive west through Italy Valley to the village of Naples passing over Valley Heads fill and then to the south end of Canandaigua Lake.

STOP 9: CANANDAIGUA LAKE

This overlook stop provides a spectacular north-view of Canandaigua Lake -- "the chosen place." Bathymetric data reveal a symmetrical "tub-shaped" lake basin with a maximum water depth of 84 m (276') near its north-south midpoint. Bedrock has been eroded as much as 261 m (856') below lake level which is 51 m (168') below sea-level (Figs. 7 and 8). Total erosion, from the top of the adjacent divides to the top bedrock beneath the lake, has been on the order of 630 m (2068'). An on-land seismic reflection profile across Valley Heads along Eel Pot Road south of Naples, suggests that although bedrock does rise to the south there may be as much as 250 m (820') of Valley Heads fill (Fig. 29).

Maximum sediment thickness beneath Canandaigua Lake is 202 m (663') which occurs in the southern half of the lake basin. Similar to the other Finger

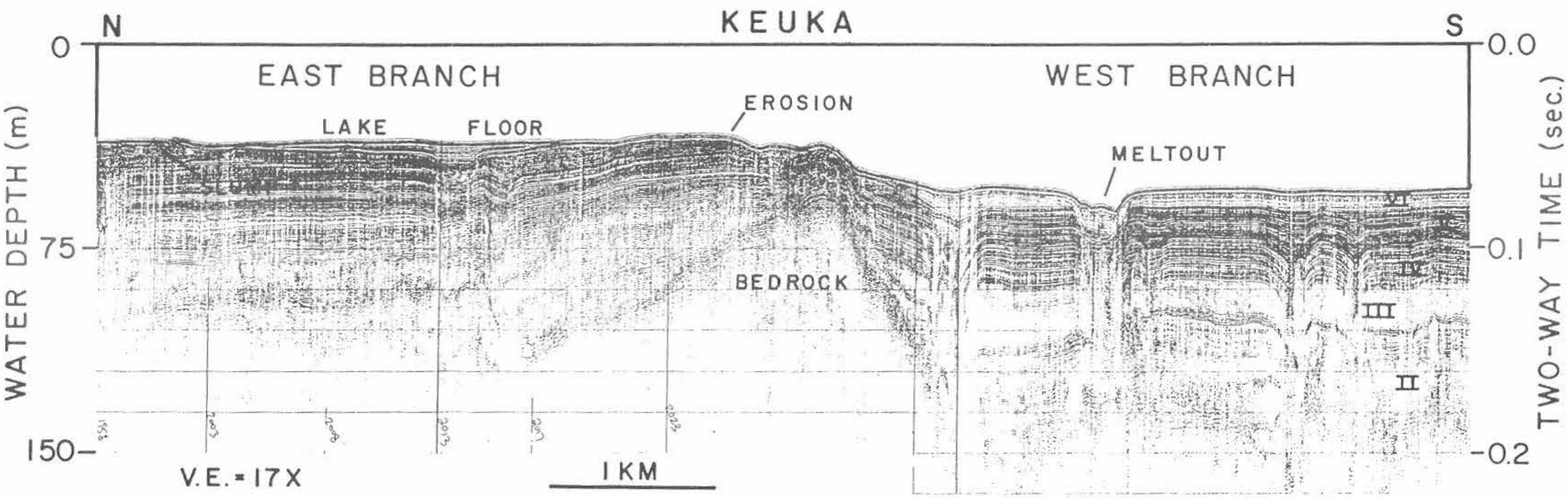


Fig. 26 - Portion of axial reflection profile from Keuka Lake illustrating bedrock "high" (sill) at juncture of east (hanging valley) and west branches near Bluff Point.

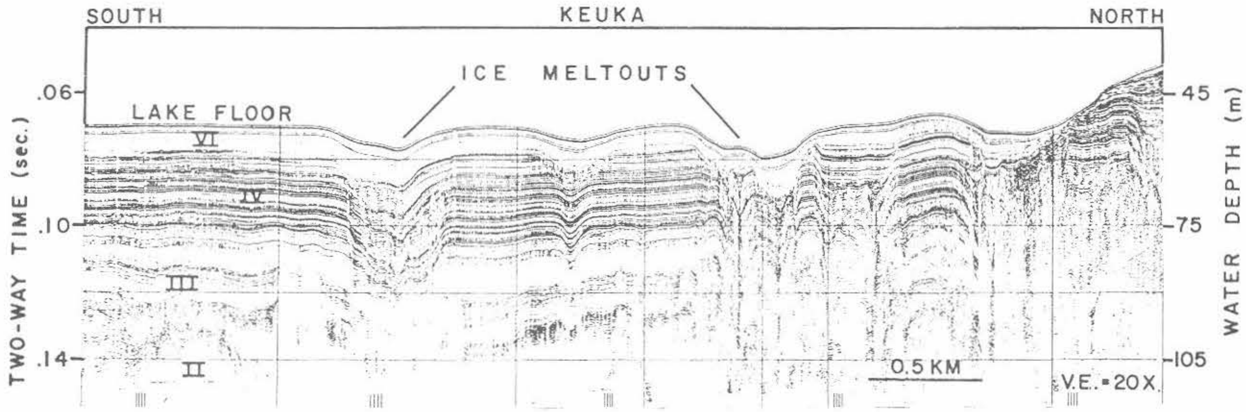


Fig. 27 - Northern end of axial reflection profile from the west branch of Keuka Lake illustrating ice-meltout features ("dead-ice sinks") that produce closed bathymetric depressions on lake floor.

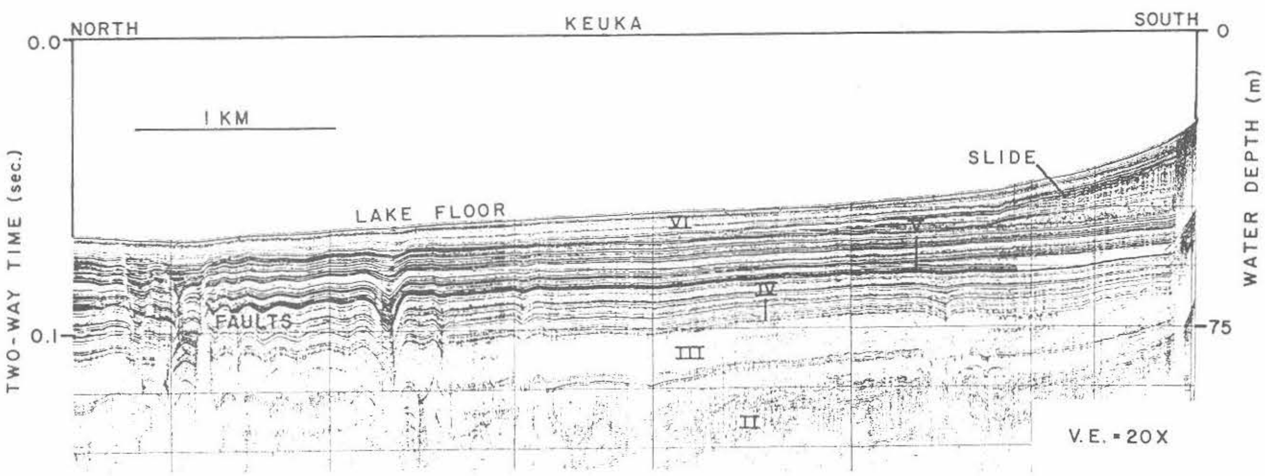


Fig. 28 - Portion of axial reflection profile from the southern end of Keuka Lake near Hammondsport illustrating seismically defined depositional sequences. Note mass flow deposits to south, and small subsidence faults to north.

Lakes we recognize six seismically-defined depositional sequences beneath Canandaigua Lake; however, there are some significant differences. Most notable is sequence III -- a transparent (reflection-free) facies -- which on our axial line (Fig. 30) occupies a restricted, "scooped-out" depression near the center of the lake. These massive fine-grained (?) sediments are up to 60 m (197') thick, and are overlain by highly reflective sequences which onlap bedrock valley walls (Fig. 31).

Another difference is that collectively, sequences IV-VI (highly reflective facies) thicken to the south implying a largely southerly derived source for these deposits. We have noticed a similar relationship in other Finger Lakes only for sequences V and VI which we have related to a lowering of glacial lake levels and a drainage reversal which accompanied complete ice withdrawal from the Finger Lake valleys. In the case of Canandaigua Lake, these data suggest that sediment input from the south end of the lake began sooner than in the other lakes. This may have been due to the fact that there are multiple valleys, which may have been ice-free at the time, that drain into the south end of Canandaigua Lake.

All seismic stratigraphic sequences beneath Canandaigua Lake, with the exception of sequence III, can be traced to the south end of the lake. These sequences do <u>not</u> pinchout here but appear to continue beneath the wetland and dry lake valley to the south. In fact, this is why we decided to locate our drillcore to the south Canandaigua Lake. Figure 29 schematically illustrates the projection of seismic stratigraphic sequences beneath Canandaigua Lake to our drillsite ~3 km to the south along Parrish Flat Road.

STOP 10: PARRISH FLAT ROAD (DRILLCORE RESULTS)

Parrish Flat Road, which cuts across the dry valley just south of Canandaigua Lake, provides a unique accessible location to correlate (as directly as possible) our lake-based seismic reflection data with drillcore samples. Between the drillsite and Canandaigua Lake (3 km to the north) are the High Tor wetlands; directly to the east is a large alluvial fan at the mouth of Conklin Gully; and 3 km to the south, Valley Heads deposits begin to crop out near the village of Naples.

A multichannel reflection profile collected along Parrish Flat Road in 1988 indicated that bedrock extends as much as 0.22 seconds of two-way travel time beneath the valley floor (Fig. 32). Using an interval velocity of 1.5 km/sec., there is about 165 m (540') of sediment fill above bedrock here. This profile also suggested the presence of four stratigraphic units: (1) a thin, upper transparent unit about 11 m (37') thick, underlain by (2) a thin (~10 m, 33' thick) high frequency sequence that thickens to east toward the alluvial fan; (3) a thick (~94 m, 308') unit with discontinuous high-amplitude reflections; and, (4) a basal unit, about 50 m (165') thick, beneath an irregular, high-amplitude reflector (Fig. 32).

Drillcore samples were recovered in July 1990 at two sites along Parrish Flat Road: site 1 extended 120 m (400') into the subsurface; and site 2, located 168 m (550') to the west, extended only 12 m (40') into the subsurface (Fig. 33). Although composite core recovery was only ~40%, vertical stratigraphic continuity was extended with the use of downhole geophysical

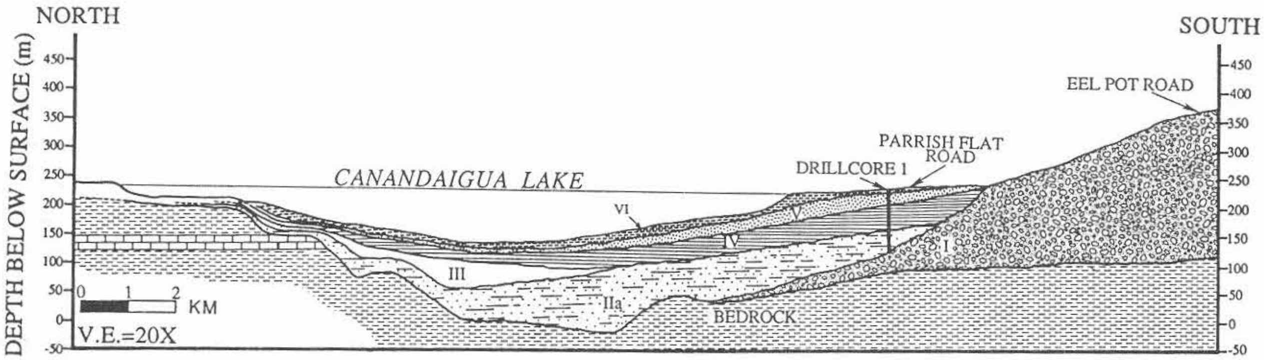


Fig. 29 - Schematic axial (N-S) profile of bedrock and stratigraphy of sediment-fill beneath Canandaigua Lake and dry valley to south. Note depositional sequences and location of drillcore. Vertical scale relative to sea-level; horizontal scale in km.

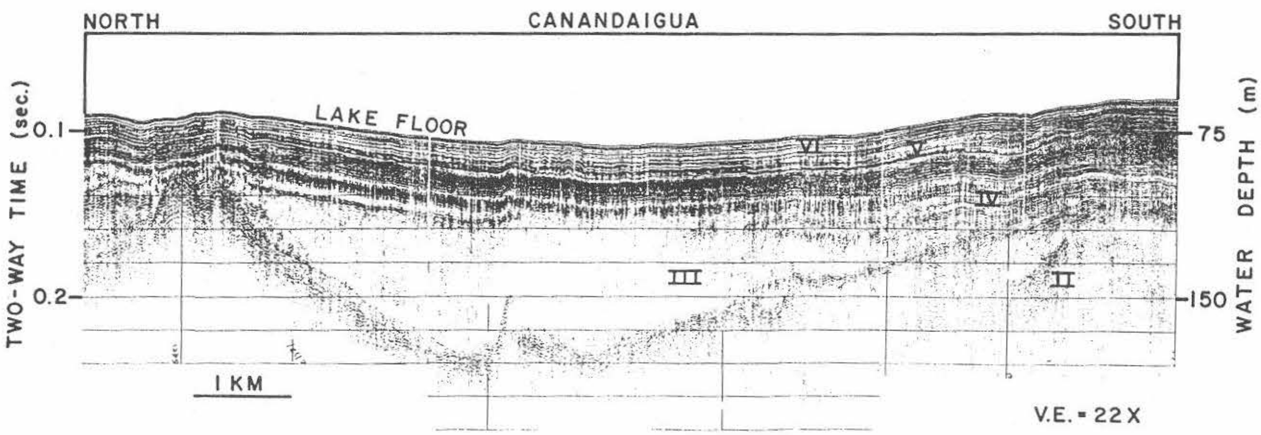


Fig. 30 - Central portion of axial reflection profile from Canandaigua Lake illustrating seismically transparent sequence III.

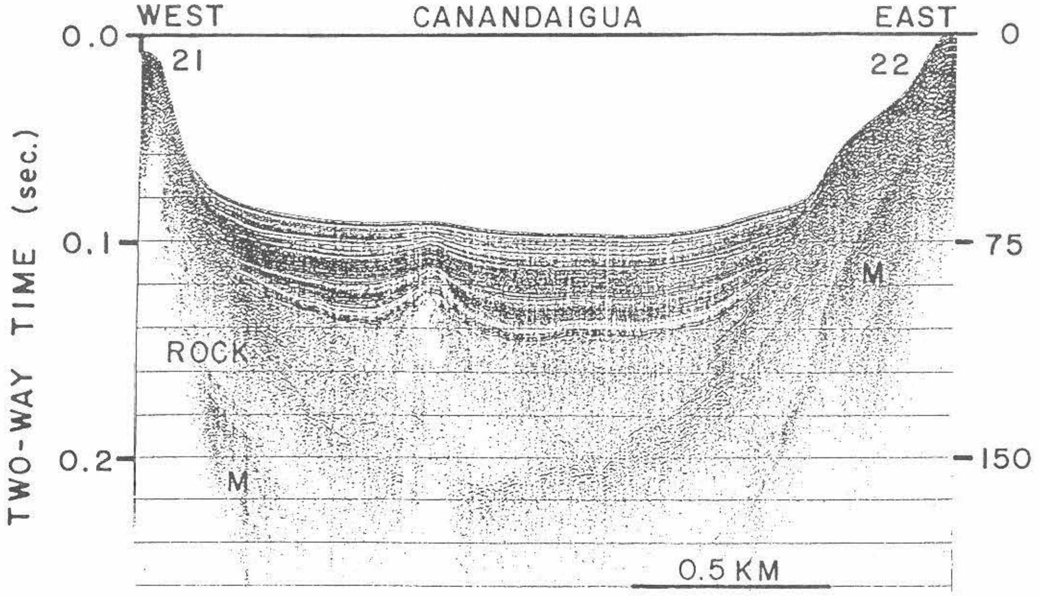


Fig. 31 - Transverse reflection profile from the central portion of Canandaigua Lake

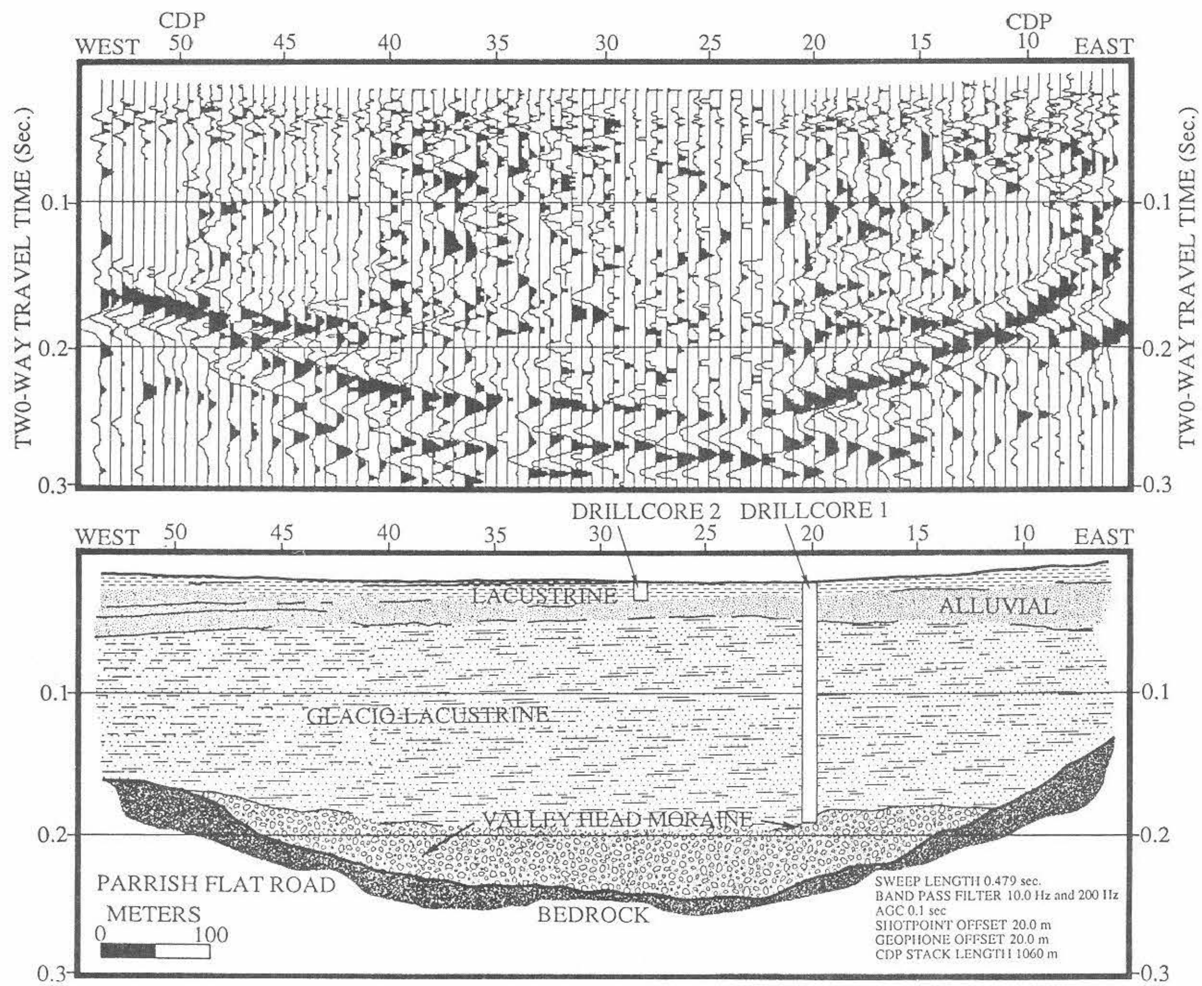


Fig. 32 - Weight-drop, multichannel reflection profile (top) and interpretation based on drillcores (bottom) along Parrish Flat Road (gamma ray and resistivity) logs (Fig. 33). Data from a vertical seismic profile shot at the deeper drillhole indicate that P-wave velocities in the upper 100 m (328') of section are all less than 1.6 km/sec. (Fig. 34) indicating undercompacted materials.

Based on physical stratigraphy, we recognize five distinct stratigraphic sequences in the Canandaigua drillcore: (1) an upper unit (12 m, 39' thick) of cyclic sequences of peat and lake clays; (2) a coarsening-upward, washed sand and gravel unit (containing artesian water) that is 14 m (46') thick; (3) a rhythmically bedded silt and sand sequence plus interbeds of massive clay with dropstones that is 60 m (197') thick; (4) a massive clay unit devoid of dropstones that is 23 m (75') thick; and (5) and basal coarse sand and gravel unit that would not support open-hole drilling (Fig. 33). Although bedrock was not reached in this drillcore, a gas well located in the center of the valley about 3 km to the south, did encounter bedrock at a subsurface depth of 143 m (470'). The age of these deposits has not yet been determined; however, ten peat samples from the upper 12 m (39') have been sent off for radiocarbon analysis and one sample of massive clay from 109 (358') has been sent off for thermoluminescence dating. Hopefully results will be in hand by the time of the field trip.

Our stratigraphic results from the Canandaigua drillcore are strikingly similar to those reported by Tarr (1904) from Ithaca (Cayuga Lake) as well as available drillcore data from Tully Valley (Fig. 6). Basically, a basal sand and gravel unit is overlain by a thick sequence of fine-grained, organic-poor sediment which in turn is capped by sands and gravels that grade up into organic-rich lake clays. We interpret this stratigraphy as a single infill sequence resulting from the last glaciation. The basal sands and gravels are interpreted as part of the Valley Heads fill (13-14 ka); the thick, organic-lean clay sequence is interpreted as glaciolacustrine facies; the upper sands and gravels as prograding alluvial valley fill that formed as glacial lake levels dropped; and, the uppermost organic-rich sequence as post-glacial swamp and lacustrine deposits. The cyclic nature of these youngest sequences suggest that lake levels have fluctuated periodically in Canandaigua Lake, and perhaps the other Finger Lakes, during postglacial time. Much more detailed study of these upper cyclic sequences is currently underway as part of Rob Wellner's doctoral dissertation.

From Parrish Flat Road we will drive north along the western margin of Canandaigua Lake to the city of Canandaigua. Toward the north end of the lake you will begin to notice drumlins which will be very common as we drive east along the New York State Thruway to our next stop at Montezuma wetlands north of Cayuga Lake.

STOP 11: MONTEZUMA CHANNELS

Montezuma National Wildlife Refuge is one of the largest freshwater wetlands in New York State. The wetlands actually occupy a southward directed, dendritic system of channels that funnel into the north end of Cayuga Lake, which has the lowest lake level (116 m, 382'). That these channels are relict, is evidenced by the fact that modern outlet drainage from Cayuga Lake is to the north.

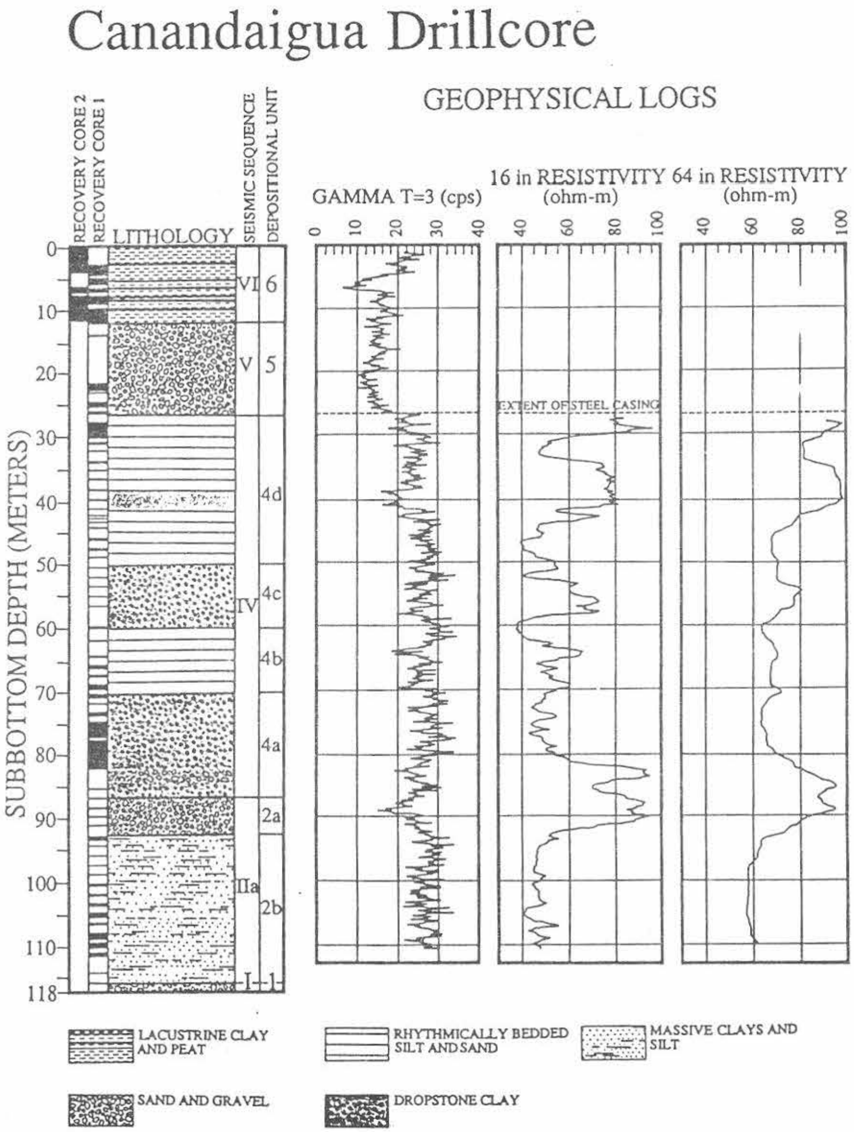


Fig. 33 - Schematic illustration of lithostratigraphy recovered at Canandaigua drillcores (left) and downhole geophysical logs (right).

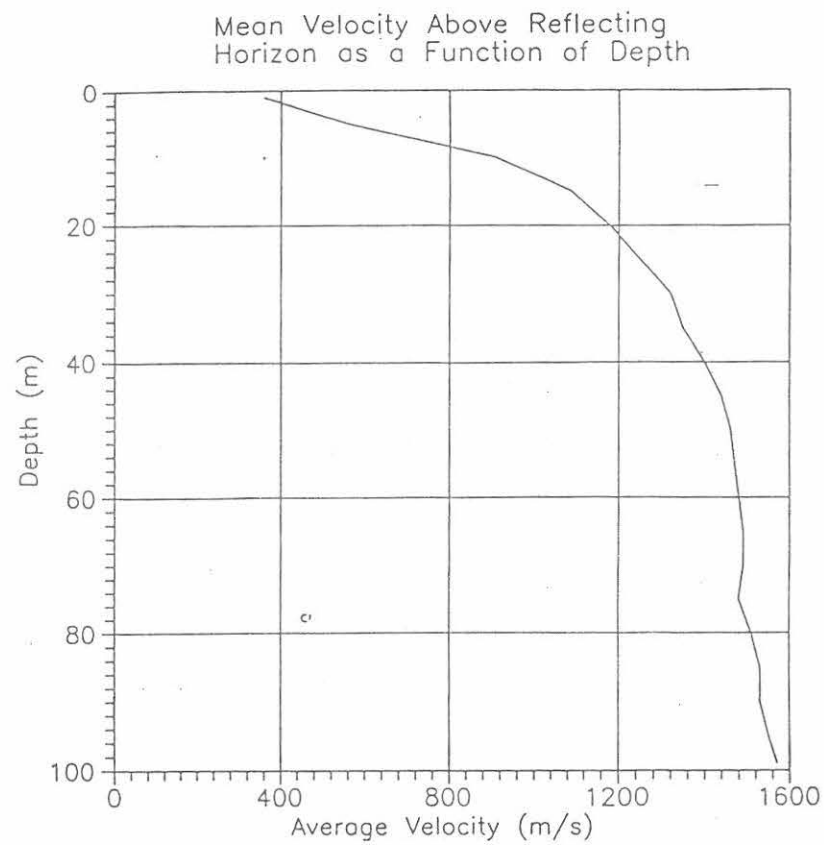


Fig. 34 - Curve of compressional wave velocity versus depth for Canandaigua drillsite along Parrish Flat Road. Data based on vertical seismic profile (VSP) shot down drillhole.

The Montezuma channel system is unusually broad (up to 5 km) for having a north-south length of only ~25 km. The channels are quite distinct on satellite images of the Finger Lakes and have been mapped largely as swamp deposits by Muller and Cadwell (1986). The channels occur within a very well-developed drumlin field but are neither drumlinized nor do they appear to truncate drumlins. The geologic significance of the Montezuma channels has been largely overlooked in previous regional analysis.

In order to better understand the subsurface geology of the Montezuma channels, we have collected about 12 km of on-land, multichannel seismic reflection profiles that have been integrated with available well data. A profile acquired at our stop along Armitage Road reveals a thalweg along the western edge of this eastern branch of the channel system (Fig. 35). Bedrock here extends as much as 40 m (131') beneath the surface of the channel or 78 m (257') above sea level. Maximum depth to bedrock beneath the channels, based on well records, is 55 m (180').

Two well records are available from Armitage Road. One well along the eastern margin of the channel penetrated about 30 m (98') of clay on top of shale before bottoming in limestone at 38 m (125'). The second well, located in the channel's thalweg, bottomed in sand without reaching bedrock (Fig. 35). Facies analysis of our reflection data suggest that the sands are restricted to the channel's thalweg (Fig. 35).

The postglacial pollen record of the Montezuma channels has been investigated via a 12 m core recovered from Crusoe Lake located about 8 km north of Armitage Road (Cox and Lewis, 1965). This core retrieved about 6 m (20') of organic-rich gytta overlying 5 m (16') of clay (Fig. 36). Two radiocarbon dates from 5.0 m (16') and 1.3 m (4') below the lake floor yielded ages of 6,850 ±150 and 3,200 ±100 years, respectively (Cox and Lewis, 1965). Pollen data (Fig. 36) indicate that a spruce-pine-fir forest dominated the region prior to 6,850 years ago. The first deciduous forests (dominated by hemlock and oak) developed about 6,500 years ago during a time when the climate here may have been warmer and more moist than today (Cox and Lewis, 1965). Between about 3,000 and 2,000 years ago, beech replaced much of the hemlock suggesting a decrease in available moisture. During the past 2,000 years, the pollen record (Fig. 36) suggests a decrease in temperature and an increase of available moisture (Cox and Lewis, 1965). These authors also suggest that a large lake, of which Crusoe is a remnant, may have persisted in Montezuma Marsh until about 700 A.D.

We also collected a "sledge-hammer" multichannel seismic reflection profile across the drumlin we are standing on (Fig. 37). Results suggest that there are about 18 m (58') of unconsolidated sediment beneath the crest of the drumlin. The bedrock reflection is quite distinct and indicates a "high" with about 6 m (18') of relief. Although these results are from only one of thousands of drumlins in the area, they do suggest that the drumlins here may be controlled by relief on an eroded bedrock surface.

What is the origin of the Montezuma channels? The facts are that it is a southward-directed, dendritic system with a maximum length to width ratio of about 5:1; it is a broad, shallow system filled with up to 55 m (180') of

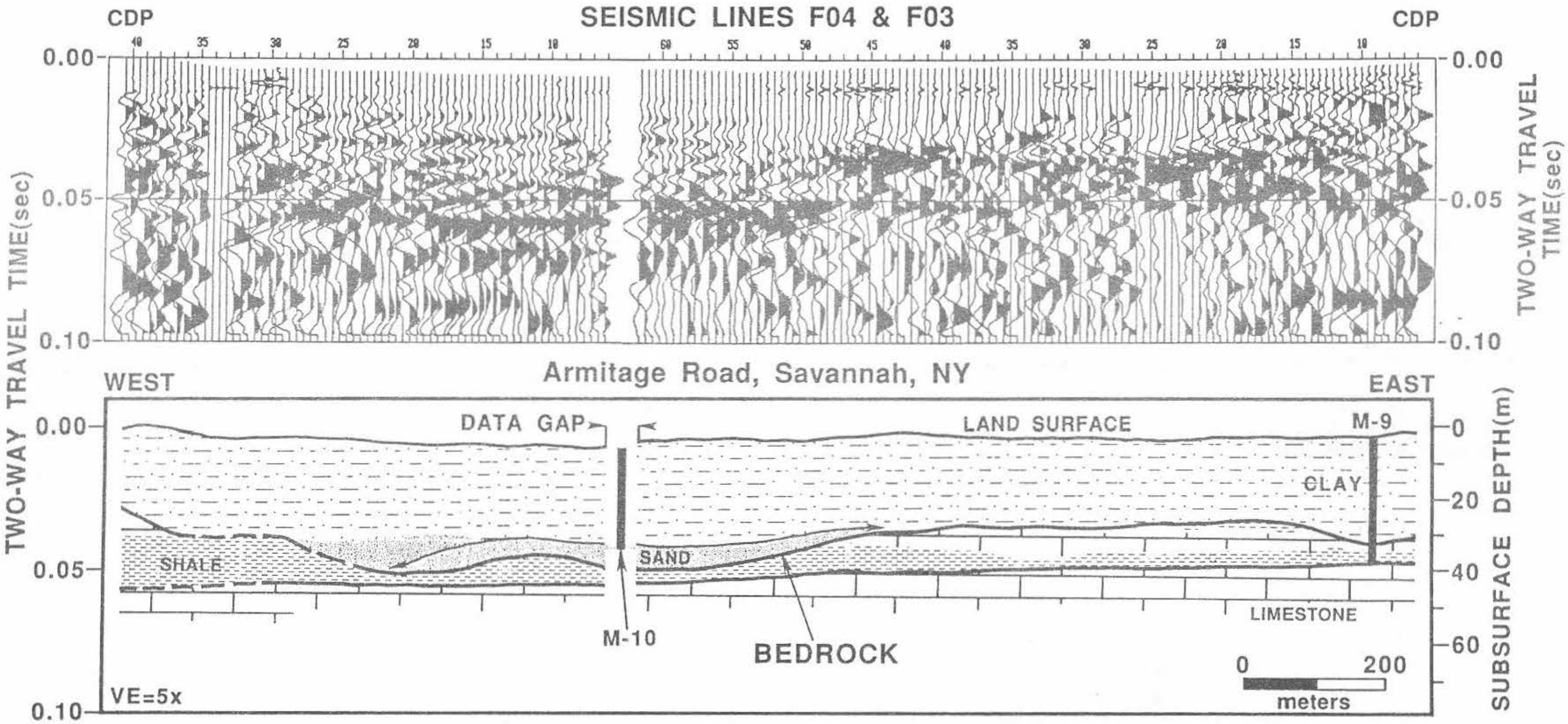


Fig. 35 - Weight-drop, multichannel reflection profile (top) and line-drawing interpretation (bottom) of west branch of Montezuma channels along Armitage Road. Well records confirm subsurface lithologies. Note thalweg along western reach of channel.

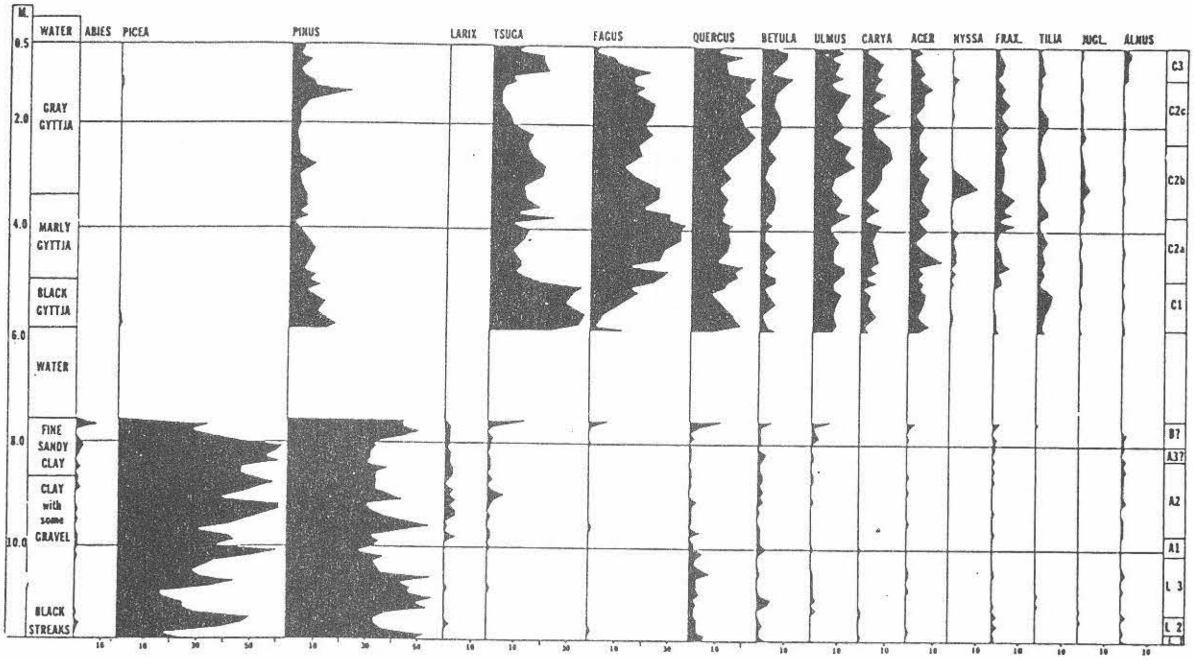


Fig. 36 - Pollen diagram for a core recovered from Crusoe Lake in the Montezuma channels. Gross stratigraphy at left; pollen zones at right; from Cox and Lewis (1965).

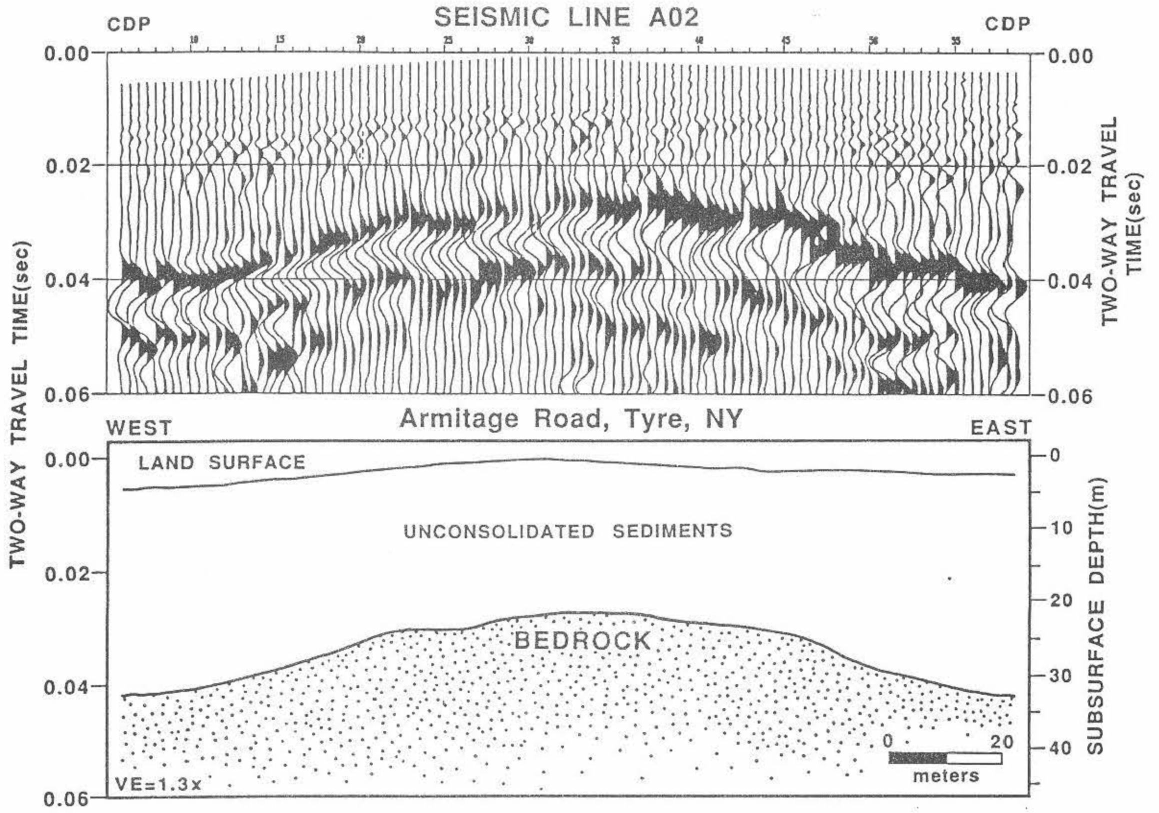


Fig. 37 - "Sledge-hammer" multichannel reflection profile across drumlin along east margin of west branch of Montezuma channels. Note bedrock "high" beneath drumlin. Depth scale assumes a P-wave velocity of 1.55 km/sec.

sediment; basal coarse-grained sediments are overlain by lake clays; and, the entire system feeds into the north end of Cayuga Lake despite the fact that contemporary outlet drainage is to the north.

One interpretation is that these channels represent subaerial proglacial drainage from a stabilized ice margin north of the channels. However, there is little direct evidence for a stabilized ice-margin here (Muller and Cadwell, 1986) and the dendritic pattern is inconsistent with proglacial streams which are typically braided. Also, a proglacial stream origin would require subaerial conditions followed by a sublacustrine environment.

An alternative hypothesis is that the Montezuma channels were carved by subglacial meltwaters that were funneled into the Cayuga Lake valley. A dendritic pattern has been theoretically predicted for subglacial drainage systems (Fig. 38) which Boulton and Hindmarsh (1987) argue are required to drain excess water during stable subglacial deformation to form drumlins. A subglacial origin for these channels would also allow for direct evolution of environments from subglacial fluvial to glaciolacustrine and then lacustrine (without an intervening subaerial stage) as the ice sheet retreated. Muller and Cadwell's (1986) surficial geologic map of the Finger Lakes (Fig. 2) indicates the presence of a similar, but less well-developed, channel at the north end of Seneca Lake which has an esker-like ridge along its northern axial thalweg.

From our stop on Armitage Road along the western branch of the Montezuma channels, we will head east and drive across the eastern branch of the channels where you can get a good "feel" for the width of these channels. We will then drive south to Route 20 and then east to our last stop at Skaneateles which will take us across the southern edge of the Weedsport drumlin field.

STOP 12: NORTH END OF SKANEATELES LAKE (WRAP-UP DISCUSSION)

This scenic stop at the north end of Skaneateles Lake is designed to provide us with an opportunity to collectively discuss the subsurface geophysical/geologic data we have examined during the past day and a half, and its implications for the origin and evolution of the Finger Lakes. These data document the large scale erosion (up to 298 m; 978' <u>below</u> sea-level) and infill (up to 270 m; 886') of the Finger Lake valleys (Mullins and Hinchey, 1989) including a thick Valley Heads sequence. Seismic stratigraphic relationships indicate that the sediment-fill beneath the Finger Lakes post-dates deposition of Valley Heads at 13-14 ka. Thus, timing of the erosion of the Finger Lakes can only loosely be constrained as sometime between the Devonian and 13-14 ka! If pre-late Wisconsin sediments were deposited beneath the Finger Lake valleys they have been removed by subsequent erosion. This implies that the Finger Lakes were at least deepened by late Wisconsin glaciation.

Hughes' (1987) model of Laurentide Ice Sheet deglaciation for 14 ka (Fig. 39) suggests that ice streamed into the Finger Lakes region from the St. Lawrence Lowlands/Ontario Basin as the ice sheet began to collapse. Based on our subsurface data integrated with Muller and Cadwell's (1986) surficial map, we suggest a highly digitate ice margin in the Finger Lakes region with streams of more rapidly flowing ice moving down the Finger Lake valleys

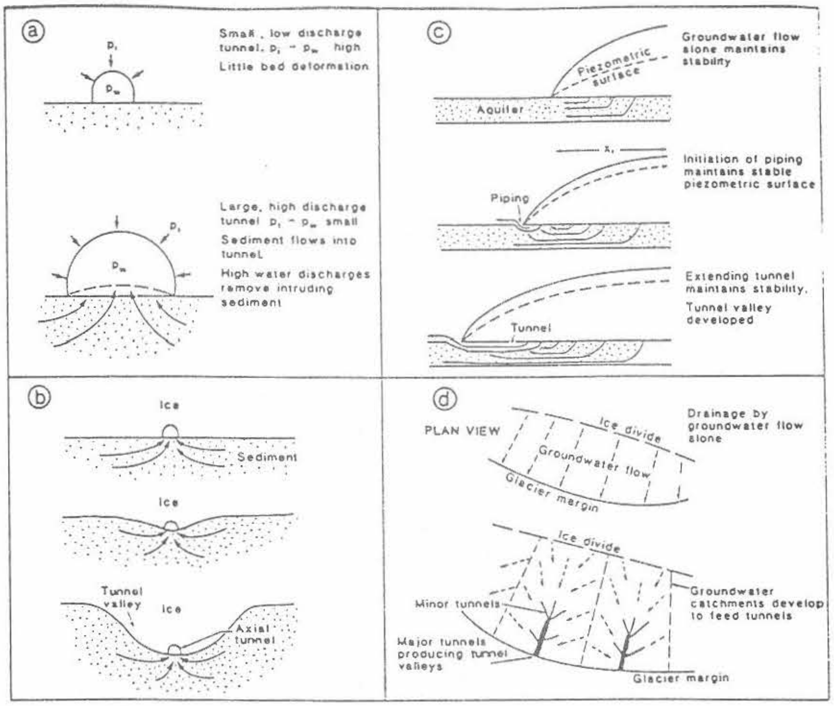


Fig. 38 - Schematic, conceptual models for discharge of subglacial meltwater and formation of tunnel valleys beneath temperate ice sheets. Predicted subglacial drainage system is illustrated in "D". From Bouldon and Hindmarch (1987).

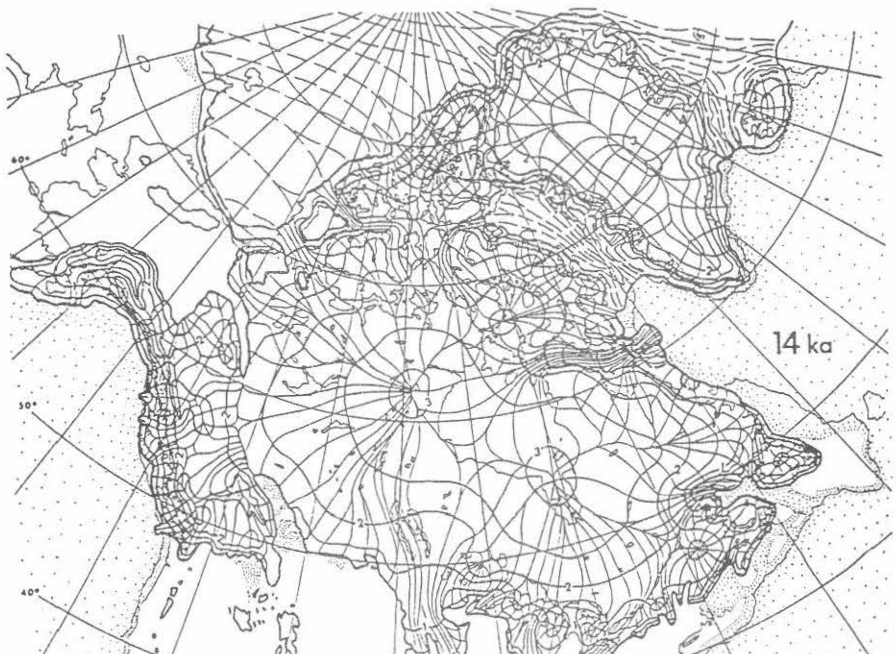


Fig. 39 - Reconstruction of Laurentide Ice Sheet at 14 ka (from Hughes, 1987). Note ice domes over James Bay and southeastern Canada with converging flow lines along St. Lawrence lowlands that extrude into Finger Lakes region.

(Fig. 40). At this time (13-14 ka) large volumes of coarse-grained debris were being pumped subglacially down through the Finger Lake valleys, perhaps by channelized subglacial meltwaters, and ultimately deposited as the Valley Heads moraines. An alternative interpretation for this stage is that an ice dome developed in the Lake Ontario basin and that rapid ice flow through the Finger Lake valleys was in the form of outlet glaciers which may have also been accompanied by channelized subglacial meltwater (Ridky and Bindshadler, 1990).

As the ice margin began to pull-off its Valley Heads position channelized subglacial meltwaters continued to pump large volumes of finer-grained debris into the Finger Lakes valleys as subaqueous outwash (sequence II; Fig. 41). In this regard we agree with Gustavson and Boothroyd (1987) that subglacial streams (rather than "dirt machines") were the primary source of glacio-fluvial and glaciolacustrine sediment along the southern margin of the Laurentide Ice Sheet.

Once ice margins had retreated to the north end of the Finger Lake valleys, classical proglacial lacustrine sedimentation continued to fill the valleys which is recorded on our reflection profiles as highly-reflective sequence IV. At this time, with the north ends of the valleys dammed by ice, lake levels were considerably higher than today. However, as soon as ice retreated from the north ends of the valleys, lake levels dropped rapidly and a drainage reversal (from south- to north-directed) occurred (Mullins and Hinchey, 1989). This event is recorded on our profiles by sequence V which thickens to the south in all the lake basins. When lake levels dropped, highstand deltas were left hanging on valley walls and numerous gorges were eroded as local base level was dramatically lowered (Fig. 41).

After the ice completely withdrew from the Finger Lakes region ~12 ka (Hughes, 1987), differential isostatic rebound in the north resulted in the flooding of the south ends of the Finger Lakes valleys as evidenced by lake clays overlying fluvial/alluvial sands and gravels in drill cores. Subsequent to this flooding event, post-glacial drainage from the south has partially filled the southern ends of the Finger Lake valleys, while accumulating a thin, transparent "blanket" of sediment beneath the lakes.

The most recent chapter in the natural history of the Finger Lakes is recorded in the cyclic sequence of peats and lake clays recovered in the Canandaigua drillcore. There are about six of these cycles that overlay fluvial/alluvial sands and gravels. Do these cycles represent climatically-driven lake level fluctuations or post-glacial rebound oscillations? Whatever their cause, the fact that there are about six cycle in the post-glacial (<~12 ka) section suggests that they have a periodicity on the order of 2 ka. Our working hypothesis is that they were climatically driven and may record an environmental record of wet and dry cycles for the Finger Lakes region during the past 12 ka. We continue to work on these cyclic sequences and should have more definitive results in the near future.

We hope that you have not only enjoyed this unique field trip but that you have also gained a better appreciation of the "unseen" geology beneath the Finger Lake valleys. If we are to fully comprehend the Quaternary geologic history of regions such as the Finger Lakes it will be necessary to record subsurface as well as surficial geology. We hope that our integrated

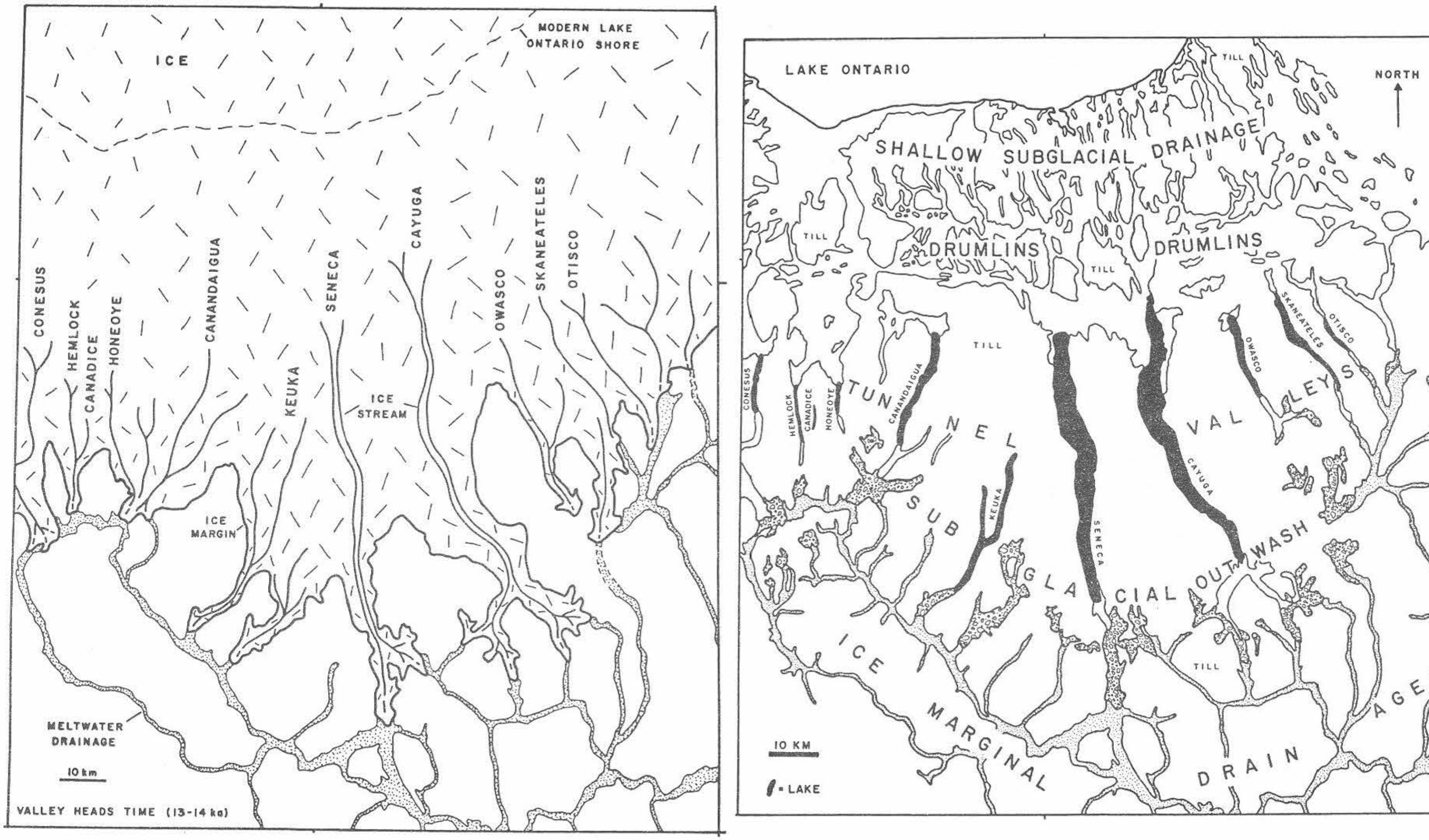


Fig. 40 - (Left) Schematic reconstruction of southern margin of Laurentide Ice Sheet in Finger Lakes region at Valley Heads time (13-14 ka). (Right) Regional interpretation of major glacial features in Finger Lakes region.

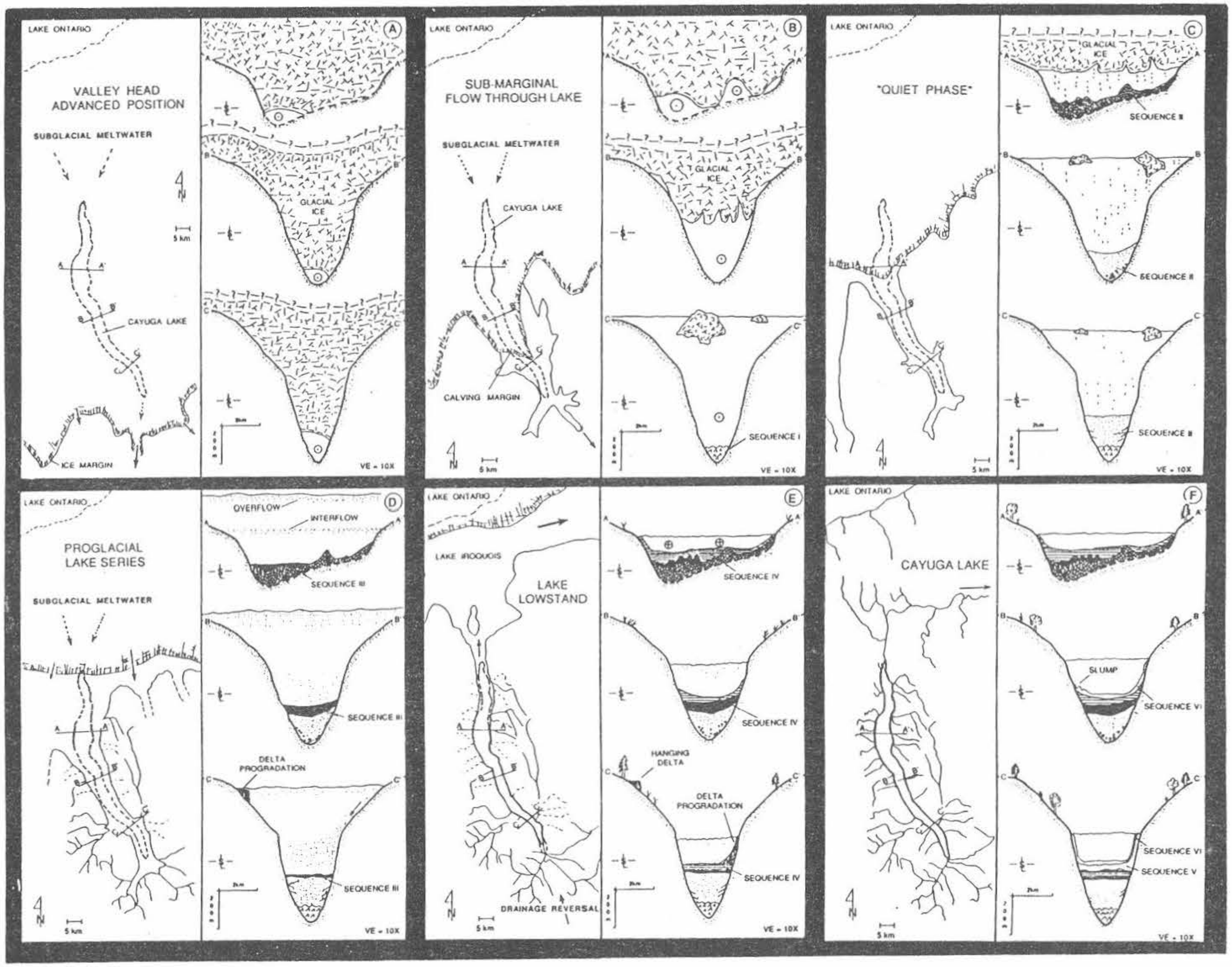


Fig. 41 - Schematic, sequential illustration of sediment infill history for the Finger Lakes based on seismic stratigraphic data from Cayuga Lake (E. Hinchey, in prep.).

subsurface geophysical and geological approach has revealed at least a part of the "barely suspected missing chapter in the history of the Finger Lakes" alluded to by Bloom (1984).

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ROAD LOG

SUBSURFACE GEOLOGY OF FINGER LAKES

Total Miles	Miles from Last Point	Route Description
0.0	0.0	Depart Heroy Geology Lab on the campus of Syracuse University. Turn left at light on to University Avenue.
0.2	0.2	Turn right at light on to Irving Avenue.
0.7	0.5	Turn left at Harrison Street.
0.9	0.2	Turn left on to I-81 south. Drive south along east margin of Onondaga Trough toward Cortland.
12.1	11.2	Exit 15 - Lafayette. Note jointed bedrock along off-ramp.
12.2	0.1	Turn left on to Route 20 west.
13.0	0.8	Turn right on to Webb Road.
13.5	0.5	Turn left on to Amidon Road.
14.1	0.6	<u>STOP #1</u> - dead end of Amidon Road - overview of Tully Valley. Turn around; return east on Amidon Road.
14.7	0.6	Turn right on Webb Road.
15.2	0.5	Turn right on Route 20 west.
16.3	1.1	Turn right on Route 11A.
16.5	0.2	Turn right on Route 11A south.
17.3	0.8	Turn right on Webster Road.
18.0	0.7	Turn left on Tully Farms Road - head south.
19.7	1.7	<u>STOP #2</u> - Otisco Road. After stop, continue south on Tully Farms Road down Tully Valley and up Valley Heads.
23.6	3.9	Turn right on Route 80. Note contact between Valley Heads moraine and bedrock of valley wall as you head west.
28.3	4.7	Bear left on to Oak Hill Road.
33.1	4.8	Turn right on Otisco Valley Road. Pass through town of Amber. Amber hanging delta 1.6 miles from turn.
35.4	2.3	Turn left on Route 174 at north end of Otisco lake. Forest Home Hotel, 1.1 miles from turn, provides view down Otisco valley.

Total Miles	Miles from Last Point	Route Description
38.6	3.2	Turn left at stop sign. Continue on Route 174 to Borodino.
39.8	1.2	Turn left on Route 41. Head south.
47.8	8.0	STOP #3 - Picnic area. Overlook of Skaneateles Lake. After stop, return to Route 41 and head north.
47.9	0.1	Turn left on secondary road at Onondaga-Cortland County line. Head down to Skaneateles Valley.
49.4	1.5	Turn right at stop sign on East Lake Road.
49.6	0.2	Turn left on Glenheaven Road. Drive around south end of Skaneateles Lake. Note bedrock outcrops along west valley wall.
55.6	6.0	Head straight across intersection with Route 41A. Drive west on New Hope Road.
56.6	1.0	Continue straight (west) on Burdock Road.
57.6	1.0	Turn left on Route 38A (Dutch Hollow Road). Head south to Moravia to end of Route 38A.
64.6	7.0	Head north (straight) on Route 38.
71.5	6.9	STOP #4 - Turn right on to Ensnore Road - Owasco Lake overview. After stop cross Route 38 and head west on Center Road to Scipio Center.
74.2	2.7	Turn left on Route 34. Head south to Venice Center.
77.3	3.1	Turn right on Poplar Ridge Road. Head toward Poplar Ridge. Cross "chevron moraine" just east of Poplar Ridge.
80.5	3.2	At Poplar Ridge cross Route 34B. Continue straight (west) on Poplar Ridge Road.
84.9	4.4	Turn left on Route 90. Head south.
85.0	0.1	Bear right on to Lake Road.
86.6	.16	STOP #5 - Turn right in to Long Point State Park on Cayuga Lake - LUNCH. After stop return to Lake Road and turn right.
88.0	1.4	Turn right on Route 90. Head south.
93.1	5.1	Turn right on Route 34B at King Ferry. Head south toward Ithaca.

Total Miles	Miles from Last Point	Route Description
105.1	12.0	Turn right on Route 34. Continue south to Ithaca.
110.6	5.5	Turn right on Route 34 south where it joins Route 13. Continue south through Ithaca. Stay on Routes 13/34.
114.4	3.8	<u>STOP #6</u> - Turn left into Buttermilk Falls State Park. After stop, leave Park and turn left on Routes 13/34.
114.9	0.5	Turn right on Route 13A. Head north toward Cayuga Lake.
116.9	2.0	Turn left on Route 79 west toward Watkins Glen. (**CAUTION**: World's most complex intersection!)
131.8	14.9	Turn left at stop sign. Continue west on Route 79 to Watkins Glen. Note kame deposits in quarry across road.
134.0	7.2	<u>STOP #7</u> - Turn right into Warren Clute Memorial (Lakeside) Park. After stop, leave Park, turn right, continue west on Route 79 into village of Watkins Glen.
139.8	0.8	Turn left on Route 14 (N. Franklin Street). Head south toward Horseheads. Note infilled lake valley to left as you head south from Watkins Glen.
155.2	15.4	Turn right at light. Route 14 south to Route 17.
156.6	1.4	Turn right on Route 17 west (Southern Tier Expressway). Head toward Corning. Drive will take you along proglacial meltwater channel.
166.6	10.0	Exit 47. Continue west on Route 17 west toward Jamestown. Pass through city of Corning.
190.8	24.2	Exit 38. Bath, N.Y. Turn right on Route 54 north.
190.9	0.1	Turn right at light. **<u>OVERNIGHT STAY</u>** at Super 8 Motel. In morning depart motel and head north on Route 54.
191.7	0.8	Left at light. Continue north on Route 54. Drive will take you across Valley Heads and down to dry lake floor.

Total Miles	Miles from Last Point	Route Description
198.6	6.9	Turn left on Route 54A into village of Hammondsport.
199.2	0.6	Continue straight at stop sign.
199.8	0.6	Go to next stop sign and turn right.
199.9	0.1	Bear left on Route 54A north. Drive along west shore of Keuka Lake to Branchport.
213.8	13.9	Turn right at blinking light in Branchport on Route 54A toward Penn Yan.
214.8	1.0	STOP #8 - Turn right into overlook at north end of northwest branch of Keuka Lake. After stop, leave overlook and turn left on Route 54A and return to Branchport.
215.8	1.0	Head straight at blinking light on Italy Hill Road. Head west towards Naples.
220.8	5.0	Turn right on Italy Turnpike. Continue west to Naples.
223.9	3.1	Yield sign. Continue straight toward Naples.
229.5	5.6	Turn right on Route 53 north. Drive down Valley Heads.
230.7	1.2	End Route 53. Continue straight on Route 21 north through village of Naples.
232.5	1.8	Bear left on County Road 12 north. Head up hill.
233.9	1.4	There will be a brief stop to overlook drillsite along Parrish Flat Road in valley.
235.6	1.7	STOP #9 - Turn right into private parking area. Overlook of Canandaigua Lake. After stop, turn left and return south along County Road 12.
236.0	0.4	Turn left on Griesa Hill Road. Head down hill.
237.2	1.2	Turn right on Route 21 south.
237.5	0.3	Turn left on Parrish Flat Road east.
238.0	0.5	STOP #10 - Drillsite in field near intersection DEC-owned old railroad bed and Parrish Flat Road. After stop, return to west on Parrish Flat Road.

Total Miles	Miles from Last Point	Route Description
238.5	0.5	Turn right on Route 21. Drive north to city of Canandaigua along west shore of Canandaigua Lake.
243.0	4.5	Bear right. Continue north on Route 21 toward Canandaigua.
255.8	12.8	Turn right. Continue north on Route 21.
256.2	0.4	Turn right (east) at light on Routes 21 north and 20 east to city of Canandaigua.
257.9	1.7	Turn left at light on route 21 north (Also Route 332).
258.9	1.0	Turn right on Route 21 north. Head toward New York State Thruway.
266.2	7.3	Turn left on to New York State Thruway (I-90). Head east toward Albany.
285.9	19.7	Exit 41. Waterloo/Clyde. Pay toll.
286.4	0.5	Turn right on Rouge 414 south.
286.7	0.3	Turn left at light on Route 318 east toward Auburn.
290.9	4.2	Turn left on Routes 20/5 east.
291.0	0.1	Turn left on Route 89 north.
293.7	2.7	Cross over New York State Thruway. Look to right (east) across Montezuma channel. Continue north on Route 89.
296.4	2.7	Turn left at stop sign on Armitage Road at Wayne County line. Head west.
297.0	0.6	<u>STOP #11</u> - Drumlin along east side of west branch of Montezuma Channels. After stop, turn around and head east on Armitage Road.
297.6	0.6	Continue straight at stop sign. Head east on Route 89 north.
298.9	1.3	Turn right on Route 31. Head east toward Syracuse.
301.1	2.2	Turn right on Route 90 south.
301.8	0.7	Turn right at blinking light. Continue south on Route 90. Note drumlins.
305.6	3.8	Turn left on Routes 20/5 east. Head to Auburn.
315.6	10.0	Turn right. Continue on Route 20 east.

Total Miles	Miles from Last Point	Route Description
315.7	0.1	Turn left. Continue on Route 20 east to Skaneateles.
322.5	6.8	STOP #12 - Park at north end of Skaneateles Lake. After stop continue east on Route 20 through village of Skaneateles.
324.5	2.0	Turn left on Route 175 toward Syracuse.
329.5	5.0	Turn right. Continue on Route 175 east toward Syracuse.
337.7	8.2	Head straight through traffic lights. Head east on Route 173 (W. Seneca Turnpike).
339.6	1.9	Turn left at light on Route 11 north.
340.7	1.1	Turn right toward I-81 north.
340.8	0.1	Turn left on to on-ramp for I-81 north.
342.4	1.6	Exit 18. Harrison and Adams Streets.
342.6	0.2	Turn right at end of off-ramp on E. Adams Street.
342.8	0.2	Turn right on top of hill on Irving Avenue.
343.0	0.2	Turn left at light on University Place.
343.1	0.1	Turn right at light on Crouse Drive. Return to Heroy Geology Lab on the campus Syracuse University.

-- END OF TRIP --

THE FOUNDERS OF AMERICAN GEOLOGY: A VISIT TO THEIR TOMBS, LABS, AND THEIR FAVORITE EXPOSURES: THE DEVONIAN LIMESTONES OF THE CAPITAL DISTRICT; A STUDY OF THE SEQUENCE STRATIGRAPHY OF THESE LIMESTONES

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INTRODUCTION

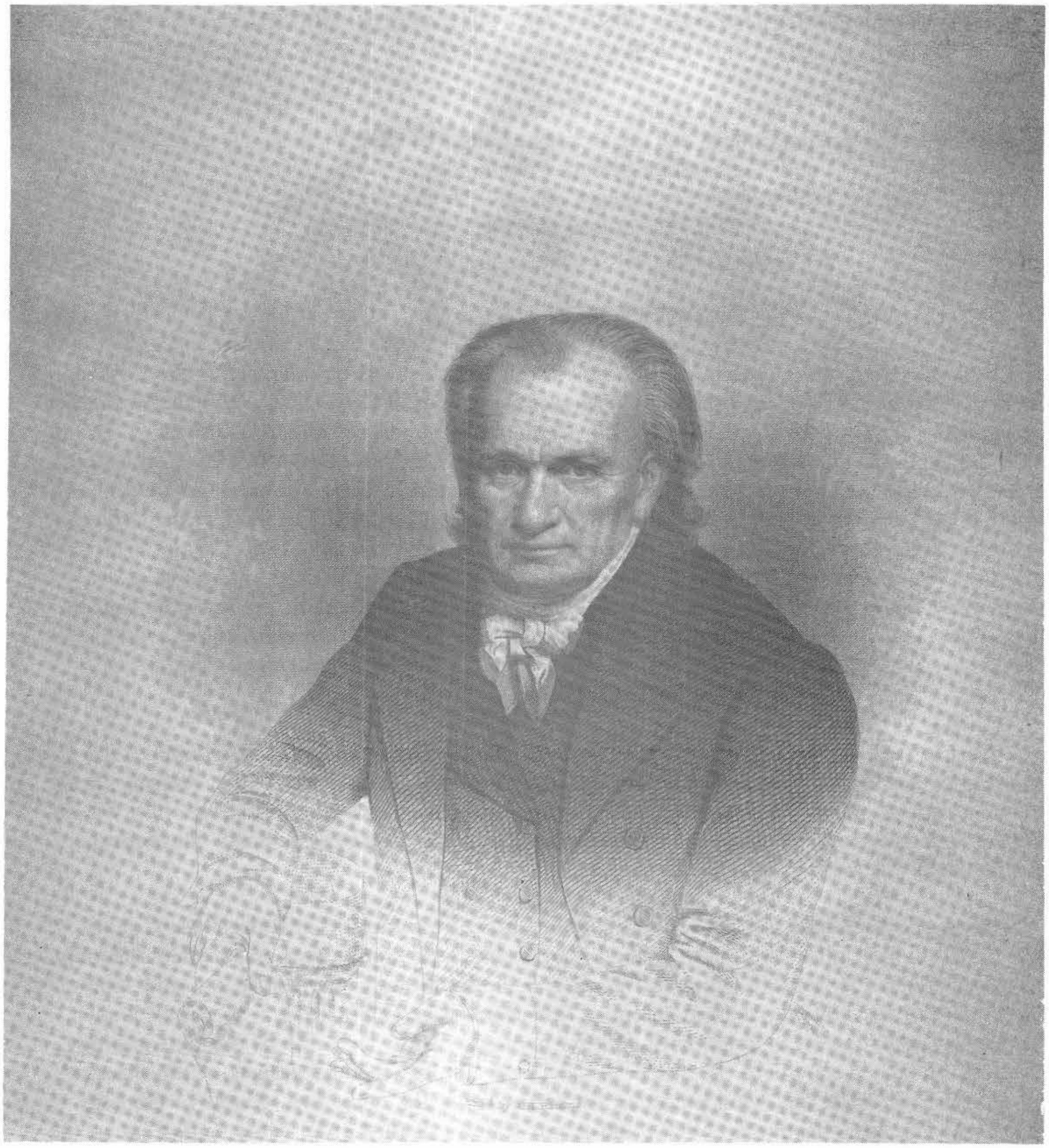
Located along the Helderberg Escarpment this classic site is on hallowed ground. Amos Eaton (1776-1842), Ebenezer Emmons (1799-1863), James Hall (1811-1898), William W Mather (1804-1859), Sir Charles Lyell (1797-1875), Benjamin Silliman (1779-1864), James D. Dana (1813-1895), Louis Agassiz (1807-1873), and Sir William E. Logan (1798-1875) have trod here before you, and we will view the memorial plaque erected in their honor. The field sites expose a large variety of Lower and Middle Devonian limestones, including coral reefs, stromatoporoid reefs, storm deposits, stromatolite facies, skeletal- and lime mud facies, solution-collapse features, and karst. This diversity of facies will be studied in terms of transgressive-regressive cycles, known as **parasequences**, which are defined as *conformable successions of genetically-related beds bounded by surfaces of erosion, called parasequence surfaces*.

HISTORY OF GEOLOGY

Before viewing the sequence stratigraphy of the Devonian we will pay tribute to the founders of American geology on whose concepts modern geology has built. Among those particular attention will be given to Amos Eaton, James Hall, and Ebenezer Emmons, whose debt we have incurred as builders of our science and at whose graves we will pay our respects. This field trip will begin in Troy, N.Y.

Troy, the hallowed ground of the geologic pioneers, is located in Rensselaer County, New York, named after the distinguished Van Rensselaer family who established the only successful Dutch Patroonship which thrived as a manorial estate from 1630 to the mid-1800's. One branch of the family produced Jeremias Van Rensselaer (1793-1871), a then well known, but now largely forgotten, nineteenth century geologist who wrote one of the first geology textbooks published anywhere. Entitled "Lectures on Geology, being Outlines of the Science" and published in 1825, this book preceded the textbooks of the other two "Giants of Geology" from Troy, Ebenezer Emmons (1826) and Amos Eaton (1830)(Fig. 1). More important to the purposes of this field trip, however is Stephen Van Rensselaer, born on November 4, 1764, who was a twelfth generation descendant of the original Dutch immigrant patroon. He graduated from Harvard University in 1782 served as New York State legislator from 1791 to 1796, as Lieutenant Governor of New York from 1795 to 1798, and as General of the New York State militia. His father, likewise Stephen Van Rensselaer, was the eighth and last Patroon and 6th Lord of the Manor of Rensselaerwyck; his mother was Catharine Livingston, daughter of Philip Livingston, one of the signers of the Declaration of Independence (Florence Van Rensselaer, 1956, p. 24,37).

In 1819 the legislature of the state of New York elected Stephen Van Rensselaer President of the Central Board of Agriculture. This board published two volumes on the geology of Albany and Rensselaer Counties authored by Amos Eaton. "It was believed then, and it is believed now, that these were the first two attempts in this country to collect and arrange



Amos Eaton

FIGURE 1. Amos Eaton in middle life.

geological facts, with the direct view to the improvement of agriculture" (Barnard 1839, p. 72). In Barnard's (1839) "Life, Services and Character of Stephen Van Rensselaer" he dwelt at length on Van Rensselaer's geological contributions: after republishing the studies on the geology of Albany and Rensselaer Counties "at his own cost, in a separate and convenient form, for extensive and gratuitous distribution" (Barnard,1839, p.74), he next turned his attention to a more extended scientific survey, to be carried through the entire length of the State on the line of the Erie Canal. This was commenced and prosecuted, under his orders, in the fall of 1822 by Professor Amos Eaton aided by two competent assistants. Van Rensselaer considered the geological studies of these two counties and the Erie Canal route part of a grander scheme, a plan for a "large and generous contribution to the science of Geology." This plan embraced a particular examination of the strata and formation of American rocks, by the survey of a transverse section, running across the Great Primitive Ranges of New England and the Transition and Secondary Ranges of eastern and western New York (Barnard,1839, p.75). He engaged Amos Eaton who completed this survey in 1823. His section extended from Boston to Lake Erie, a distance of about 550 miles, stretching across 9° of longitude and embracing a belt about 50 miles wide. At the same time, Professor (Edward) Hitchcock was employed to make a similar survey of a section across New England, a few miles north of that taken by Professor Eaton. In 1824, a publication was made, containing the results of these surveys, with maps exhibiting a profile view of the rocks in each of the sections. This work presents a connected actual inspection and survey, of greater extent than had ever been offered to geology.

However, according to Barnard (p.76) "the crowning glory of this good man's life" resulted on November 5, 1824 in the founding of the Rensselaer School, now Rensselaer Polytechnic Institute, to which he appointed two professors, a senior professor Amos Eaton, pathfinder of North American stratigraphy and one of the founders of American geology, and a junior professor, Lewis C. Beck, later to be the famous State Mineralogist of New York, who was followed by Ebenezer Emmons , one of the giants of nineteenth century American geology. By 1839, Rensselaer " had furnished to the community more State Geologists than has been furnished, in the same time, by all the colleges of the Union" (Barnard, 1839, p. 83), found in Friedman (1979,1981).

Amos Eaton, has been acclaimed the "Father of American Geology." He completed geological surveys of Albany and Rensselaer Counties, commissioned by the New York State Agricultural Society, but paid for by Stephen Van Rensselaer. Van Rensselaer also supported Eaton's geological survey of the territory adjoining the Erie Canal route during 1823-1824. In 1818 Eaton published a textbook, "An Index to the Geology of the Northern States." In this book Eaton not only incorporated a time and rock classification scheme, but also introduced a local guidebook, and published a cross section extending from the Atlantic Ocean to the Catskill Mountains. In 1824 Eaton appealed to Van Rensselaer for $300 as part of the effort to establish the Rensselaer School in Troy. Van Rensselaer provided these funds immediately and continued his financial support until 1829 when he ceased direct support of the school. Despite a heavy load of teaching and administration, Eaton published in 1830 a "Geological Textbook, Prepared for Popular Lectures on North American Geology;" its second edition appeared in 1832. In the second edition Eaton emphasized the importance of field work: students "must be shown the nearest rocks, from day to day." Eaton took his students on long field excursions into the mountains of New England and along the Erie Canal. At the time of his death in 1842, Eaton had become the most influential American geologist. In 1841 Sir Charles Lyell, during his trip to North America visited Eaton at Rensselaer. Eaton likewise received the respects of the Rev. William Buckland (1784-1856), the first professor of mineralogy and geology in the University of Oxford, England. In American geology the period 1818 and 1836 is known as the "Eatonian Era".

Among the most influential alumni of Rensselaer was James Hall, State Geologist of New York, known as the "Father of Geosyncline." In 1857 (published in 1859) Hall observed that, where the Paleozoic marine strata are thin (thicknesses of only few hundred or few thousands of meters), they are flat lying. In contrast,within the Appalachians thicknesses of equivalent age strata amount to tens of thousands of meters and the strata are not horizontal. Hall hypothesized that the subsidence of the strata within a trough, where they would be extra thick, provided the mechanism for folding them (Friedman and Sanders, 1978, p.435). Hall likewise, has become known as "Father of American Stratigraphy" and similarly, "Father of American Paleontology." Probably no other single person exerted a more influential role in the development of paleontology in North America.

James Hall is alleged to have originally literally walked the 220 miles from his home in Hingham, Massachusetts, to Rensselaer so that he might enroll and study under the great Eaton. Hall's first job at Rensselaer included whitewashing one of its buildings and tidying up the school; later he became librarian, and by 1835 he was listed as a full professor. Persuaded by Eaton, the New York State Legislature established a Geological and Natural History Survey in 1836 to which James Hall was appointed.

Another early alumnus who became a giant in the nineteenth century was Ebenezer Emmons. A graduate of Rensselaer in the first class of 1826, Emmons had been inspired by Eaton. Earlier, Emmons had studied under Eaton at Williams College when Eaton taught there in 1817. Emmons became Junior Professor at Rensselaer, a position he held for ten years, and a member of the New York State Geological Survey in 1836.

FIELD PROGRAM

We will first convene at the Rensselaer Center of Applied Geology, a center serving as Headquaters of the Northeastern Science Foundation, and affiliated with Brooklyn College and Graduate School and University Center of the City University of New York. This center was named after Jeremias Van Rensselaer whose 1823 publication An Essay on Salt, Containing Notices of its Origin, Formation and Geological Position and Principal Localites, Embracing, A Particular Description of the American Salines inspired the creation of one of the Foundation's journal Carbonates and Evaporites. Jeremias Van Rensselaer has been called the father of evaporite geology.

At the center we will view originals of the following publications which are central to understanding the development of American geology:

Eaton, Amos, 1822, A Geological and Agricultural Survey of Rensselaer County in the State of New York: Albany, E. and A. Hosford, 70 p.

__________, 1824, A Geological and Agricultural Survey of the District Adjoining the Erie Canal, in the State of New York: Albany, Parkard and Van Benthuysen, 163 p.

__________, 1830, Geological Textbook. Albany, Webster and Skinners, 63 p.

Emmons, Ebenezer, 1842, Geology of New York, Part II; Survey of the Second Geological District: Albany, W. & A White and J. Visscher, 427 p.

______________, 1846, Agriculture of New York: Albany, C. Van Benthuysen & Co., 371 p.

______________, 1854, Agriculture of New York: Albany, C. Van Benthuysen & Co., 267 p. and 47 plates.

Hall, James, The Natural History of New York, "Geology" and "Paleontology", 1843, 1847, 1852, 1859, 1861, 1867, 1879, 1884, 1885, 1893, 1894.

The local field trip begins with a visit to the grave of Amos Eaton at Oakwood Cemetary in Troy where we pay our respects (Fig.2) (STOP 1). Not far from Amos Eaton's grave is that of another distinguished American, Samuel Wilson (1766-1854), better known as Uncle Sam. The United States is nicknamed "Uncle Sam" in much of the world without knowing that "Uncle Sam" was in fact a distinguished citizen. During the War of 1812 Sam Wilson marked military supplies with the initials U.S., which came to stand for both United States and Uncle Sam. Wilson was one of the prime movers for incorporation of Troy as a village (1794) and as a city (1816).

After our visit to Eaton's grave we follow in the footsteps of Sir Charles Lyell. In 1841 (Lyell,1845) he reported on landslides in the city of Troy in which many people were killed. Lyell does not provide his source of information, but it may be the January 4,1837 edition of the *Albany Argus* which under the headline "Dreadful Calamity - Several Lives Lost" reported that in Troy "an avalanche of clay came tumbling from an eminence of 500 feet, moving down the base of the hill to level land, and then continued from the impulse it received to the distance of about 800 feet, covering up acres of ground, accompanied with a cataract of water and sand, which kept up a terrible roar. The mass moved along with great rapidity, carrying with it two stables and three dwelling houses." Lyell does not describe the geological setting for these slides. At the end of the Pleistocene the slope between the Hudson River and the plateau on which South Troy is located formed the shelf and slope of a vanished glacial lake, now known as Lake Albany. Slippery varved lake clays mark its sediments. This kind of setting is an invitation for disaster, and we will view the scarps resulting from slides at South Troy's Prospect Park. Although his location is not precise, Sir Charles Lyell probably inspected these same scarps.

From the Oakwood Cemetary in Troy we will cross the Hudson River to visit the Albany Rural Cemetary to see the graves of James Hall and Ebenezer Emmons (STOP 2).

Our next stop is in Lincoln Park, Albany, where James Hall and his co-workers did their research (STOP 3). A plaque on the building states: "This building was erected by James Hall, State Geologist of New York 1836-1898. For fifty years it served as his office and laboratory and from it graduated many geologists of merit and distinction. During most of that period it was an influential and active center of geological science in this country. Erected by the Association of American State Geologists 1916".

From Lincoln Park, Albany, we move on to the Indian Ladder Trail at the John Boyd Thacher State Park near Albany (Fig. 3).

Located in the Altamont and Voorheesville quadrangles, this classic site is on hallowed ground. A plaque erected in 1933 in memoriam of those pioneer geologists whose researches in the Helderbergs in the nineteenth century made this region classic ground includes not only almost all American pioneer geologists, but in addition lists pioneers of British and Canadian geology. Among those listed are Amos Eaton (1776-1842), the John Gebhards Sr. and Jr., James Hall (1811-1898), William W. Mather (1804-1859), Lardner Vanuxem (1792-1848), James Eights (1797-1882), Sir Charles Lyell (1797-1875), Benjamin Silliman (1779-1864), Edouard de Verneuil (1805-1873), James D. Dana (1813-1895), Henry D. Rogers (1808-1866), William B. Rogers (1804-1882), Ferdinand Roemer (1818-1891), Louis Agassiz (1807-1873), and Sir William E. Logan (1798-1875).

Sir Charles Lyell visited the "Helderberg Mountains", as he called them, in September 1841 and although he rejoiced, noting that "the precipitous cliffs of limestone, render this region more picturesque than is usual where the strata are undisturbed" (1845, p. 67), he was more concerned in his account with the "Helderberg war" between Van Rensselaer and his tenants. On his return to the "Helderberg Mountains" in May 1846 the "Helderberg war" absorbed him again since he states that "the anti-renters have not only set the whole militia of the site at defiance, but have actually killed a sheriff's officer, who was distraining for rents" (Lyell,

FIGURE 2. Gravestone of Amos Eaton in Oakwood Cemetary, Troy, N.Y.

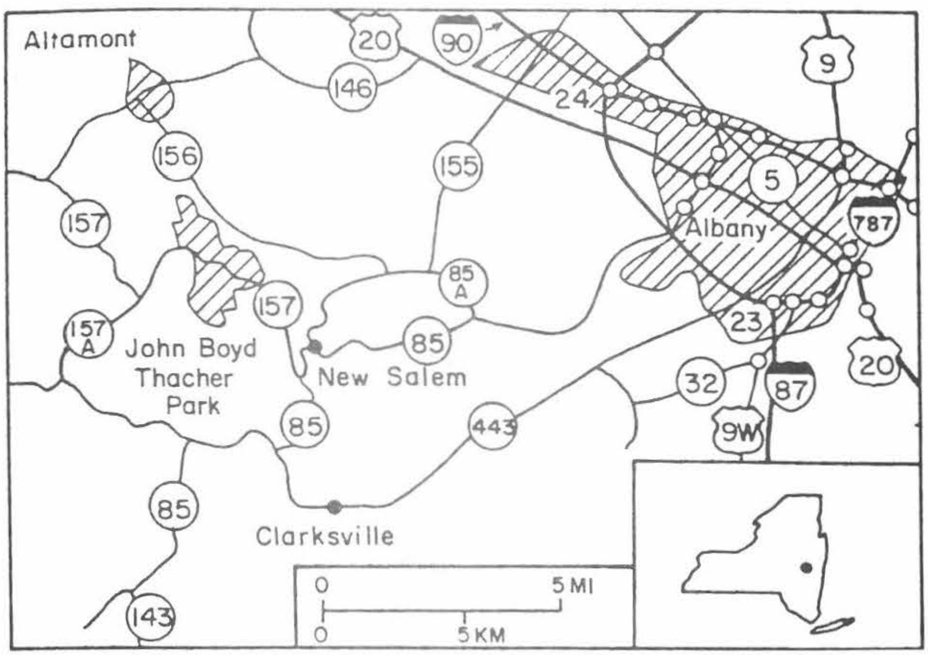


FIGURE 3. Location of John Boyd Thacher State Park (Fisher, 1987). The Indian Ladder Trail is located on Route 157.

1849, p. 260).

The definitive studies of the Lower Devonian carbonates of the Helderberg Escarpment exposed at John Boyd Thacher Park date to the early New York State Geological Survey and were written by Vanuxem (1842), Mather (1843), and Hall (1843). Their reports were supplemented and complemented later in the nineteenth century by the many volumes and papers by James Hall detailing stratigraphy and paleontology, followed in this century by many others, including Fisher (1987), Goldring (1935, 1943), Goodwin and Anderson (1982, 1985), Laporte (1967, 1969), Rickard (1962, 1975). and Friedman (1990). Gurney and Friedman (1986) described vertical parasequences of Lower Devonian limestones as transgressive-regressive cycles in Cherry Valley, approximately 35 miles west of Thacher Park.

The Indian Ladder Trail site provides an unusual opportunity to study a vertical cliff of limestone strata: a vertical exposure of approximately 80 ft or 24 m exposed in the cliff is accessible by stairway and footpath; handrailings assure safety. One can view the entire sequence of the rocks at close quarter, including by hand lens; comparable physical settings in quarries never allow such inspection.

LOCATION AND LOCAL HISTORY

Figure 3 shows the location of the John Boyd Thacher State Park where the Indian Ladder Trail reveals the vertical sequence of Lower Devonian limestones. Why the name Indian Ladder? Verplanck Colvin, one of the earliest men to write about the Helderbergs, in 1869 wrote:

"What is this Indian ladder so often mentioned? In 1710 this Helderberg region was a wilderness; nay all westward of the Hudson River settlement was unknown. Albany was a frontier town, a trading post, a place where annuities were paid, and blankets exchanged with Indians for beaver pelts. From Albany over the sand plains... "Schenectada", (pine barrens) of the Indians... led an Indian trail westward. Straight as the wild bee or the crow the wild Indian made his course from the white man's settlement to his own home in the beauteous Schoharie valley. The stern cliffs of these hills opposed his progress; his hatchet fells a tree against them, the stumps of the branches which he trimmed away formed the round of the Indian ladder."

The trail ended where the cliff did not exceed twenty feet in height. Here stood "the old ladder." In 1820 this ladder" was yet in daily use (Goldring, 1935). The modern stairway crosses the old Indian ladder road which ran to the top of the escarpment where the trail begins.

Entering Thacher State Park from Albany on Route 157 stop at the "Cliff Edge Overlook" for a view of the Taconic and Berkshire Mountains, Adirondacks, Hudson River, and City of Albany, then drive to the next parking lot which has a sign La Grange Bush Picnic Area - Indian Ladder Trail. Descend here for study of the Lower Devonian carbonate facies. Examine also the memorial plaque near the cliff edge at the Mine Lot Creek parking lot which has been attached to a vertical rockwall. It says "in memory of those pioneer geologists whose researches in the Helderbergs from 1819 to 1850 made this region classic ground". The names of these pioneers have been cited in the introduction to this paper.

SIGNIFICANCE

The cliff face exposes an excellent case history of sequence stratigraphy. Lower Devonian limestones of the Helderberg Group reveal three parasequences which may be recognized among the exposed formations (Rondout, Manlius, and Coeymans formations) (Figs.4&5). Parasequences are the building blocks of vertical sequences. A parasequence is defined as a relatively conformable succession of genetically related beds bounded by surfaces (called parasequence surfaces) of erosion, nondeposition, or their correlative conformities (Van Wagoner. 1985). Each

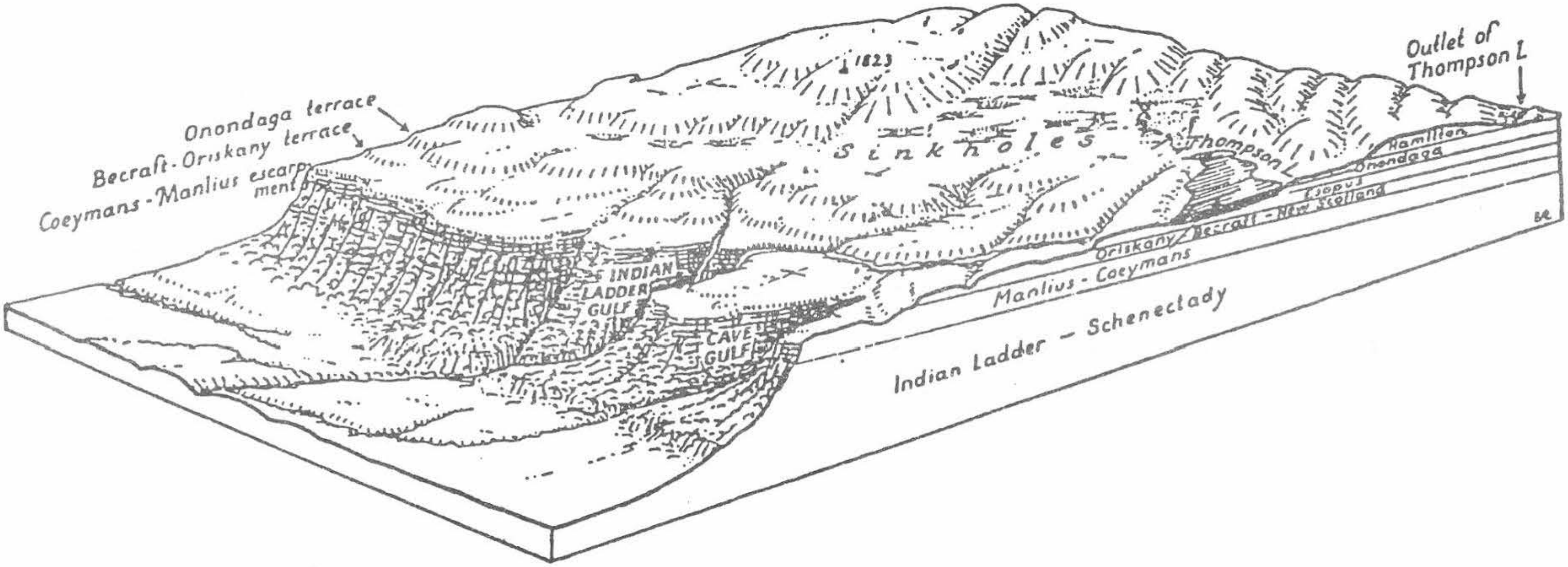


FIGURE 4. Block diagram of Helderberg Escarpment showing Indian Ladder location (labeled Indian Ladder Gulf), dip-slope of Paleozoic strata, prominent terraces, and sinkhole topography (H. F. Cleland, 1930; modified by Winifred Goldring, 1935).

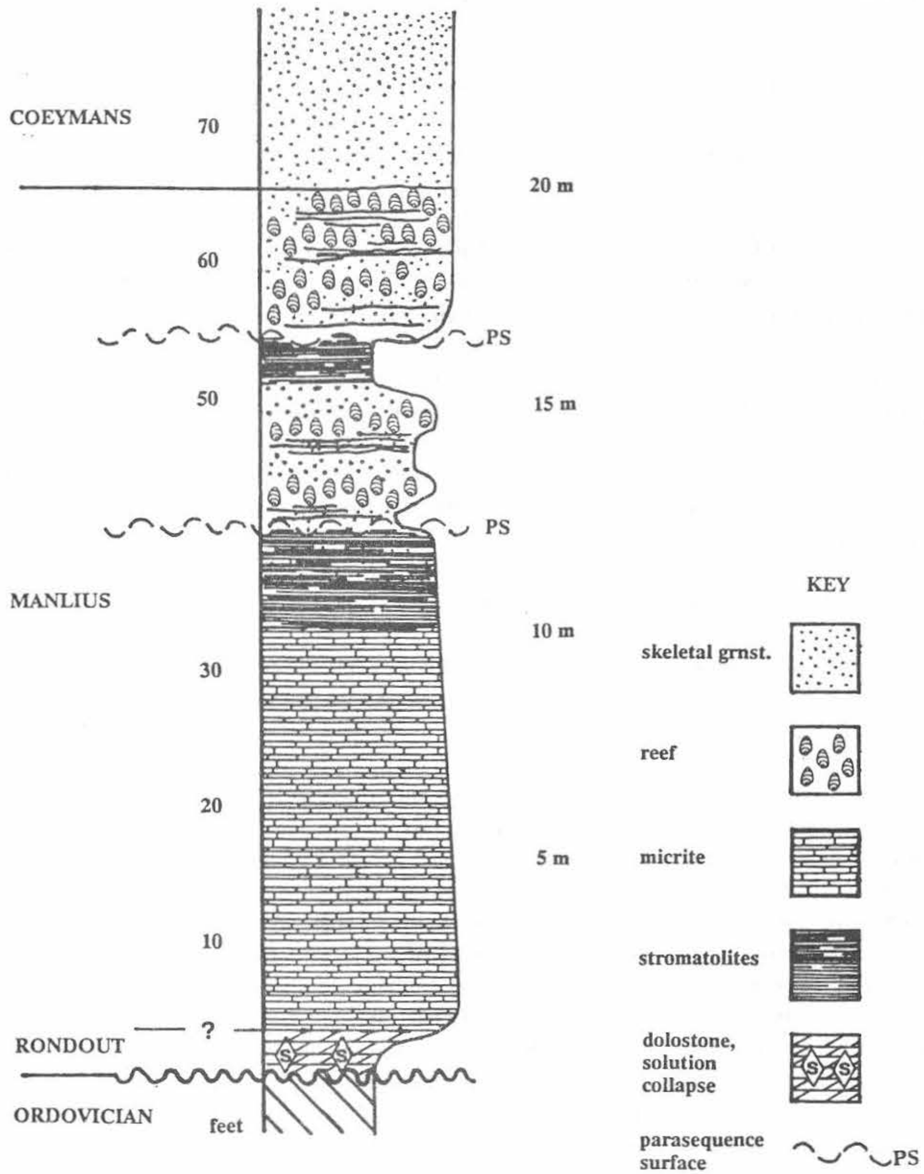


FIGURE 5. Stratigraphic section and vertical sequence revealing parasequences at Indian Ladder Trail.

sequence is initiated by a eustatic fall in sea level rapid enough to overcome subsidence or by epeirogenic upward motion. A parasequence surface commonly is an unconformity surface.

THE STRATIGRAPHIC COLUMN

Below the Indian Ladder Trail, where a waterfall known as Minelot Falls spouts across the path, sandstones and shales of the Middle Ordovician Schenectady formation are mostly concealed beneath a cover of blocks of Devonian limestone forming a talus slope. At the waterfall a major unconformity just below the trail separates the Ordovician strata from the Rondout Formation exposed at the base of the cliff. The nonfossiliferous Rondout Formation has been classed as latest Silurian or earliest Devonian (Fisher, 1987).

Figure 4 shows the columnar stratigraphic section, the parasequence surfaces, and facies distribution of the Lower Devonian carbonates. The Rondout Formation at the base is overlain by the Manlius Formation, and the top of the section extending to the break in slope at the top of the cliff is occupied by the Coeymans Formation. The stratigraphic section exposed on this trail is the type locality for the Thacher Member of the Manlius Formation, proposed by Rickard (1962).

PARASEQUENCES

Studies of vertical sequences should normally be worked from the base of the section upward. However, at this exposure it is best to work the section downward following the stairway from the edge of the cliff.

The top of the section is composed of skeletal grainstone (locally skeletal packstone) in which fossils, especially brachiopods, corals and crinoids, are evident (Fig. 5); the pentamerid *Gypidula coeymanensis* is prevalent. This facies is part of the Coeymans Formation. Its lower contact is sharp and obvious in the field. Below this contact follows the Manlius Formation which underlies most of this escarpment. A stromatoporoid reef with loclly intercalated skeletal grainstone represents the top of this formation (Fig. 5). The stromatoporoids show their distinctive globular concentric structures resembling cabbage heads. Previous authors (Fisher, 1987; Rickard. 1962) have termed this reef facies a biostrome, presumably because its geometry in outcrop is sheetlike rather than moundshaped. In my experience with reefs of all ages I have observed that most large reefs are flat on top and bottom, especially on the scale of this exposure. Other geologists share this experience, thus Shaver and Sunderman (1989) note "virtually all large reefs seen on outcrop have eroded, flattened tops, whereas smaller reefs that were not naturally aborted and that were unaffected by erosion as seen on outcrop have convex-upward rounded tops".

Close examination of the reef facies reveals a fine-grained matrix between the framework-building stromatoporoids. This matrix resembles micrite, a lithified former lime mud; hence this facies may be misinterpreted as representing a low-energy setting.
However, in analogous modern reefs cement forms millimeters to centimeters beneath the living part which in thin section is finely crystalline (cryptocrystalline) and semi-opaque. Hence the matrix in such reef rock looks just like low-energy micrite (Friedman et al., 1974). Case histories abound where unwary geologists have confused high-energy reef rock with a supposed "low-energy" lime-mud facies (Friedman, 1975). Therefore the observation of a fine-grained matrix between the framework builders does not deter, in fact confirms, the interpretation that this part of the section formed as a high-energy reef facies, and not in a low-energy setting.

The stromatoporoids are massive which in the ecologic zonation of Devonian reefs represents the shallowest-water zone of a subtidal setting.

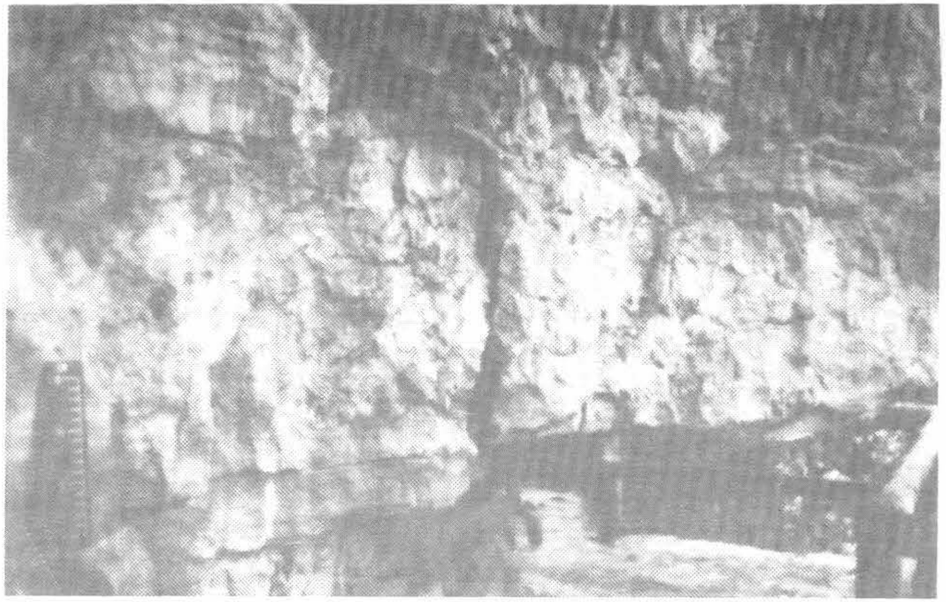


FIGURE 6. Photograph showing recessed underlying stromatolitic (finely-laminated) facies and overlying stromatoporoid reef facies. Sharp contact between the two facies on which scale rests is a parasequence surface (see fig. 5). Manlius Formation.

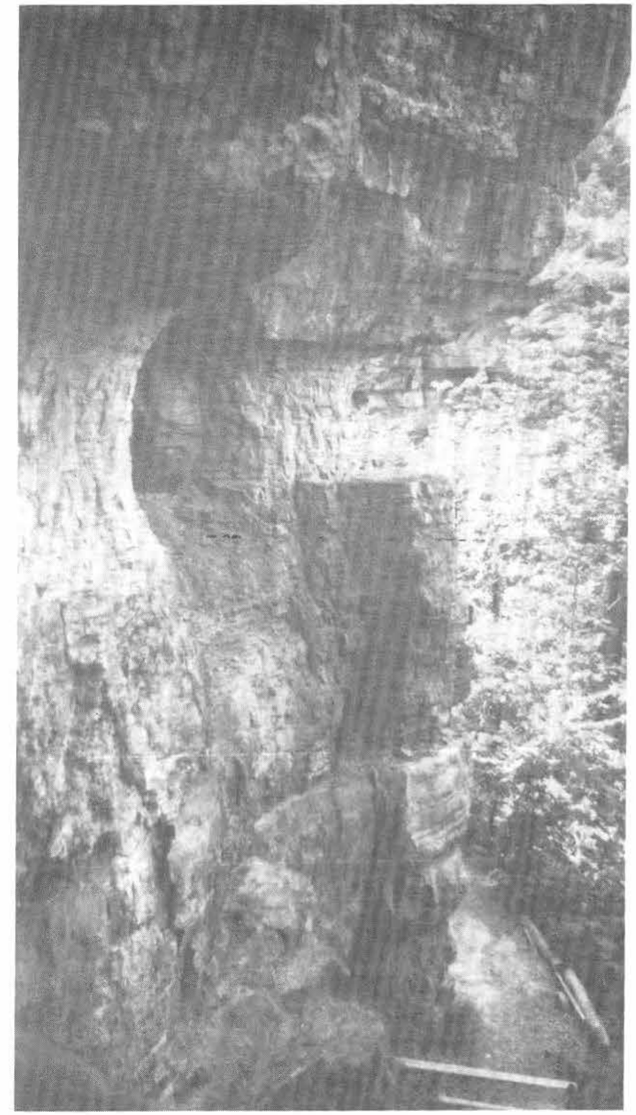


FIGURE 7. Photograph showing from top downward: massive projecting reef separated by parasequence sur face from underlying recessed stromatolites ("Upper Bear Path"). Below recessed stromatolite facies note in descending order bedded skeletal grainstone, nonbedded reef, bedded skeletal grainstone, reef, parasequence surface, and non-recessed stromatolite facies recognizable as a well-bedded facies. For detail compare with figure 5. Manlius Formation.

Below the reef facies occurs a stromatolitic (finely laminated) facies which is recessed back creating a near cavelike morphologic feature (Fig. 6). This recessed feature can be traced throughout Thacher State Park and is known as "Upper Bear Path". By analogy with modern environments the stromatolitic facies represents a low-energy intertidal or supratidal setting. The sharp contact between the intertidal or supratidal low-energy stromatolitic facies and overlying subtidal high-energy reef facies represents a parasequence surface (Figs. 5 and 6). Downward from the stromatolites a stromatoporoid reef facies once again recurs, separated by bedded skeletal grainstone from the stromatolites; in fact the reef facies is present twice (Fig. 7). Hence downward the setting changes from intertidal or supratidal to subtidal shallow water. Below this double reef section the change is again to interpreted intertidal or supratidal stromatolites. Hence once again a parasequence surface separates the subtidal high-energy reef facies from the underlying intertidal to supratidal stramatolites (Fig. 5). Interestingly, this stromatolitic facies is resistant to erosion (Fig. 7), hence projects out in the cliff, whereas the upper stromatolite facies is recessed almost cavelike. Below this lower stromatolite facies the lithology and facies are that of a low-energy thin-bedded micrite with local skeletal grainstone occurring as finely intërbedded couplets, scour-and-fill structures, local cross-bedding, and some beds containing abundant spiriferid brachiopods, tentaculitids, ostracodes, and bryozoans. Near the base of the Manlius Formation occur several thicker beds, up to about 20 cm in thickness.

Near the base of the section is the Rondout Formation. Its exact contact with the overlying Manlius Formation is subject to debate. In the columnar section (Fig. 5) the Rondout Formation is identified where solution-collapse features are prominent and the lithology changes to dolomitic, especially dolomitic stromatolites, with sporadic intercalated calcitic laminae and shale laminae, an interpreted supratidal facies. Clasts of solution-collapse breccia are prominent together with gypsum-filled veins. The angular clasts of collapse breccia resulted from collapse and brecciation of overlying carbonate strata when evaporites underlying them were dissolved. It represents a karst setting. The Rondout Formation is commonly known as Rondout Waterlime. Its base is at or below the trail.

ITINARY

From Oneonta drive to Troy, where this itinary begins.

An informal stop will be made at the Rensselaer Center of Applied Geology, 15 Third Street, for informal discussion and scanning of the historical geological publications of Eaton, Hall and Emmons (for titles see HISTORY OF GEOLOGY).

Road Log

CUMULATIVE MILEAGE	MILES FROM LAST POINT	ROUTE DESCRIPTION
0	2.1	From Rensselaer Center of Applied Geology (Third Street, Troy) go via Broadway, River Street, Fulton Street and Sixth Ave. (turn left) to Hoosick St. (Rte 7), turn right; one block on Hoosick St., turn left onto Tenth St. which becomes Oakwood Ave. Enter Oakwood Cemetery. On entering the cemetery keep right 0.1 mile to section S2.

STOP 1.Visit the grave of the father of American geology Amos Eaton (1776-1842). For details on his life read the section on History of Geology.

Return to Hoosick Street(Rte. 7) and cross Hudson River to I 787 via Collar Bridge. Drive I 787 south to I 90 and take Watervliet exit (NY 378) west.

0 2.9 Take NY 378 to first traffic light; turn right into historic Albany Rural Cemetery)(established 1841).

STOP 2. Visit the graves of James Hall (1811-1898) (Lot 93, Section 18) and Ebenezer Emmons (1799-1863)(Lot 46, Section 6).

James Hall. - James Hall was the single most prominent American geologist of the 19th century. The impressive honors he received are worth listing chronologically. Hall obtained the degrees of A.B. and A.M. from Rensselaer in 1831 and 1832. In 1837 he was elected Member of the Imperial Mineralogical Society of St. Petersburg, Russia. In 1840 he was one of the founders of the American Association of Geologists which became the American Association for the Advancement of Science. In 1843 he was elected correspondent of the Academy of Natural Sciences of Philadelphia. In 1848 he was elected Foreign Member of the Geological Society of London, the number of foreign members being limited to 50. In 1848 he was elected fellow of the American Academy of Arts and Sciences. In 1856 he became President of the American Association for the Advancement of Science and in 1863 he was named, by an Act of Congress, to be one of the 50 original members of the National Academy of Science. The King of Italy conferred on him the title of Commander of the Order dei Santi Maurizzio é Lazzaro. Many other honors too numerous to be cited may be added.

Ebenezer Emmons.--alumnus and Junior Professor at Rensselaer, member of the New York State Geological Survey, founder of the North Carolina Geological Survey, and State Geologist of North Carolina; father of the Taconic System.

A graduate of Rensselaer in the first class of 1826, Emmons had been inspired by Eaton. Emmons became Junior Professor at Rensselaer, a position he held for ten years, and a member of the New York State Geological Survey in 1836. Later he was State Geologist of North Carolina, spreading Rensselaer's influence in American geology through his textbooks and advocacy of the Taconic system. Emmons had noted the presence of a group of rocks between the Potsdam Sandstone, the lowest of the sedimentary formations in New York and what was at the time called the Primitive Rocks of Central Vermont. This interval he proposed to call the Taconic System. Emmons acquainted the public with the Adirondack Region and gave the names to principal mountains. Classics which Emmons published include Manual of Mineralogy and Geology (1826), Report on the Second Geological District of New York (1842), Natural History of New York (1848), American Geology Containing a Statement of Principles of the Science With Full Illustrations of the Characteristic American Fossils (1854), Treatise Upon American Geology (1854), The Swamplands of North Carolina (1860), and Textbook of Geology (1860).

The antagonism of Hall helped chase Emmons out of New York. Yet in death their graves, close to each other, make their old controversy seem remote. Hall never acknowledged Emmons' thought-provoking studies on the Adirondack and Taconic Regions. One of the Adirondack peaks has been named Mount Emmons.

0	1.3	Leave Albany Rural Cemetery. At Exit (traffic light) turn left , enter the village of Menands, drive to junction with 787, go south to Albany.
5.7	4.4	Take exit US 20 at Port of Albany, go straight on Madison Avenue (Rts. 20 and
5.9	0.2	32) to South Pearl St. (make left turn on to South Pearl St.)
6.25	0.35	Go 3 blocks to Morton Avenue and make a right turn on to Morton Ave.
6.8	0.45	Drive 6 blocks to South Swan Street (count blocks on right).

6.9 0.1 Make right turn into South Swan St. and take first left into Lincoln Park. Alight at Sunshine School.

STOP 3. The annex to Sunshine School is the historical laboratory of James Hall. For fifty years this building served as his office and laboratory.

Return to I 787 North

0	1.8	To I 90 West
4.1	5.9	Take I 90 to Slingerlands (Exit4) Route 85. At traffic light turn right (west) into NY 85.
6.8	12.7	Enter Town of New Scotland
5.4	18.1	Take 157 west
2.2	20.3	Enter John Boyd Thacher Park
1.2	21.5	Park in La Grange Bush Picnic Area. Indian Ladder Trail.

STOP 4. Indian Ladder Trail at the John Boyd Thacher State Park. Before descending to see the exposures read the sections titled*Significance, the Stratigraphic Column, and Parasequences.* Study especially figure 5. Then descend.

Study the section as you descend to the base of the trail: identify the facies, parasequence surfaces, contacts, and formations. On your return to the top of the section measure the thickness of various stratigraphic and facies units and plot them on a striplog or in a notebook. Compare your data with those shown on figure 5. After completion of the exercise visit the memorial plaque at the top of the trail near Mine Lot Creek parking lot. It is this plaque which explains why the ground on which you tread at this site is sacred.

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Geology and Geochronology of the Southern Adirondacks

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PROLOGUE

Geologic investigations of the Adirondack Region began in the nineteenth century with the early surveys of Ebenezer Emmons and later with important contributions from Kemp, Smyth, and Cushing (see Buddington 1939, for complete bibliographic references of these and other early workers). in the first three decades of the twentieth century a large number of significant publications were forthcoming from Miller, Newland, Alling, and Balk, as well as those cited previously, including Buddington. The essence of these contributions is summarized in Buddington's classic Geological Society of America Memoir (1939) entitled "Adirondack Igneous Rocks and their Metamorphism". The central, dominant theme of this work is that the Adirondacks consist of a vast collection of intrusive igneous complexes ranging in composition from anorthosite to granite. This consensus view was strongly supported by a wealth of field and chemical data and may be regarded as comprising Phase I in the history of Adirondack geology. The culmination of Phase I corresponds with the publication of U.S. Geological Survey Professional Papers 376 and 377 by Buddington and Leonard (1962) and Leonard and Buddington (1964), respectively. These works were the result of government sponsored mineral exploration efforts during World War II. Both reports manifest the classical view of the Adirondacks as an igneous-plutonic domain.

Phase II begins with Engel and Engel (1958, 1964) in the lowlands and Walton and deWaard (1963) and deWaard (1964, 1965) in the highlands. Conceptually, this phase is characterized by paradigms of stratigraphy and stratigraphic correlation. Its modus operandi was the recognition and definition of rock sequences, interpreted as stratigraphic, and correlated over great distances. Depending upon the particular investigators, this stratigraphic framework was cross-bred with varying degrees of granitization and metasomatic transformations.

A late stage of Phase II is represented by the investigations of McLelland in the highlands (McLelland and Isachsen, 1986) and Carl, Foose, deLorraine, and others in the lowlands (see Carl et al. 1990 for references). In this stage, metavolcanics played a burgeoning role in the interpretation of Adirondack layered sequences (Whitney and Olmsted 1989). In addition, structural investigations documented the existence of large fold-nappe structures within most of the Adirondacks (deWaard 1964, McLelland 1984). A characteristic of this stage of Phase

II was to downgrade the importance of widespread igneous intrusion and to substitute for it metamorphic processes involving recrystallization of stratified volcanics and sediments into layered gneisses modified by local anatectic effects (Carl et al. 1990; Whitney and Olmsted 1989).

Phase III of Adirondack geology began in the mid-1970's when Eric Essene, together with his students John Valley and Steve Bohlen (see Bohlen et al. 1985 and Valley et al. 1990) established a quantitative framework for Adirondack pressure, temperature, and fluid conditions during metamorphism. This approach has been significantly augmented by the oxygen isotope studies of John Valley and Jean Morrison (see Morrison and Valley 1988 for complete reference) and the U-Pb studies of Klaus Mezger (1990), all of which have provided critical data that constrain models of Adirondack evolution. Simultaneously, McLelland and Chiarenzelli (McLelland et al. 1988, 1991a,b; McLelland and Chiarenzelli 1990, 1991) have conducted a U-Pb zircon study of the Adirondacks in order to follow up on Silver's (1969) pioneering, landmark investigations. The quantitative results of these research programs have provided unequivocal boundary conditions with which any interpretations of Adirondack geology must be consistent. These results and associated boundary conditions are presented in the text that follows. Significantly, and interestingly, the numbers demonstrate that Phase I interpretations were much closer to the truth than the elaborate stratigraphic models of Phase II. It has become increasingly clear that, in the Adirondacks, intrusive igneous rocks greatly dominate over metavolcanics, or even possible candidates for metavolcanics. Accordingly, it has become evident that layering in orthogneisses is not of primary origin but represents examples of tectonic layering of the sort described by Davidson (1984) in tectonites referred to as straight gneisses. Highly strained rocks of this sort have been described for the Piseco anticline by McLelland (1984) and are common throughout the region.

In conclusion, modern quantitative petrologic and isotopic data strongly indicates that the early, and classic, interpretations of the Adirondacks were, in the main, very nearly correct and herein lies a lesson worth pondering. These results offer additional support for the well documented thesis that granites are plutonic, intrusive rocks and that attempts to form them by circuitous, non-intrusive mechanisms are both outdated and destined to failure. This assessment applies equally well to long discredited examples of granitization and to more modern attempts to account for granites by metamorphosing acidic volcanics. Among the critical observations related to these conclusions are quantitative results from geothermometry, geobarometry, geochronology, and petrochemistry. Combined with a modern understanding of tectonic layering, these considerations can greatly constrain the interpretation of complex geologic terranes, as described below for the southern Adirondack region.

INTRODUCTION AND GEOCHRONOLOGY

The Adirondacks form a southwestern extension of the Grenville Province (fig. 1) and are physiographically divided into the Adirondack highlands (granulite facies) and lowlands (amphibolite facies) by a broad zone of high strain referred to as the Carthage-Colton Mylonite Zone (figs. 2,3) which is continuous with the Chibougamau-Gatineau line (AB on fig. 1). Together these two zones separate the Grenville Province into two major blocks with the Central Granulite Terrane (CGT) lying east of AB and the Central metasedimentary Belt (CMB) and Central Gneiss Belt (CGB) lying to the west. Within the southwestern portion of the Grenville Province further subdivisions exist and are shown in figure 3.

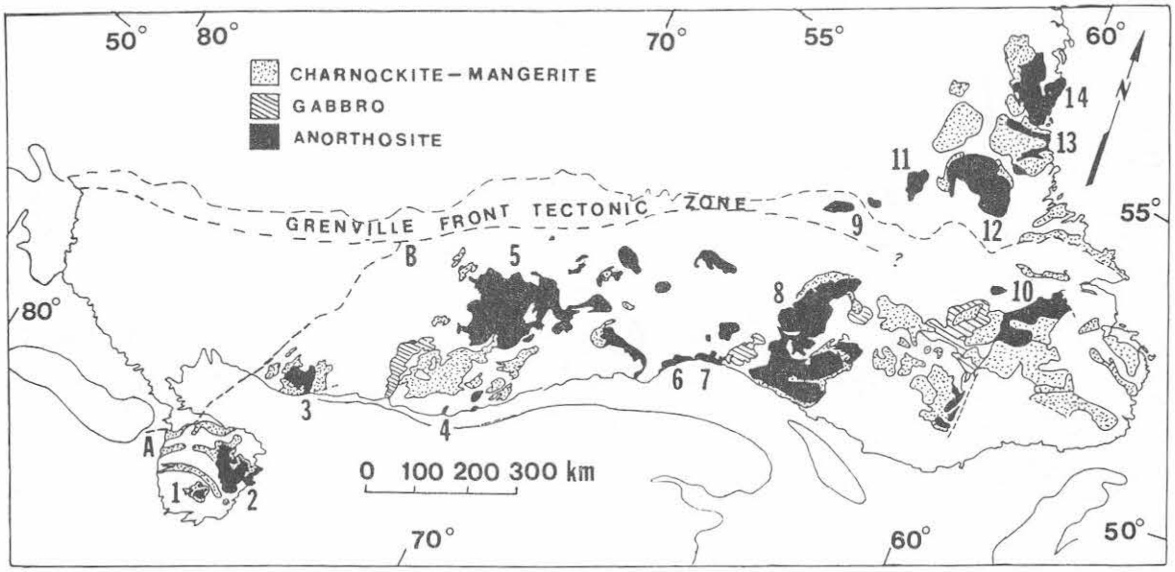


Fig. 1. Generalized map of anorthositic massifs within the Grenville Province and adjacent Labrador. The dashed line, AB, separates terranes with anorthosite massifs on the east from ones lacking them on the west and corresponds to the Carthage-Colton-Gatineau-Chibougamau Line. 1-Snowy Mt. and Oregon domes (ca. 1130 Ma); 2-Marcy massif (ca. 1135 Ma); 3-Morin anorthosite and Lac Croche complex (1160±7 Ma); 4-St. Urbain anorthosite (ca. 1070 Ma); 5-Lac St. Jean complex (1148±4 Ma); 6-Sept Isles (1646±2 Ma); 7-8–Harvre St. Pierre complex (1126±7 Ma) including the Pentecote (1365±7 Ma) anorthosite; 9-Shabagamo intrusives; 10-Mealy Mts. anorthosite (1646±2 Ma); 11-12-Harp Lake anorthosite (ca. 1450 Ma); 13-Flowers River complex (ca. 1260 Ma); 14-Nain complex (1295 Ma) including Kiglapait intrusive (1305±5 Ma). From McLelland (1989).

As demonstrated by recent U-Pb zircon and Sm-Nd geochronology summarized (table 1) by Daly and McLelland (1991), McLelland and Chiarenzelli (1991) and Marcantonio et al. (1990), the Adirondack-CMB sector of the Grenville Province contains large volumes of metaigneous rocks that represent recent (i.e., ca. 1400-1200 Ma) additions of juvenile continental crust. These results (fig. 4) indicate that the Adirondack-CMB region experienced wide-spread calcalkaline magmatism from ca. 1400-1230 Ma. Associated high grade (sillimanite-K-feldspar-garnet) metamorphism has been fixed at 1226±10 Ma by Aleinikoff (pers. comm.) who dated dust air abraded from metamorphic rims on 1300 Ma zircons. Identical rocks, with identical ages, have been described from the Green Mts. of Vermont by Ratcliffe and Aleinikoff (1990), in northern Ireland by Menuge and Daly (1991), and in the Texas-Mexico belt of Grenville rocks (Patchett and Ruiz 1990). It appears, therefore, that a major collisional-magmatic belt was operative along the present southern flank of the Grenville Province during the interval 1400-1220 Ma and may have been related to the assembly of a supercontinent at this time. More locally, this magmatism along with associated metamorphism, represents the Elzevir Orogeny of the Grenville Orogenic Cycle. as defined by Moore and Thompson (1980). Within the Adirondacks Elzevirian rocks are

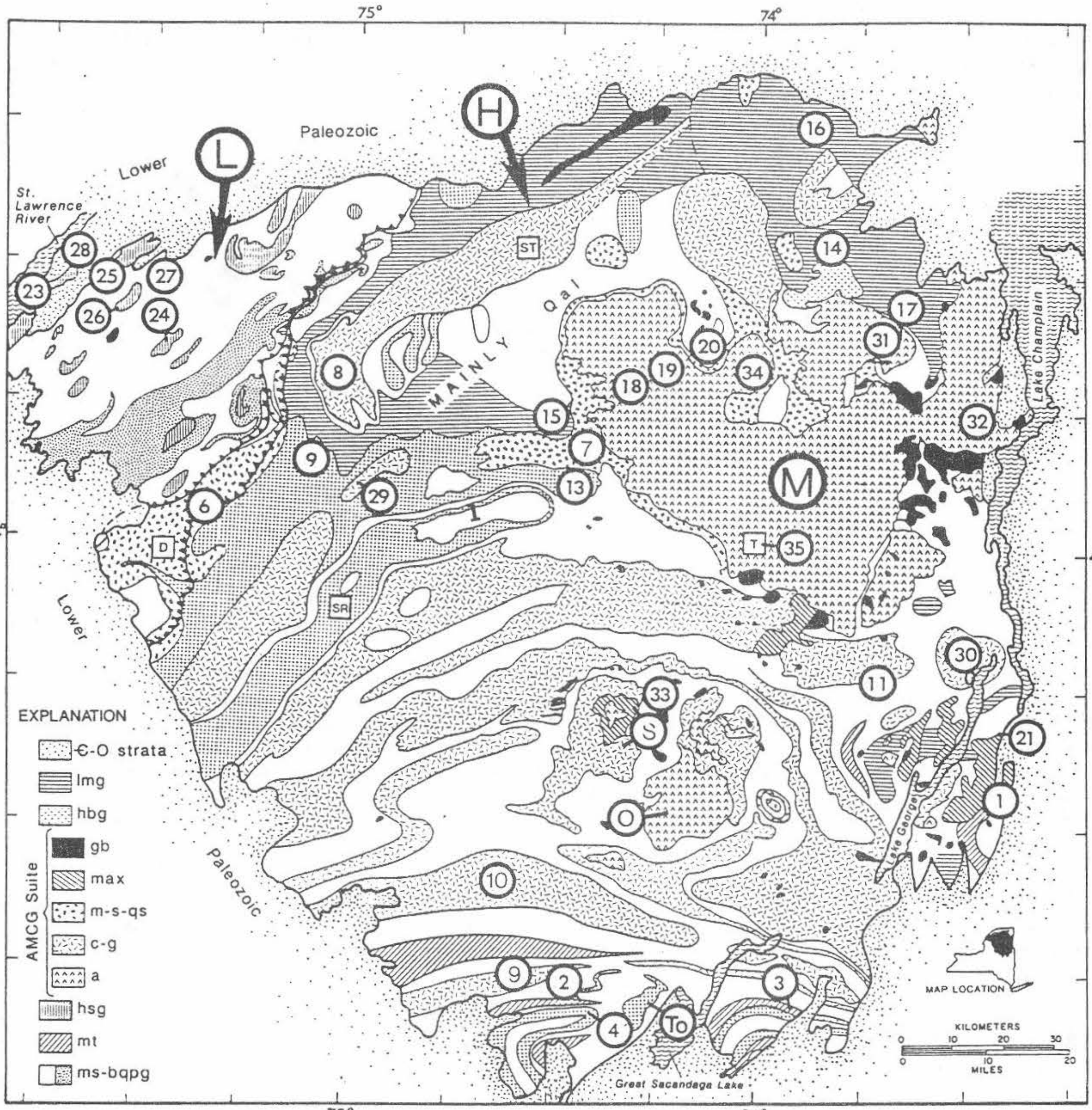


Fig. 2. Generalized geologic map of the Adirondack Highlands (H) and Lowlands (L). The Carthage-Colton Mylonite Zone (CCMZ) is shown with sawteeth indicating directions of dip. Numbers refer to samples listed in Tables 1 and 2. Map symbols: lmg=Lyon Mt. Gneiss, hbg=hornblende-biotite granitic gneiss, gb=olivine metagabbro, max=mangerite with andesine xenocrysts a=metanorthosite, m-s-qs=mangeritic-syenitic-quartz-syeniticgneiss, ms=metasediments, bqpg=biotite-quartz-plagioclasegneiss, hsg=Hyde School Gneiss, mt=metatonalitic gneiss. Locality symbols: A=Arab Mt. anticline, C=Carthage anorthosite, D=Diana complex, O=Oregon dome, S=Snowy Mt. dome, ST=Stark complex, SR=Stillwater Reservoir, T=Tahawus, To=Tomantown pluton. From McLelland and Chiarenzelli (1990) and Daly and McLelland (1991).

TABLE I
U-PB ZIRCON AGES FOR META-IGNEOUS ROCKS OF THE ADIRONDACK MOUNTAINS

No.	Age (Ma)	Location	Sample No.
HIGHLANDS			
Tonalitic gneiss and older charnockite			
1	1329 ±37	South Bay	AM-87-12
2	1301*	Canada Lake	AM-86-12
3	1336*	Lake Desolation	LDT
4	1233*	Canada Lake	AM-87-13
Mangeritic and charnockitic gneiss			
5	1155 ±4	Diana complex	
6	1147 ±10	Stark complex	AM-86-15
7	1134 ±4	Tupper Lake	AC-85-6
8	1125 ±10	Schroon Lake	9-23-85-7
Older hornblende granitic gneiss			
9	1156 ±8	Rooster Hill	AM-86-17
10	1150 ±5	Piseco dome	AM-86-9
11	1146 ±5	Oswegatchie	AC-85-2
Younger hornblende granitic gneiss			
12	1100 ±12	Carry Falls	AM-86-3
13	1098 ±4	Tupper Lake	AM-86-6
14	1093 ±11	Hawkeye	AM-86-13
Alaskitic gneiss			
15	1075 ±17	Tupper Lake	AM-86-4
16	1073 ±6	Dannemora	AM-86-10
17	1057 ±10	Ausable Forks	AM-86-14
Anorthosite and metagabbro			
18	1054 ±20	Saranac Lake	AC-85-8§
19	1050 ±20	Saranac Lake	AC-85-7‡
20	996 ±6	Saranac Lake	AC-85-9
Xenolith-bearing olivine metagabbro			
21	1144 ±7	Dresden Station	AM-87-11
22	1057	North Hudson	CGAB**
LOWLANDS			
Leucogranitic gneiss			
23	1415 ±6	Wellesley Island	AM-86-16
Alaskitic gneiss			
24	1284 ±7	Gouverneur dome	AC-85-4
25	1236 ±6	Fish Creek	AM-87-4
26	1230 ±33	Hyde School	AC-85-5
Granitic and syenitic gneiss			
27	1160 ±4	Edwardsville	AM-87-5
28	1155 ±15	North Hammond	AM-87-3
HIGHLAND SAMPLES OF SILVER (1969)§§			
29	1113 ±10	Fayalite granite, Wanakena	
30	1109 ±11	Charnockite, Ticonderoga	
31	1084 ±15	Undeformed syenite dike, Jay	
32	1074 ±10	Anorthosite pegmatite, Elizabethtown	
33	1064 ±10	Metanorite, Snowy Mountain dome	
34	1054 ±20	Sheared anorthosite pegmatite, Jay	
35	1009 ±10	Magnetite-ilmenite ore, Tahawus‡‡	

Note: Errors at two sigma.
*Minimum Pb/Pb age.
Data from Grant et al (1986).
§Contains zircon cores, >1113 Ma, air abraded.
‡Baddeleyite age of >1086 ±5 Ma from this sample.
**Contains baddeleyite >1109 Ma.
Monazite age of 1137 ±1 Ma.
§§Decay constants of Steiger and Jager (1977).
‡‡Location same as Sanford Lake (SL) in Figure 1.

Table 2.: Sm-Nd data (sample numbers in Table 1)

sample	L	Zircon age[1]	t_{DM}[2]
ADIRONDACK HIGHLANDS			
Tonalites			
1 :AM87-12	t	1329 ± 36	1403
2 :AM86-12	t	1307 ± 2	1366
3 :LDT	t	>1366	1380
AMCG granitoids			
5 :DIA	s	1155 ± 4	1430
6 :AM86-15	r	1147 ± 10	1495
7 :AC85-6	m	1134 ± 4	1345
9 :AM86-17	e	1156 ± 8	1436
10 :AM86-9	g	1150 ± 5	1346
Younger granitoids			
13 :AM86-6	gd	1098 ± 4	1314
15 :AM86-4	a	1075 ± 17	1576
(repeat)			
:SK2A	tr	c.1060	1330
(repeat)			1373
Metasediment			
:JMCL-1	p	>c.1330	2075
Gabbro			
21 :Ali-1	g	1144 ± 7	1331
ADIRONDACK LOWLANDS			
Wellesely Island			
23 :AM86-16	l	1415 ± 6	1440
Fish Creek			
25 :AM87-4	a	1236 ± 6	1210
:5/90-5	t		
Hyde School			
26 :AC85-5	a	1230 ± 33	1351
:HS3	t	1230 ± 33	1397
:HS4	t	1230 ± 33	1350
Governeur			
24 :AC85-5	a	1284 ± 7	1525
ELZEVIR TERRANE			
Northbrook			
9/88-9	t	1250	1245
Elzevir			
9/88-10	t	1275	1397

1: U-Pb zircon ages in Ma from McLelland and Chiarenzelli (1990a,b) and Grant et al. (1986); 2: Sm-Nd model ages in Ma (DePaolo 1981) from Daly and McLelland (1991) for the Highlands and McLelland, Daly and Perham (1991) for the Lowlands; L: lithologies, a=alaskite, e=enderbite, g=granite, gd=granodiorite, m=mangerite, p=pelite, s=syenite, t=tonalite, tr=trondhjemite, l=leucogranite, initial digits of sample numbers refer to localities in Fig. 2.

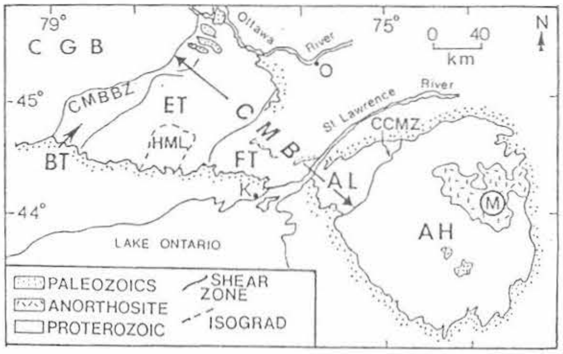


Fig. 3. Southwestern Grenville Province. CMB=Central Metasedimentary Belt, CGB=Central Gneiss Belt, BT=Bancroft Terrane, ET=Elzevir Terrane, FT=Frontenac Terrane, AL=Adirondack Lowlands, HL=Adirondack Highlands, HML=Hastings metamorphic low, K=Kingston, O=Ottawa, CCMZ=Carthage-Colton Mylonite Zone, M=Marcy massif.

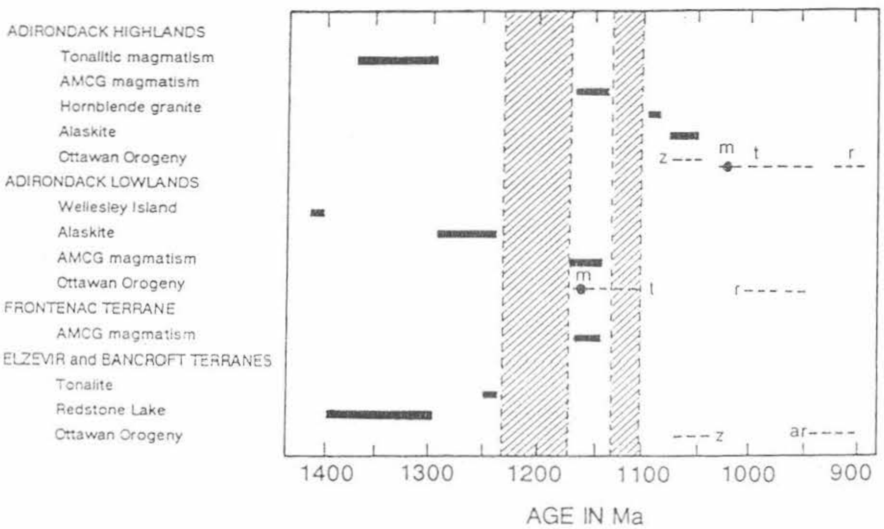


Fig. 4. Chronology of major geological events in the southwestern Grenville Province. z=zircon, t=titanite, m=monazite, r=rutile, ar=Ar/Ar. Diagonal ruling=quiescence. From McLelland and Chiarenzelli (1991).

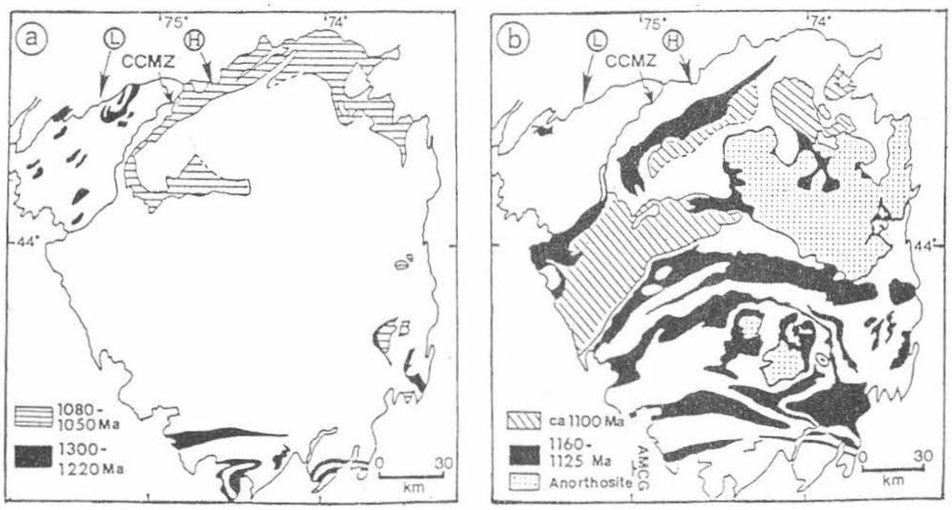


Fig. 5. Chronological designation of Adirondack units. L=Adirondack Lowlands, H=Adirondack Highlands, CCMZ=Carthage Colton Mylonite Zone. From Chiarenzelli and McLelland (1991).

represented by 1300-1220 Ma tonalites and alaskites whose distribution is shown in figure 5. The apparent absence of this suite from the central Highlands is believed to be the combined result of later magmatic intrusion and recent doming long a NNE axis.

Shown in figure 6 are paleoisotherms determined by Bohlen et al. (1985) and zircon ages divided according to origin and degree of disturbance. Note that the locus of disturbed ages corresponds with peak paleotemperatures. This result is discussed again in the metamorphic section. Within the Frontenac-Adirondack region, the Elzevir Orogeny was followed by 40-50 Ma of quiescence terminated at 1170-1130 Ma by voluminous anorogenic (fig. 4) magmatism referred to as the anorthosite-mangerite-charnockite-granite (AMCG) suite. The older ages are characteristic of AMCG magmatism in the Frontenac Terrane (including the Lowlands) while the Highlands commonly exhibit ages of 1150-1130 Ma (fig. 5). The large Marcy anorthosite massif (fig. 2) and its associated granitoid envelope have been shown to have an emplacement age of ca. 1135 Ma (McLelland and Chiarenzelli 1990). These ages are similar to those determined (Emslie and Hunt 1990) for the Morin, Lac St. Jean, and several other large massifs farther northeast in the Grenville Province (fig. 1). Rocks of similar age and chemistry (i.e., Storm King Granite) have been described within the Hudson Highlands (Grauch and Aleinikoff 1985). The extremely large dimensions of the AMCG magmatic terrane emphasize its global-scale nature corresponding, perhaps, to supercontinent rifting with the rifting axis located farther to the east. Valley (1985), McLelland and Husain (1986), and McLelland et. al. (1991a,b) have provided evidence that contact, and perhaps also regional, metamorphism accompanied emplacement of hot (~1100°C, Bohlen and Essene 1978), hypersolvus AMCG magmas. Wollastonite and monticellite occurrences related to thermal pulses from AMCG intrusions occur in proximity to AMCG intrusions (Valley and Essene 1980). In the Lowlands and the Canadian sector of the Frontenac Terrane monazite (table 1., no. 28), titanite (Rawnsley et al. 1987), and garnet ages (Mezger 1990) all indicate high temperatures (~600-800°C) at ca. 1150 Ma. Rutile ages and Rb/Sr whole rock isochron ages document temperatures not exceeding ~500 °C at ca. 1050-1000 Ma.

Following approximately 30 Ma of quiescence (Fig. 4), the Adirondacks, along with the entire Grenville Province, began to experience the onset of the Ottawan Orogeny of the Grenville Orogenic cycle (Moore and Thompson 1980). Initially the Ottawan Orogeny appears represented by 1090-1100 Ma hornblende granites in the northwest Highlands. These rather sparse granites were followed by deformation, high grade metamorphism, and the emplacement of trondhjemitic to alaskitic magnetite-rich rocks (Lyon Mt. Gneiss of Whitney and Olmstead 1988) in the northern and eastern Adirondacks. The zircon ages of these rocks fall into an interval of 1050-1080 Ma (table 1) which corresponds to the peak of granulite facies metamorphism when crust currently at the surface was at ~25 km. Accordingly, the alaskitic to trondhjemitic rocks are interpreted as synorogenic to late-orogenic intrusives. They were followed by the emplacement of single bodies of fayalite granite (ca. 1050 Ma) at Wanakena and Ausable Forks (fig. 2).

Sm-Nd analysis (Daly and McLelland 1991) demonstrates that the emplacement ages of the ca. 1300 Ma tonalitic rocks of the Highlands correspond closely to their neodymium model ages (table 1 and fig. 7a) indicating that these represent juvenile crustal additions. As seen in figure 7a, ϵ_{Nd} evolution curves for AMCG and younger granite suites pass within error of the tonalitic rocks and suggest that the tonalites, together with their own precursors (amphibolites?), served as source rocks for succeeding magmatic pulses. Remarkably, none of these igneous suites gives evidence for any pre-1600 Ma crust in the Adirondack region and the entire terrain appears to have come into existence in the Middle to Late Proterozoic. Significantly, Sm-Nd analysis for the ca. 1230-1300 Ma tonalitic to alaskitic Hyde School Gneiss (table 1, fig. 7b) demonstrates that it has model neodymium ages and ϵ_{Nd} values closely similar to Highland tonalites. The results are interpreted to reflect the contiguity of the Highlands and Lowlands at ca. 1300 Ma. Given this, the Carthage-Colton Mylonite Zone is interpreted as a west-dipping extensional normal fault that

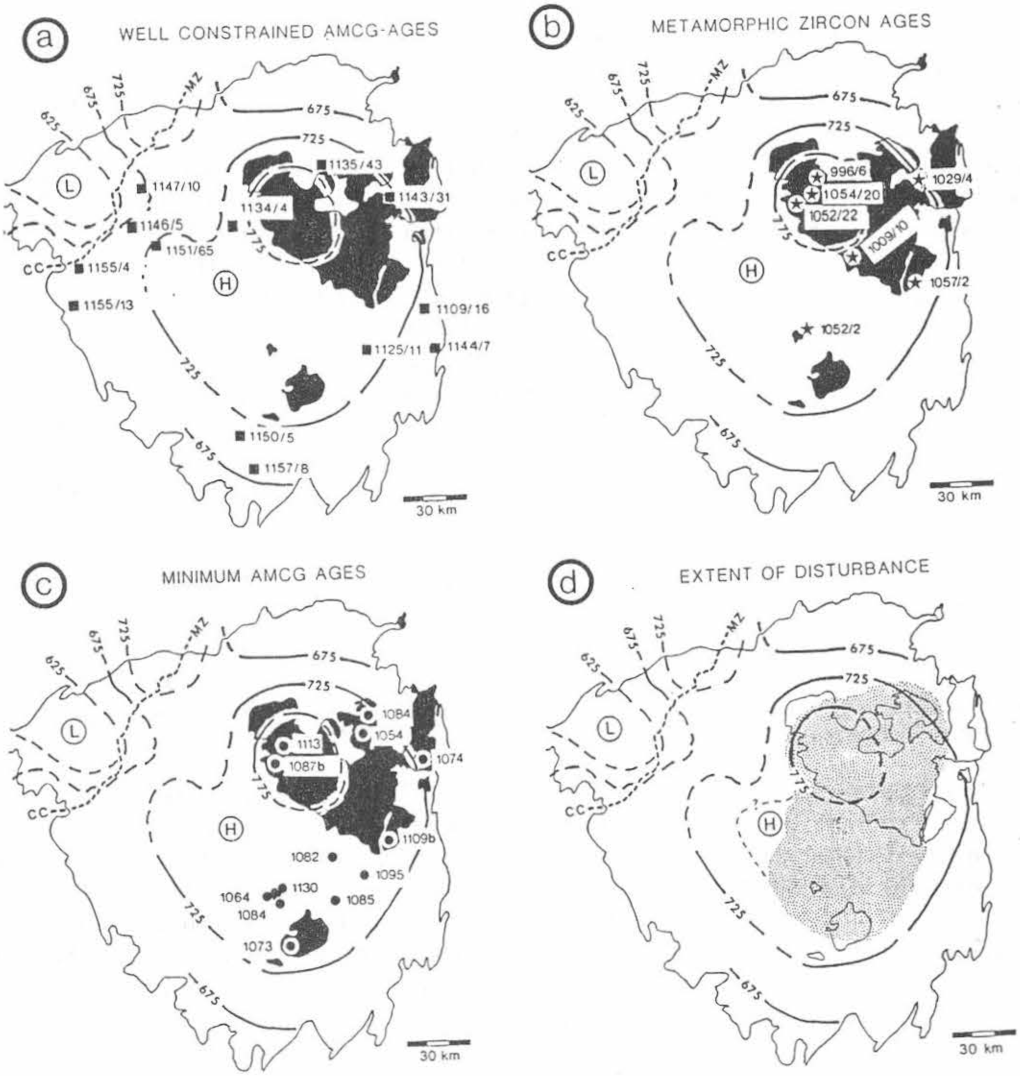


Fig. 6. Relationship between various U/Pb zircon ages and Adirondack paleoisotherms. Shaded area in (d) shows the extent of zircons whose U/Pb systematics have been disturbed. H=Highlands, L=Lowlands, CC=Carthage Colton Mylonite Zone. From Chiarenzelli and McLelland (1991).

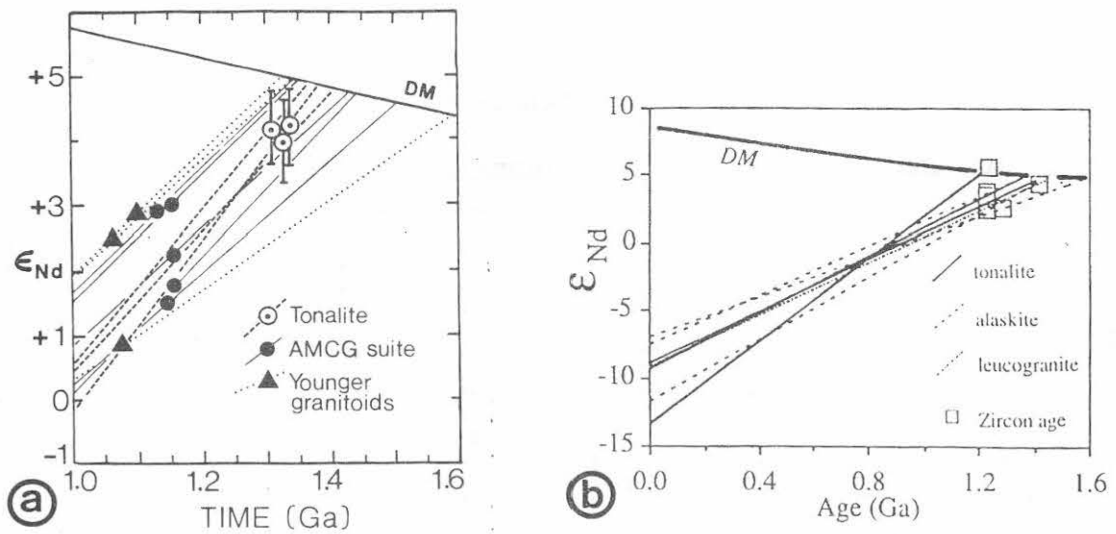


Fig. 7. ϵ_{Nd} evolution diagrams for (a) Adirondack highlands (Daly and McLelland 1991), (b) Adirondack lowlands (Hyde School Gneiss). U-Pb zircon ages are indicated by circles, triangles and squares (from table 1). DM=depleted mantle evolution curve (DePaolo 1981).

formed during the Ottawan Orogeny in response to crustal thickening by thrust stacking (Burchfiel and Royden 1985). East dipping extensional faults of this sort and age have been described by van der Pluijm and Carlson (1989) in the Central Metasedimentary Belt. Motion of this sort along the Carthage-Colton Mylonite Zone would help to explain the juxtaposition of amphibolite and granulite facies assemblages across the zone. A downward displacement of 3-4 km would satisfactorily explain the somewhat lower grade of the Lowlands terrane.

PETROLOGY CHARACTERISTICS OF THE PRINCIPAL ROCK TYPES IN THE SOUTHERN ADIRONDACKS

The following discussion is divided into igneous and metasedimentary sections. Only rock-types occurring within the southern Adirondacks are discussed.

Igneous Rocks

A) Tonalites and related granitoids. Typical whole rock chemistries for these rocks are given in table 3. Figure 8 shows the normative anorthite (An)-albite (Ab)-orthoclase (Or) data for these rocks and compares them to similar rocks in the Lowlands. AFM plots are given in figure 9 and calc-alkali index versus silica plots in figure 10; both figures illustrate the strongly calcalkaline nature of the Highland tonalite to granitoid suite.

The tonalitic rocks, which will be visited at Stop 4, outcrop in several E-W belts within the southern Adirondacks. in the field they can be distinguished from, otherwise similar, charnockitic rocks by the white alteration of their weathered surfaces and the bluish grey on fresh surfaces. A distinctive characteristic is the almost ubiquitous presence of discontinuous mafic sheets. These have been interpreted as disrupted mafic dikes coeval with emplacement of the tonalites.

Associated with the tonalitic rocks are granodioritic to granitic rocks containing variable concentrations of orthopyroxene. These are best represented by the Canada Lake Charnockite (Stop 3) and by the large Tomantown pluton (fig. 2) whose minimum emplacement age is 1184 Ma (table 1).

TABLE 3

	EARLY CALCALKALINE ROCKS			OLDER ANOROGENIC PLUTONIC ROCKS					
	AM-87-13	TOE	CL-6	AM-86-17	AM-86-1	AC-85-1	AM-86-9	AM-86-15	AC-85-2
SiO_2	65.00	65.68	74.63	68.90	71.88	73.72	69.14	67.47	75.17
TiO_2	0.75	1.16	0.37	0.59	0.36	0.04	0.89	0.72	0.20
Al_2O_3	15.10	14.97	14.22	14.50	14.82	13.54	13.78	15.12	12.63
FeO	Nd	Nd	Nd	2.16	1.27	0.87	2.83	3.34	1.11
Fe_2O_3	6.03	6.79	1.53	1.1	0.96	0.11	1.82	1.59	1.07
MnO	0.10	0.08	0.04	0.02	0.03	0.01	0.04	0.10	0.02
MgO	0.46	1.15	0.55	0.84	0.43	0.20	0.45	0.51	0.19
CaO	2.71	2.56	1.66	2.3	1.87	0.85	2.26	2.57	0.88
Na_2O	4.10	2.80	3.56	3.06	3.93	5.71	3.07	3.41	2.99
K_2O	5.13	4.52	4.26	4.18	3.99	4.42	4.91	5.18	5.49
P_2O_5	0.10	0.51	0.10	0.24	0.09	0.01	0.23	0.19	0.04
LOI	0.39	0.74		0.40	0.27	0.17	0.19	0.39	0.17
Σ	99.87	100.96	99.42	99.63	99.65	99.61	100.59	99.96	
Ba(ppm)	1230	710	510	680	660	1100	736	810	442
Rb (ppm)	100	97	170	160	200	406	81	128	230
Sr (ppm)	260	230	260	200	130	26	211	215	99
Y (ppm)	70	70	37	40	110	321	60	71	77
Nb (ppm)	20	19	17	30	30	15	19	21	15
Zr (ppm)	790	345	160	270	670	118	538	546	284

YOUNGER ANOROGENIC PLUTONIC ROCKS

	AC-85-6	AC-85-10	AM-86-7	WPG	SLC	AM-87-9	AM-86-8	AM-87-10
SiO_2	62.15	54.94	58.50	69.20	60.64	61.05	60.94	62.14
TiO_2	0.88	1.55	0.66	0.51	1.14	0.78	1.39	0.36
Al_2O_3	16.40	14.87	20.32	13.90	15.27	15.98	15.91	12.35
FeO	3.96	10.25	2.81	3.1	9.28	4.60	6.51	10.32
Fe_2O_3	1.49	2.80	0.43	1.34	1.77	2.10	1.02	1.7
MnO	0.09	0.24	0.01	0.05	0.19	0.05	0.14	0.01
MgO	1.06	0.96	1.47	0.52	0.74	1.64	1.70	0.83
CaO	3.27	5.52	6.16	2.03	3.97	3.63	4.53	3.65
Na_2O	4.81	3.45	5.02	3.02	3.34	3.41	3.55	6.05
K_2O	5.13	3.83	3.35	5.48	3.70	4.76	3.86	1.26
P_2O_5	0.30	0.65	0.32	0.11	0.31	0.42	0.46	0.09
LOI	0.41	0.37	0.43	0.39	0.01	0.91	0.37	0.67
Σ	99.95	99.43	99.50	99.65	100.46	99.63	100.38	99.24
Ba(ppm)	850	625	Nd	1279	823	Nd	1100	Nd
Rb (ppm)	106	47	Nd	124	87	Nd	83	29
Sr (ppm)	335	367	Nd	184	215	Nd	410	180
Y (ppm)	60	55	Nd	48	121	Nd	110	63
Nb (ppm)	21	14	Nd	14	38	Nd	25	79
Zr (ppm)	464	431	Nd	382	647	Nd	1200	309

	YOUNGER GRANITIC ROCKS						LATE LEUCOGRANITIC ROCKS				
	AM-86-3	AM-86-6	NO-Fo1	AM-86-13	AM-87-6	GHA	AM-86-11	AM-87-7	AM-86-4	AM-86-10	AM-86-14
SiO_2	68.62	68.05	71.75	76.30	69.00	73.2	69.98	67.80	70.01	69.05	72.39
TiO_2	0.48	0.55	0.41	0.18	1.43	0.35	0.46	0.42	0.69	0.57	0.38
Al_2O_3	14.37	14.67	13.49	11.64	12.12	13.1	12.37	15.76	12.43	13.06	12.63
FeO	3.01	3.51	2.39	1.13	4.94	0.84	5.13	1.2	4.23	3.50	1.6
Fe_2O_3	0.93	1.18	1.12	0.61	1.16	2.1	1.11	2.8	1.5	1.42	4.13
MnO	0.06	0.07	0.05	0.01	0.02	0.02	0.14	0.03	0.03	0.01	0.03
MgO	0.49	0.45	0.11	0.01	0.67	0.33	0.08	0.56	0.01	0.17	0.29
CaO	1.99	1.81	1.43	0.45	0.75	1.55	1.25	2.35	0.94	0.35	1.07
Na_2O	3.71	3.81	2.99	3.32	2.79	4.16	3.99	3.71	1.92	1.08	6.63
K_2O	5.67	5.61	5.79	5.22	6.50	4.01	4.91	4.91	8.34	9.64	0.52
P_2O_5	0.13	0.12	0.07	0.03	0.17	0.07	0.04	0.13	0.17	0.12	0.70
LOI	0.40	0.58	0.5	0.30	0.25	0.39	0.21	0.31	0.23	0.41	0.10
Σ	99.86	100.41	100.00	99.20	99.80	100.3	99.67	99.98	100.50	99.38	100.47
Ba(ppm)	861	715	692	Nd	1014	1249	160	Nd	840	290	98
Rb (ppm)	161	148	182	188	214	178	190	Nd	315	330	16
Sr (ppm)	209	240	115	132	303	211	20	Nd	73	60	42
Y (ppm)	65	62	72	157	35	86	120	Nd	66	75	117
Nb (ppm)	20	20	25	29	17	19	30	Nd	18	23	27
Zr (ppm)	394	542	595	392	507	338	1230	Nd	414	600	786

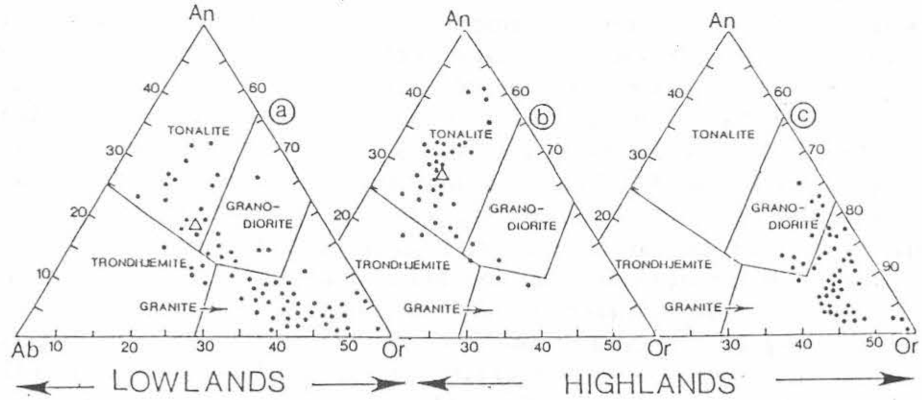


Fig. 8. Plots of normative albite (Ab)-anorthite(An)-orthoclase (Or) for (a) Hyde School Gneiss, (b) Highlands tonalites, and (c) Tomantown pluton. Open triangles give average values for tonalitic samples. Definition of fields due to Barker (1979).

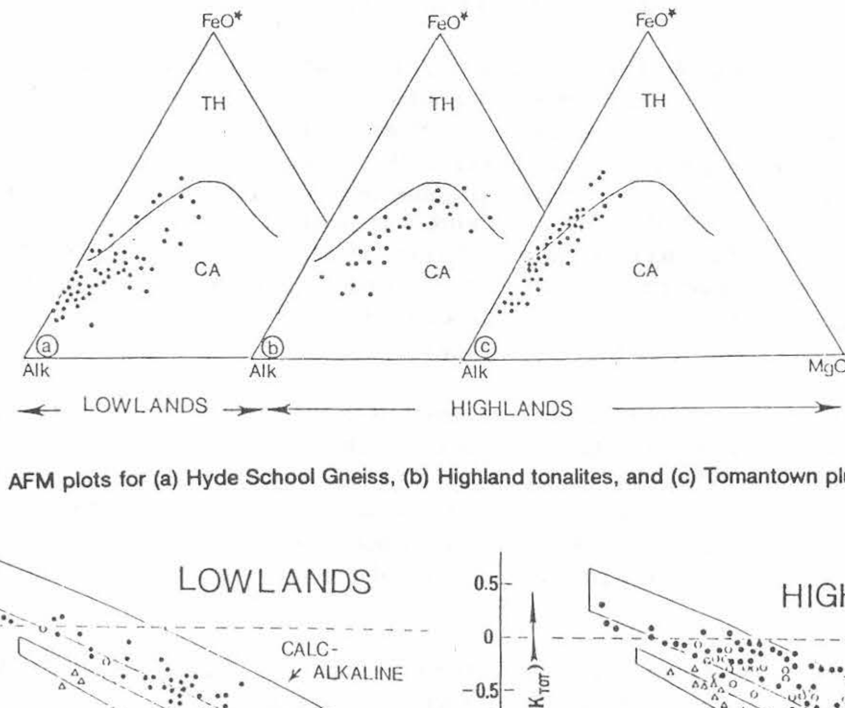


Fig. 9. AFM plots for (a) Hyde School Gneiss, (b) Highland tonalites, and (c) Tomantown pluton.

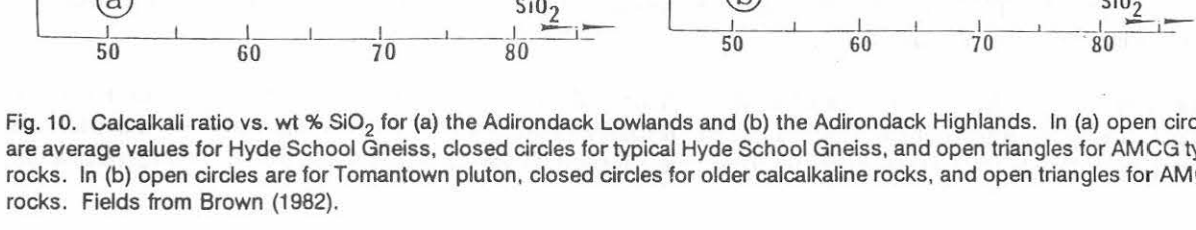


Fig. 10. Calcalkali ratio vs. wt % SiO_2 for (a) the Adirondack Lowlands and (b) the Adirondack Highlands. In (a) open circles are average values for Hyde School Gneiss, closed circles for typical Hyde School Gneiss, and open triangles for AMCG type rocks. In (b) open circles are for Tomantown pluton, closed circles for older calcalkaline rocks, and open triangles for AMCG rocks. Fields from Brown (1982).

McLelland et al. (1991b) have interpreted the early calcalkaline rocks of the Highlands as correlative with the Hyde School Gneiss of the Adirondack Lowlands (fig. 2). This interpretation is consistent with the Sm-Nd results (table 1b) discussed previously and shown in figure 7.

B) AMCG Suite. Within the southern Adirondacks AMCG rocks are widely developed and abundantly represented in the Piseco anticline (Stop 6) as well as the Oregon (Stop 8) and Snowy Mt. Domes. The chemistry of granitoid (mangeritic to charnockitic varieties of these rocks is given in table 3, especially for the older anorogenic plutonic rocks to which southern Adirondack suites belong. As shown in figure 10, the AMCG rocks have calcalkali-silica trends that are distinctly different than those shown by the tonalitic suites. McLelland (1991) and McLelland and Whitney (1991) have shown that the AMCG rocks exhibit anorogenic geochemical characteristics (figs. 11, 12, 13) and also constitute bimodal magmatic complexes in which anorthositic to gabbroic cores are coeval with, but not related via fractional crystallization to the mangeritic-charnockitic envelopes of the AMCG massifs (i.e., Marcy massif, fig. 2). Bimodality is best demonstrated by noting the divergent differentiation trends of the granitoid members on the one hand and the anorthositic-gabbroic rocks on the other (Buddington 1972). This divergence is nicely exhibited by the variation of the FeO-MgO ratio with wt.% SiO_2 (fig. 12) Ga-Al_2O_3 trends (fig. 13), and by Harker variation diagrams for AMCG rocks of the Marcy massif (fig. 14) (McLelland 1989). The extreme low-SiO_2, high-iron end members (fig. 14) of the anorthosite-gabbro family will be seen at Stop 8 and are believed to represent late liquids developed under conditions of low oxygen fugacites (i.e., dry, Fenner-type trends).

C) Metasedimentary Rocks. Within the southern Adirondacks the metasedimentary sequence is dominated by quartzites and metapelites with marbles being virtually absent. The quartzites are exceptionally thick and pure and comprise an ~1000 m-thick unit referred to as the Irving Pond Quartzite (Stop 2). Of even greater extent, as well as thickness, is the Peck Lake Formation which consists of garnet-biotite-quartz-oligoclase ± sillimanite gneiss (referred to as kinzigite) together with sheets, pods, and stingers of white, minimum melt granite that commonly contains garnets (Stop 1). McLelland and Husain (1986) interpreted the kinzigites and their leucosomes as restite-anatectite pairs and attributed partial melting to heating accompanying AMCG magmatism. It is now believed that an additional period of anatexis probably preceded the 1130-1150 Ma AMCG magmatism during the 1300-1220 Elzevir Orogeny.

The occurrence of anatexis within the kinzigites is corroborated by the presence of sparse hercynitic spinel within either garnets or sillimanite-rich wisps in leucosomes. McLelland et al. (1991a) have shown that extraction of anatectic material from the least altered kinzigites can satisfactorily account for the composition of more aluminous, lower-silica kinzigites. The ultimate evolution of this process would be to produce assemblages of aluminous sillimanite-garnet-biotite gneiss together with granitic material of the sort that characterizes the Sacandaga Formation (Stop 9).

Based on the bulk chemistry of kinzigites in the southern Adirondacks, McLelland and Husain (1986) interpreted their protoliths as Proterozoic greywackes and shales. More recently, McLelland et al. (1991b) have provided evidence to support the conclusion that the Peck Lake Fm. kinzigites of the southern Adirondacks can be correlated with the markedly similar Major Paragneiss of the Adirondack Lowlands (bqpg on fig. 2). McLelland and Isachsen (1986) have also argued that the Peck Lake Fm., and associated rocks, continues eastward into the eastern Adirondacks in the vicinity of Lake George.

In contrast to the southern and eastern Adirondacks, the central Adirondacks contain only sparse kinzigite, and metasediments are principally represented by synclinal keels of marble and calcsilicate (Stop 7). It is possible that the change from carbonate to pelitic metasediments

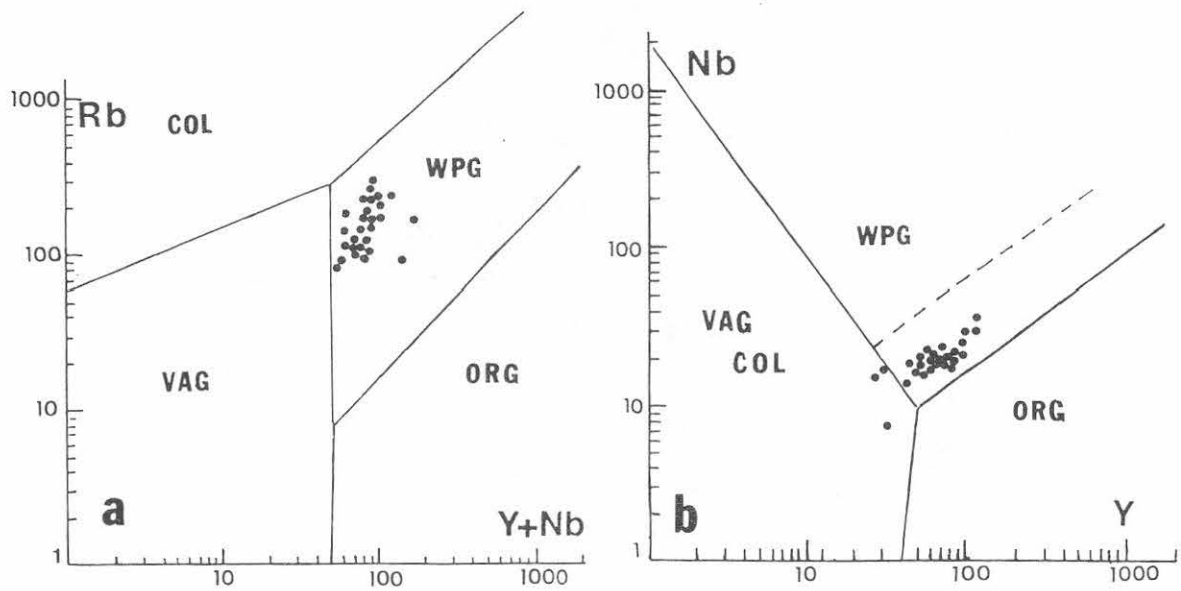
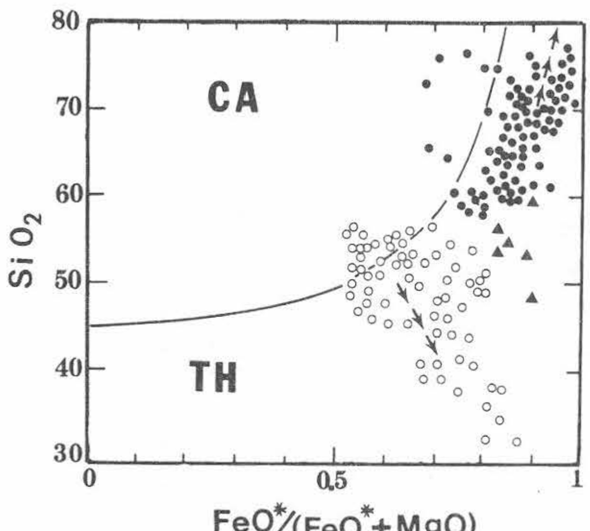


Fig. 11. Tectonic discrimination diagrams (Pearce et al. 1984) for AMCG granitoids from the Marcy massif. COL=collisional, WPG=within plate granites, ORG=-ocean ridge granites, VAG=volcanic arc granites.



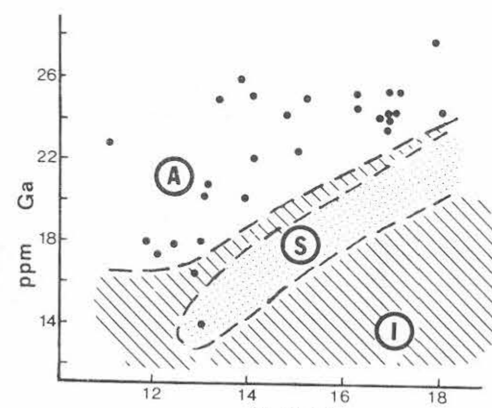


Fig. 12. FeO*-MgO vs. wt.% SiO_2 variation for anorthositic suite (open circles) and mangeritic-charnockite suite (filled circles). Triangles designate mixed rocks at contacts. Arrows indicate differentiation trends of the two suites, CA=calcalkaline, TH=tholeiitic. (after Anderson 1983)

Fig. 13. Ga vs. Wt.% Al_2O_3 for AMCG suite granitoids of the Marcy massif. Fields of A-,S-, and I-type granites and shown (after White and Chappell 1983).

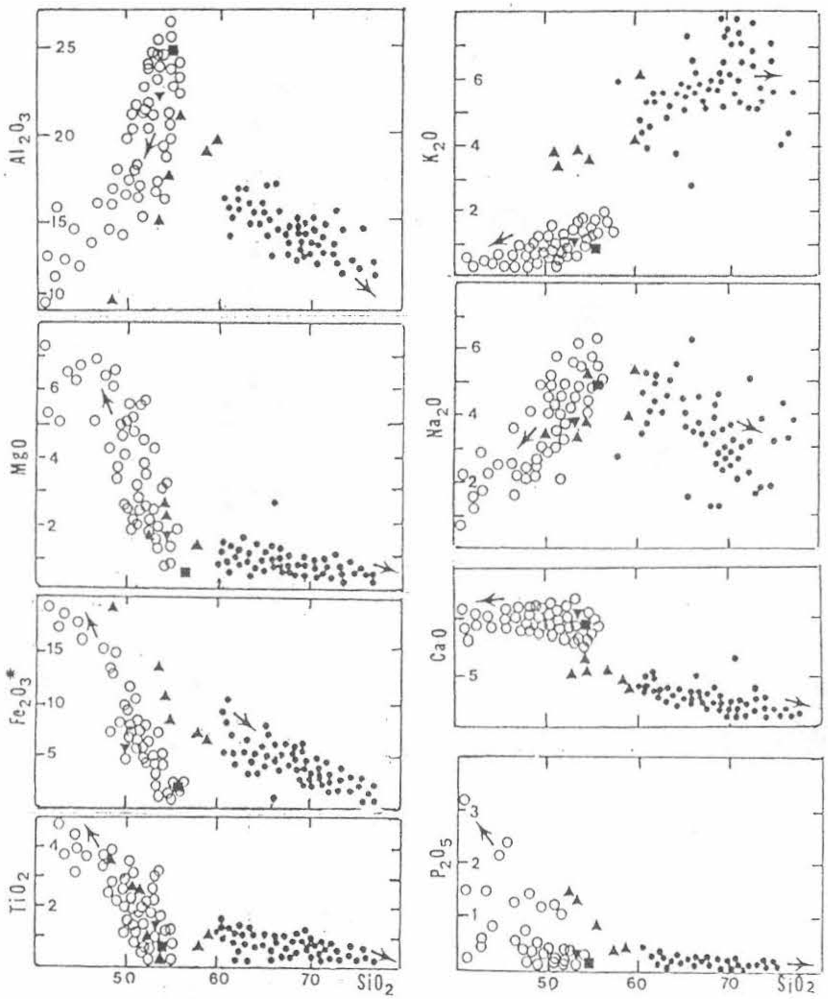


Fig. 14. Harker variation diagrams or AMCG-rocks of the Marcy massif. Open circles=anorthositic suite, filled circles=granitoid suite, upright triangles=mixed rocks, inverted triangles=Whiteface facies, square=Marcy facies. Arrows indicate differentiation trends.

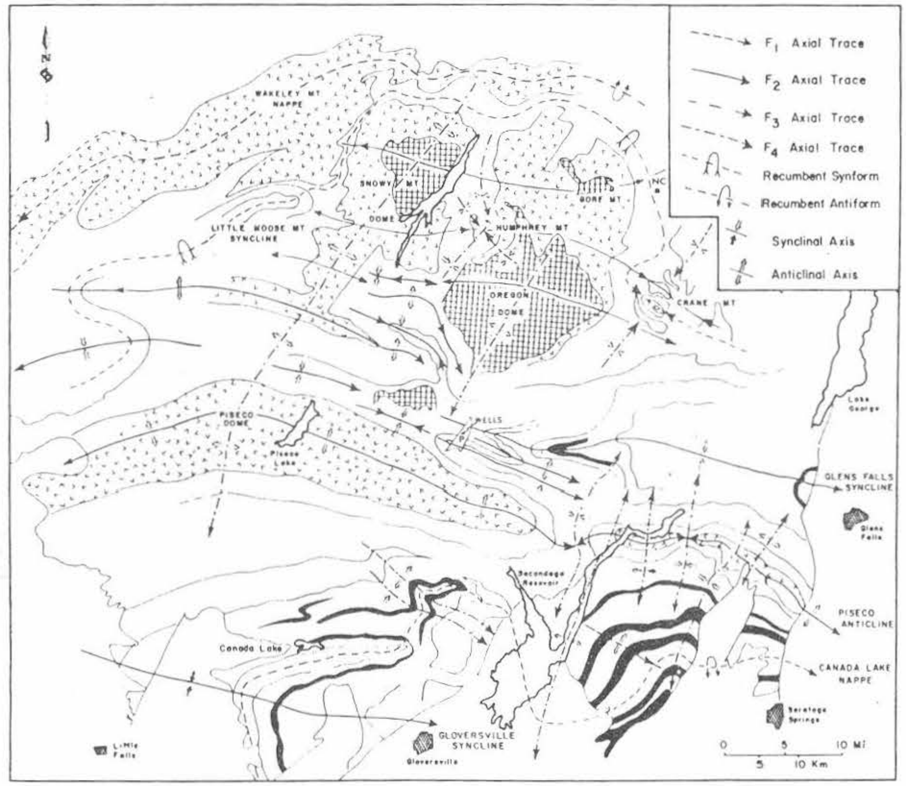


Fig. 15. Fold axes within the southern and central Adirondacks. Designation of folds as synclines and anticlines is provisional, since facing directions are not yet known.

corresponds to an original shelf to deep water transition, now largely removed by later intrusion, doming, and erosion. The Irving Pond quartzite may represent a siliceous clastic cap closing out the earlier deep water basin.

A single specimen of metapelite (no. 21, table 2) has yielded a T_{DM} of 2075 Ma. This model age approximates the time at which source rocks for the metasediment separated from the mantle. Although the age may be the result of mixing rocks >2075 Ma with younger components, the older material clearly predates any possible Adirondack sources.

STRUCTURAL GEOLOGY

The southern Adirondacks is an area of intense ductile strain, essentially all of which must postdate the ca. 1150 Ma AMCG rocks which are involved in each of the major phases of deformation, i.e., the regional strain is associated with the Ottawan Orogeny.

As shown in figures 2 and 15, the southern Adirondacks are underlain by very large folds. Four major phases of folding can be identified and their intersections produce the characteristic fold interference outcrop patterns of the region (fig. 16).

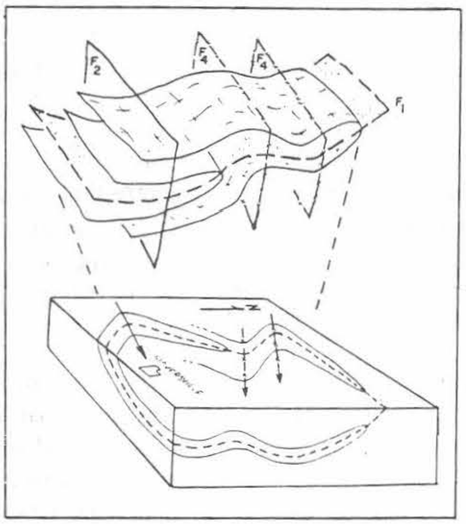


Fig. 16. Block diagram showing how interference between F_1, F_2, and F_4 fold sets produce the outcrop pattern of the F_1 Canada Lake isocline. The axial plane of the F_1 fold is stippled and its fold axis plunges 10-15° to the southeast. The city of Gloversville is shown.

The earliest recognizable map-scale folds (F_1) are exceptionally large isoclinal recumbent structures characterized by the Canada Lake, Little Moose Mt., and Wakely Mt. isoclines, whose axes trend E-W and plunge 10-15° about the horizontal. The Little Moose Mt. isocline is synformal (deWaard 1964) and the other two are antiformal, and suspected to be anticlinal, but the lack of stratigraphic facing directions precludes any certain age assignments although these are designated in figure 15 on a provisional basis. All of these structures fold an earlier tectonic foliation consisting of flattened mineral grains of unknown age and origin. An axial planar cleavage is well developed in the Canada Lake isocline, particularly in the metapelitic rocks.

F_2-folds of exceptionally large dimensions trend E-W across the region and have upright axial planes (fig. 15). They are coaxial with the F_1 folds suggesting that the earlier fold axes have been rotated into parallelism with F_2 and that the current configurations of both fold sets may be the result of a common set of forces. An intense ribbon lineation defined by quartz and feldspar rods parallels the F_2-axes along the Piseco anticline, Gloversville syncline, and Glens Falls syncline and documents the high temperatures, ductile deformation and mylonitization that accompanied the formation of these folds.

Large NNE trending upright folds (F_3) define the Snowy Mt. and Oregon domes (fig. 15). Where the F_3 folds intersect F_2 axes structural domes (i.e., Piseco dome) and intervening saddles result. A late NW-trending fold set results in a few F_4 folds between Canada Lake and Sacandaga Reservoir (fig. 15).

Kinematic indicators (mostly feldspar tails) in the area suggest that the dominant displacement involved motion in which the east side moved up and to the west (McLelland 1984). In most instances this implies thrusting motion, however, displacement in the opposite sense has also been documented. This suggests that relative displacement may have taken place in both senses during formation of the indicators. A movement picture consistent with this is still under investigation, although regional extension analogous to that in core complexes might resolve the situation.

METAMORPHISM

Figure 6 shows the well known pattern of paleoisotherms established by Bohlen and Essene (1977) and updated in Bohlen et al. (1985). Paleotemperatures have been established largely on the basis of two-feldspar geothermometry but (Fe, Ti)-oxide methods have also been used and, locally, temperature-restrictive mineral assemblages have been employed (Valley 1985). The bull's eye pattern of paleoisotherms, centering on the Marcy massif, is believed to the due to late doming centered on the massif. Paleopressures show a similar bull's eye configuration with pressures of 7-8 kbar decreasing outward to 6-7 kbar away from the massif and reaching 5-6 kbar in the Lowlands (Bohlen et al. 1985).

Bohlen et al. (1985) interpret the paleotemperature pattern of figure 6 as representative of peak metamorphic temperatures in the Adirondacks, and paleopressures are interpreted similarly. Chiarenzelli and McLelland (1991) show that disturbance of U-Pb systematics in zircons corresponds with Bohlen et al.'s (1985) paleoisotherms (fig. 6), and this correlation strengthens the conclusion that the pattern is one of peak temperatures rather than a retrograde set frozen in from a terrane of uniform temperatures in the range ~750°-800°C.

The P,T conditions of the Adirondack are those of granulite facies metamorphism, and for the most part conditions correspond to the hornblende-clinopyroxene-almandine subfacies of the high-pressure portion of the granulite facies. These conditions must have been imposed during the Ottawan Orogeny in order to have affected rocks as young as 1050 Ma. The identification of ca. 1050-1060 Ma metamorphic zircons by McLelland and Chiarenzelli (1990) fixes the time of peak metamorphic conditions and corresponds well with titanite and garnet U-Pb ages of ca. 1030-1000 Ma in the Highlands (Mezger 1990). Rb-Sr whole rock isochron ages of ca. 1100-1000 Ma also reflect Ottawan temperatures and fluids. Despite the high-grade, regional character of the Ottawan Orogeny, the preservation of foliated garnet-sillimanite xenoliths in an 1147±4 Ma metagabbro (McLelland et al. 1987a), and the report of some 1150 Ma U-Pb garnet ages (Mezger 1990), reveals that earlier assemblages from the Elzevirian and AMCG metamorphic pulses managed to survive locally. The dehydrating effects of these high temperature events, as well as the anhydrous nature of the AMCG rocks themselves, are thought to be responsible for creating a water-poor terrane throughout the Adirondack Highlands prior to the Ottawan Orogeny.

The present day depth to the Moho beneath the Adirondack Highlands is ~35 km (Katz 1955). Since metamorphic pressures of 7-8 kbar correspond to ~20-25 km depth of burial, it follows that during metamorphism the Adirondack region consisted of a double thickness of continental crust. Present day examples of doubly thickened continental crust are found in continent-continent collisional margins such as the Himalayas or Andean margins such as along the coast of South America. The latter model is not readily applicable to the Ottawan-age Adirondacks, because of the absence of calcalkaline magmatism of that age. On the other hand, the Himalayan-Tibetan analogue provides a strikingly consistent model, including the rather limited amount of associated magmatism. Because no suggestion of a suture exists between the Green Mts. of Vermont and the Grenville Tectonic Front, and because of the dominance of tectonic vergence to the northwest throughout the region, the Ottawan plate margin has been placed east of the Grenville inliers of the Appalachians and assigned an eastward dip. Although highly speculative, this possibility, together with other plate tectonic reconstructions are shown in figure 17.

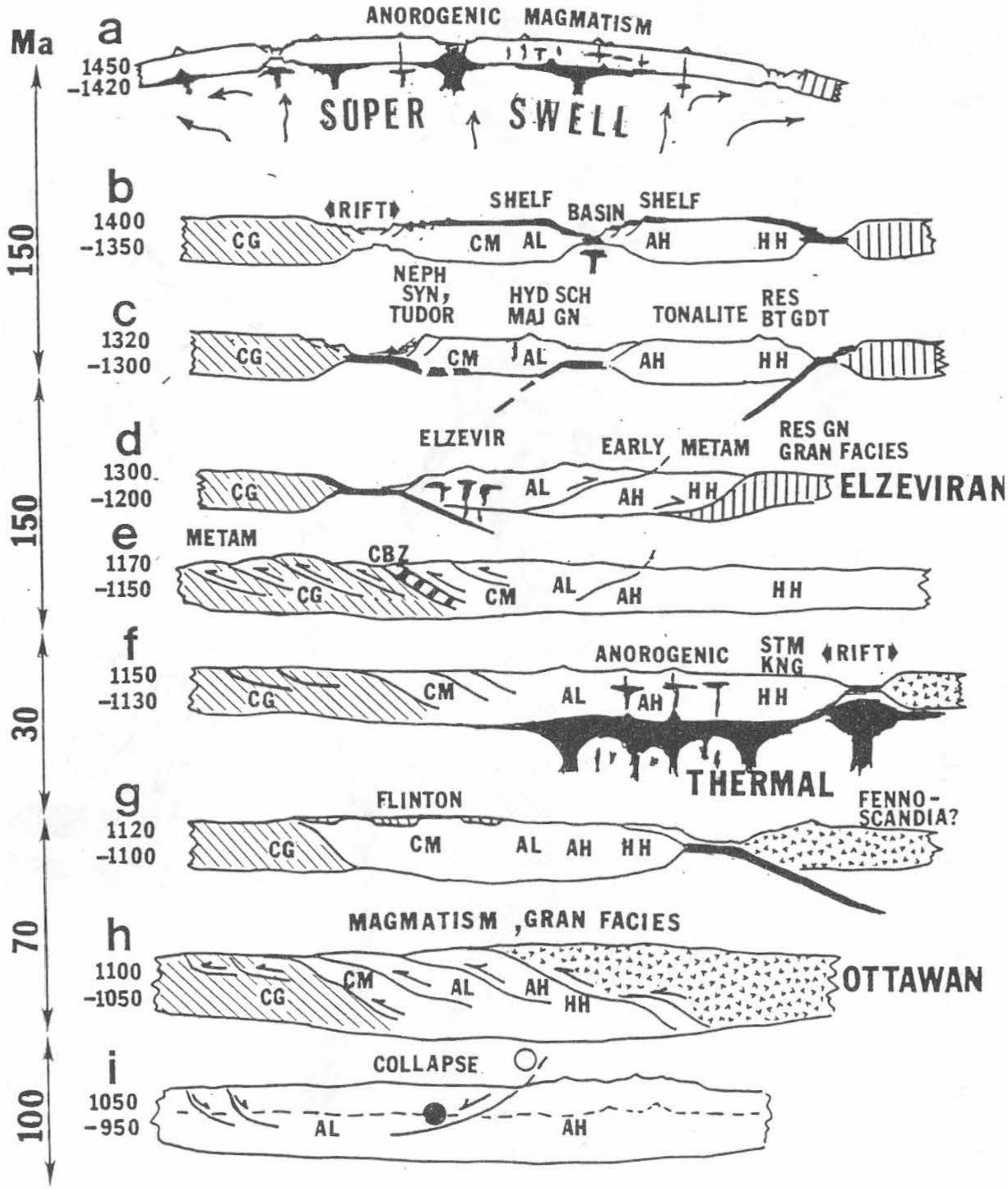


Fig. 17. Hypothetical plate tectonic scenarios for the southwestern Grenville Province.

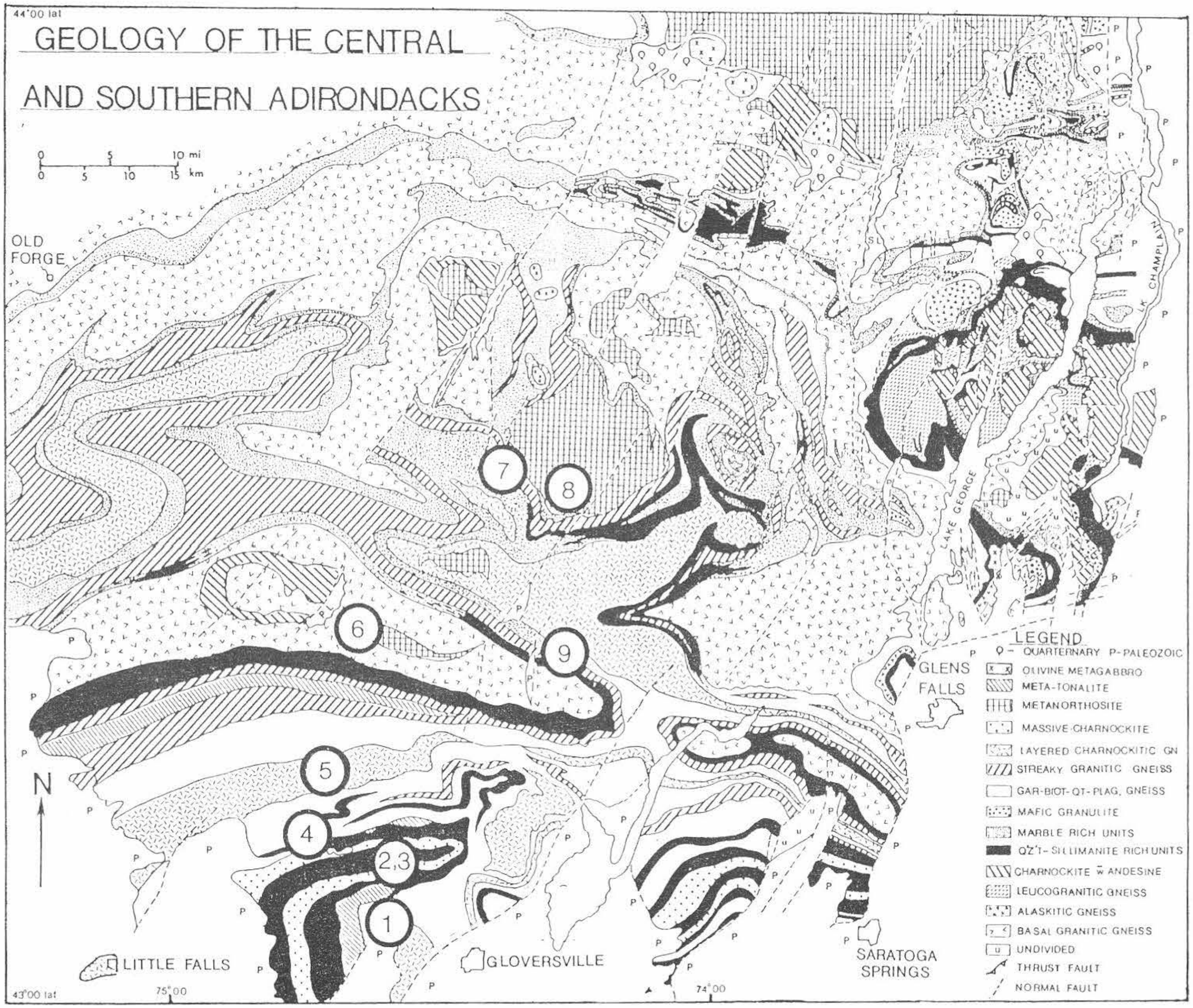


Fig. 18. Geologic map of the southern and central Adirondacks with field trip stops 1-9 indicated (McLelland and Isachsen 1986).

ROAD LOG

(See fig. 18 for stop locations)

CUMULATIVE MILEAGE	MILES FROM LAST POINT	ROUTE DESCRIPTION
	0	Junction of Willie Road, Peck Hill Road, and NY Rt. 29A
1.3	1.3	Mud Lake to northeast of NY Rt. 29A
1.8	1.5	Peck Lake to Northeast of NY Rt. 29A
3.6	1.8	STOP 1. Peck Lake Fm.

STOP 1.

This exposure along Rt. 29A just north of Peck Lake is the type locality of the sillimanite-garnet-biotite-quartz-feldspar gneisses (kinzigites) of the Peck Lake Fm. in addition, there are exposed excellent minor folds of several generations. Note that the F_1 folds rotate an earlier foliation. The white quartzo-feldspathic layers in the kinzigites consist of quartz, two feldspars,and garnet and ar believed to be anatectic and have been folded by F_1 indicating pre-F_1 metamorphic events. Typical whole rock compositions are shown below. Spinel has been found enclosed in garnets at this outcrop. The similarity of the Peck Lake Fm. to the Major Paragneiss of the Lowlands suggests that the Adirondacks were contiguous at the time of deposition of the rocks.

Table 4.

COMPOSITIONS OF REPRESENTATIVE LEUCOSOME AND HOST ROCK — SELECTED CLASTICS

	Leucosome		Host Rock		Average Greywacke[a] (Σ = 23)	Average PC Slate[b] (Σ = 33)	Average Slate[c] (Σ = 36)
	LL1	9-17-2A	10-29-1B	9-11-4B			
SiO_2	75.61	74.60	68.04	64.24	64.70	56.30	60.64
Al_2O_3	13.75	13.49	13.93	16.16	14.80	17.24	17.32
TiO_2	.02	.19	.86	.90	.50	.77	.73
Fe_2O_3	.51	1.47	6.08	7.44	4.10	7.22	4.81
MgO	.11	.54	1.45	1.57	2.20	2.54	2.60
CaO	.36	1.64	1.65	3.41	3.10	1.00	1.20
Na_2O	2.19	3.25	2.84	3.20	3.10	1.23	1.20
K_2O	6.82	4.69	3.27	2.92	1.90	3.79	3.69
MnO	.02	.04	.06	.09	.10	.10	...
P_2O_5	.09	.08	.18	.17	.20	.14	...
LOI	.31	.25	.66	.66	2.40	3.70	4.10
TOTAL	99.78	100.24	99.80	99.76	101.00	98.70	98.00

6.1	2.5	Junction NY Rt. 29A and NY Rt. 10
8.0	1.9	Nick Stoner's Inn on west side of NY Rt. 29A-10
8.6	.6	STOP 2. Irving Pond Fm., .5 mile north of Nick Stoner's Inn, Canada Lake. Very near hinge line of F_1 Canada Lake isocline.

STOP 2.

The outer portion of the Irving Pond Fm. is exposed in low cuts along the east side of Rt. 29A just prior to the crest in the road heading north.

At the southern end of the cut typical, massive quartzites of the Irving Pond are seen. Proceeding north the quartzites become "dirtier" until they develop sillimanite-garnet-biotite-feldspar (kinzigites) layers along with quartzite.

At the northern end of the cut, and approximately on the Irving Pond/Canada Lake Fm. contact there occurs an excellent set of F_1 minor folds. Polished slabs and thin sections demonstrate that these fold an earlier foliation defined by biotite flakes and flattened quartz grains.

At the southern end of the outcrop dark, fine grained metadiabase sheets crosscut the quartzite. Near the telephone pole erosional remnants of diabase appear to truncate approximately horizontal foliation in the quartzite suggesting that the diabase was emplaced after an early metamorphism. At the north end of the cut a diabase sheet of variable thickness is folded in the F_1 fold. The folding is interpreted as Ottawan, the diabase as AMCG in origin, and the early foliation as Elzevirian. This is consistent with the presence of quartzite xenoliths in the ca. 1300 Ma tonalites.

The Irving Pond Fm. is the uppermost unit in the lithotectonic sequence of the southern Adirondacks. Its present thickness is close to 1000 meters, and it is exposed across strike for approximately 4000 meters. Throughout this section massive quartzites dominate.

8.8	.2	STOP 3. Canada Lake Charnockite (>1233 Ma, table 1, sample AM-87-13. Now fixed at 1251±43)

STOP 3.

Large roadcuts expose the type section of the Canada Lake charnockite. Lithogically the charnockite consists of 20-30% quartz, 40-50% mesoperthite, 20-30% oligoclase, and 5-10% mafics. The occurrence of orthopyroxene is sporadic. These exposures exhibit the olive-drab coloration that is typical of charnockites. Note the strong foliation in the rock. Farther north along the highway there are exposed pink leucogranitic variants of this unit. The chemical composition of these is given in table 3 (ab-6). The whole rock chemistry of the charnockitic phase is similar to AM-86-17 in table 3. The lateral continuity of the Canada Lake is striking (fig. 2) but the presence of xenoliths reveals an intrusive origin.

10	1.2	STOP 4. Royal Mt. Tonalite (>1301 Ma, table 1, sample AM-86-12, now fixed at 1307±2 Ma).

STOP 4.

Steep roadcuts, exposed across from the Canada Lake Store, expose typical examples of the early tonalitic rocks that occur within the southern and eastern Adirondacks and that manifest the presence, throughout the region, of collisional magmatic arcs of calcalkaline chemistry that existed along the eastern margin of Laurentia from ca. 1400-1200 Ma. Amalgamation of these arcs culminated in the Elzevirian Orogeny at ca. 1250-1220 Ma.

The whole rock chemistry of the tonalitic rocks is given in table 3 and important chemical trends are portrayed in figures 8, 9, and 10. Figure 7 shows the ϵ_{Nd} characteristics of these rocks and emphasizes their petrologically juvenile character, i.e., they are not derived from any crustal rocks with long-term crustal residence but are essentially mantle derived (including derivation from melting of basaltic rock derived from the mantle at ca. 1300-1400 Ma). The ϵ_{Nd} characteristics are compared with those from Lowland tonalites and granitoids of similar age, and the similarity suggests that they are essentially the same, strongly suggesting contiguity across the entire Adirondacks at that time (~1300 Ma).

A disrupted layer of amphibolitic material runs down the outcrop to road level at the east end of the outcrop. This, and other mafic sheets in the outcrop, are interpreted as dikes an sheets coeval with the tonalite. In the eastern Adirondacks it has been possible to document mutually crosscutting relationships between these rock types. Also documented there are xenoliths of kinzigitic rock in the tonalites. Within the southern Adirondacks xenoliths of quartzite similar to the Irving Pond Fm. have been found in the tonalite.

11.8	1.8	Pine Lake, Junction NY Rt. 29A and NY Rt. 10. Proceed north on NY Rt. 10.
17.5	5.7	STOP 5. Rooster Hill megacrystic gneiss at the north end of Stoner Lake (1156±8 Ma, table 1, sample AM-86-17).

STOP 5.

This distinctive unit belongs to the AMCG suite and is widespread in the southern Adirondacks. Here the unit consists of a monotonous series of unlayered to poorly layered gneisses characterized by large (1-4") megacrysts of perthite and microcline perthite. For the most part these megacrysts have been flattened in the plane of foliation, however, a few megacrysts are situated at high angles to the foliation and show tails. The groundmass consists of quartz, oligoclase, biotite, hornblende, garnet, and occasional orthopyroxene. An igneous rock analogue would be monzonite to quartz-monzonite (see table 3 for chemical composition) and the presence of orthopyroxene makes the rock mangeritic to charnockitic.

The contacts of the Rooster Hill megacrystic gneiss are everywhere conformable, but the presence of xenoliths of kinzigite indicate its intrusive nature. Rocks such as the Rooster Hill are interpreted as derived from melting of ca. 1300 Ma tonalitic and lower crustal granitoid rocks with heat derived from large AMCG gabbroic intrusions that would ultimately differentiate to anorthosite. This suggestion is consistent with the ϵ_{Nd} trends of AMCG and tonalitic rocks in figure 7a and with the REE distributions shown in figure 19, where it appears that melting of tonalite so as to leave a plagioclase-rich residue can give the AMCG REE-trends.

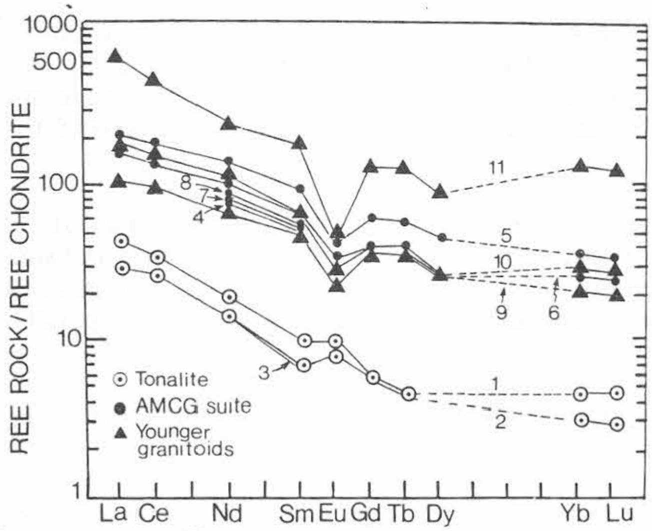


Fig. 19. Chondrite normalized REE concentrations for the Adirondack highlands. Numbers refer to samples in table 1 of Daly and McLelland (1991).

20.0	2.5	Low roadcut in kinzigites.
21.4	1.4	Avery's Hotel on west side of NY Rt. 10
22.5	1.1	Long roadcuts of pink quartzofeldspathic gneisses and metasediments of intruded metagabbro and anorthosite metagabbro. The igneous rocks are believed to belong to the AMCG suite.
23.6	1.1	Roadcut of anorthositic metagabbro and metanorite of AMCG suite.
23.9	.3	Roadcut on west side of highway shows excellent examples of anorthositic gabbros intrusive into layered pink and light green quartzofeldspathic gneisses.
24.0	.1	Pink granitic gneiss of AMCG suite intruded by anorthositic AMCG gabbros and gabbroic anorthosites. Large boudin of calcsilicate in granite.
24.3	.3	Roadcuts of quartzites and other metasediments of the Sacandaga Fm. Mezger (1990) obtained a U-Pb garnet age of ca. 1154 from these rocks.
31.0	5.7	Red-stained AMCG quartzofeldspathic gneisses that have been faulted along NNE fractures.
31.5	.5	Junction of NY Rt. 10 and NY. Rt. 8. End Rt. 10. Turn east on NY Rt. 8.
32.0	.5	STOP 6. Core rocks of the Piseco anticline (1150±5 Ma, table 1, sample AM-86-9).

STOP 6.

This stop lies along the hinge line of the F_2 Piseco anticline near its domical culmination at Piseco Lake. The rocks here are typical ofthe granitic facies of quartzofeldspathic gneisses such as occur in the Piseco anticline and in other large anticlinal structures, for example Snowy Mt. dome, Oregon dome.

The pink "granitic" gneisses of the Piseco anticline do not exhibit marked lithologic variation. Locally grain size is variable and in places megacrysts seem to have been largely grain size reduced and only a few small remnants of cores are seen. The open folds at this locality are minor folds of the F_2 event. Their axes trend N70W and plunge 10-15° SE parallel to the axis of the Piseco anticline.

The most striking aspect of the gneisses in the Piseco anticline is their well-developed lineation. This is expressed by rod, or pencil-like, structures which are clearly the result of ductile extension of quartz and feldspar grains in a granitic protolith. The high temperature, grain size reduction that has occurred results in a mylonite. Where recognizable, early F_1 isoclinal fold axes parallel the lineation.

These rocks are similar in age and chemistry to other AMCG granites and are considered to be part of that suite.

Smooth outcrops of Piseco Core rocks showing exceptionally strong mylonitic ribbon lineations.

43.5	11.5	Junction NY Rt. 8 and NY Rt. 30 in Speculator. Head southeast on NY Rt. 8-30.
47	3.5	STOP 7. Northern intersection of old Rt. NY 30 and new Rt. NY 30, 3.3 miles east of Speculator, New York.

STOP 7.

Typical Adirondack marble is exposed in roadcuts on both sides of the highway. These exposures show examples of the extreme ductility of the carbonate-rich units. The south wall of the roadcut is particularly striking, for here relatively brittle layers of garnetiferous amphibolite have been intensely boudinaged and broken. The marbles, on the other hand, have yielded plastically and flowed extensively during the deformation. As a result, the marble-amphibolite and marble-charnockite relationships are similar to those that would be expected between magma and country rock. Numerous rotated, angular blocks of amphibolite and charnockite are scattered throughout the marble in the fashion of xenoliths in igneous intrusions. At the eastern end of the outcrop tight isoclinal folds of amphibolite and metapelitic gneisses have been broken apart and rotated. The isolated fold noses that remain "floating" in the marble have been aptly termed "tectonic fish". The early, isoclinal folds rotate on earlier foliation. The garnetiferous amphibolites have typical igneous compositions and are interpreted as flows or sills.

Near the west end of the outcrop a boudin of charnockite is well exposed. McLelland and others (1987) have presented evidence that boudin represents a local example of charnockitization by carbonic metamorphism. However, granites of similar composition outside the marble do not develop orthopyroxene, demonstrating the local nature of the process and the limited permeation of the fluid phase.

Exposed at several places in the roadcut are crosscutting veins of tourmaline and quartz displaying a symplectic type of intergrowth. Other veins include hornblende- and sphene-bearing pegmatites.

Almost certainly these marbles are of inorganic origin. No calcium carbonate secreting organisms appear to have existed during the time in which these carbonates were deposited (>1200 Ma ago). Presumably the graphite represents remains of stromatolite-like binding algae that operated in shallow water, intertidal zones. This is consistent with the presence of evaporitic minerals, such as gypsum, in Lowland marbles.

At the eastern end of the outcrop coarse diopside and tremolite are developed in almost monomineralic layers. Valley et al. (1983) showed that the breakdown of almost Mg-pure tremolite to enstatie, diopside, and quartz in these rocks requires low water activity at the regional P,T conditions. Similarly, the local presence of wollastine requires lowering of CO_2 activity, presumably by H_2O. These contrasts demonstrate the highly variable composition of the fluid phase and are consistent with a channelized fluid phase within a largely fluid-absent region.

47.5	.5	Extensive roadcuts in lower part of marble. Quartzites, kinzigites, and leucogneisses dominate. Minor marble and calcsilicate rock is present.
47.9	2.5	Large roadcuts in well-layered, pink quartzofeldspathic gneisses with subordinate amphibolite and calcsilicate rock. The layering here is believed to be tectonic in origin, and the granitic layers represent an intensely deformed granite. The calcsilicate layers may be deformed xenoliths.

49.0 1.1 STOP 8. One half mile south of southern intersection of old Rt. 30 and with new Rt. 30. Anorthositic rocks on the SW margin of the Oregon Dome.

STOP 8.

On the west side of the highway a small roadcut exposes typical Adirondack anorthosite and related phases.

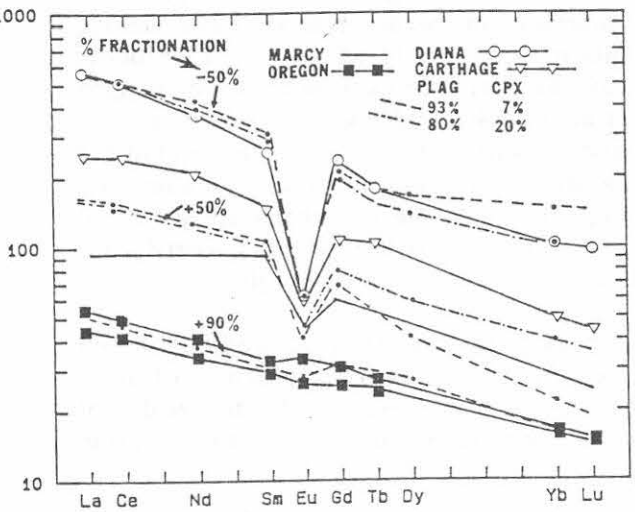


Fig. 20. Chondrite normalized REE concentrations for several Adirondack ferrogabbroic occurrences. Percentage fractionation of plagioclase and clinopyroxene are shown for a starting composition given by Carthage ferrogabbro (trianges). The Diana occurrence corresponds to sheets of breccia-bearing mafic material referred to by Buddington (1939) as shonkinite. The breccia consists of K-feldspar fragments from the host pyroxene syenite of the Diana complex.

The glacially smoothed upper surface of the roadcut reveals the presence of three major igneous phases: 1) a dark, pyroxene-rich dike that crosscuts the anorthosite, contains anorthosite xenoliths, and contains a large irregular, disrupted mass of sulfidic calcsilicate; 2) a coarse grained, Marcy-type anorthosite facies with andesine crystals 6-8" across; and 3) a fine grained anorthositic phase. Some of the coarse grained facies has been crushed and these portions bear some resemblance to the finer grained phase (note, for example, those places where fractures cross large andesine grains and produce finer grained material). However, close inspection of the finer grained material reveals the presence of ophitic texture with pyroxenes of approximately the same size as the plagioclase, and this texture and association are much better explained as igneous in origin. Therefore, the texture of the fine grained phase is interpreted as igneous in origin and may be due to chilling near the contact of the Oregon Dome massif. By contrast, large (3-4 cm) rafts of coarse grained, ophitic gabbroic anorthosite seem to be "flat" in the fine grained phase. Analyses of typical anorthositic rocks are shown in Table 5.

The pyroxene-rich ferrogabbro dike shows "soft" contacts with the anorthosite and is interpreted as essentially coeval. Zircons from it give a minimum age of 1087 Ma and, by comparison with other Adirondack anorthosites, its emplacement age is set at ca. 1135 Ma. The composition of the ferrogabbro is shown in table 5 where it is seen to be rich in TiO_2 and P_2O_5. Similar rocks occur together with other Adirondack anorthosites and are interpreted as late, Fe-enriched differentiates of a Fenner-type fractionation trend (see fig. 14). It is suggested that further differentiation within these rocks can result in liquid immiscibility and the production of magnetite-ilmenite liquids.

Table 5.

	Jotunite	Anorthositic Gabbro Green Mt.	Anorthosite Green Mt.	Anorthosite Owls Head	Mangerite Tupper Lake	Keene Gn. Hulls Falls	Marcy-type Anorthosite[a]	Whiteface Anorthosite[a]
No. of Samples	1	1	1	1	3	2	4	7
SiO_2	47.16	55.88	56.89	53.65	62.12	51.63	54.54	53.54
TiO_2	2.20	1.6	0.47	0.52	0.87	3.1	0.67	0.72
Al_2O_3	17.23	18.23	23.82	24.96	16.48	14.23	25.61	22.50
Fe_2O_3	2.75	2.4	1.21	0.41	1.49	2.1	1.00	1.26
FeO	9.24	6.57	1.3	0.70	3.96	13.5	1.26	4.14
MnO	0.15	0.09	0.02	0.02	0.09	0.16	0.02	0.07
MgO	2.71	2.08	0.65	1.45	1.06	2.63	1.03	2.21
CaO	9.04	4.87	8.19	12.21	3.27	6.5	9.92	10.12
Na_2O	6.61	4.26	5.38	3.92	4.81	2.67	4.53	3.70
K_2O	2.27	2.76	1.13	1.20	5.13	2.41	1.01	1.19
P_2O_5	0.59	0.48	0.09	0.09	0.30	0.57	0.09	0.13
H2O	0	0.09	0.42	0.04	0.32	0.07	0.55	0.12
Total	99.70	99.94	99.57	99.17	99.90	99.57	100.17	100.00

The upper, weathered surface of the outcrop affords the best vantage point for studying the textures and mineralogy of the anorthositic rocks. In several places there can be seen excellent examples of garnet coronas of the type that are common throughout Adirondack anorthosites. These coronas are characterized by garnet rims developed around iron-titanium oxides and pyroxenes. Recently McLelland and Whitney (1977) have succeeded in describing the development of these coronas according to the following generalized reaction:

Orthopyroxene + Plagioclase + Fe-bearing oxide + Quartz = Garnet + Clinopyroxene.

This reaction is similar to one proposed by de Waard (1965) but includes Fe-oxide and quartz as necessary reactant phases. The products are typomorphic of the garnet-clinopyroxene subfacies of the granulite facies (de Waard 1965). The application of various geothermometers to the phases present suggests that the P,T conditions of metamorphism were approximately 8 kb and 700±50°C respectively.

51.0	2.0	Minor marble, amphibolite, and calcsilicate rock. Predominantly very light colored sillimanite-garnet-quartz-feldspar leucogneisses interpreted as minimum-melt granitic due to anatexis of kinzigite near Oregon dome anorthosite. Enclaves of spinel- and sillimanite-bearing metapelite are present.
52.0	1.0	Junction to NY Rt. 8 and NY Rt. 30. Continue south on NY Rt. 30. To the west of the intersection are roadcuts in garnetiferous metasediment. A large NNE normal fault passes through here and fault breccias may be found in the roadcut and the woods beyond.
52.5	.5	Entering granitic-charnockitic gneiss on northern limb of the Glens Falls syncline. Note that dips of foliation are to the south.
54.8	2.3	Entering town of Wells which is situated on a downdropped block of lower Paleozoic sediments. The minimum displacement along the NNE border faults has been

		determined to be at least 1000 meters.
58.3	3.5	Silver bells ski area to the east. The slopes of the ski hill are underlain by coarse anorthositic gabbro that continues to the west and forms the large sheet just south of Speculator.
60.3	2.0	Entrance to Sacandaga public campsite. On the north side of NY Rt. 30 are quartzo-feldspathic gneisses and calcsilicates. An F_1 recumbent fold trends sub-parallel to the outcrop and along its hinge line dips become vertical.
60.8	.5	Gabbro and anorthositic gabbro.
62.0	1.2	STOP 9. Pumpkin Hollow.

STOP 9.

Large roadcuts on the east side of Rt. 30 expose excellent samples of the Sacandaga Fm. At the northern end of the outcrop typical two pyroxene-plagioclase granulites can be seen. The central part of the outcrop contains good light-colored garnet-microcline-quartz gneisses (leucogneisses). Although the weathered surfaces of these rocks are often dark due to staining, fresh samples display the typical white vs. grey color of the Sacandaga Fm. The characteristic and excellent layering of the Sacandaga Fm. is clearly developed. Note the strong flattening parallel to layering. Towards the southern end of the outcrop calc-silicates and marbles make their entrance into the section. At one fresh surface a thin layer of diopsidic marble is exposed.

At the far southern end of the roadcut there exists an exposure of the contact between the quartzo-feldspathic gneisses of the Piseco anticline and the overlying Sacandaga Fm. The hills to the south are composed of homogenous quartzo-feldspathic gneisses coring the Piseco anticline (note how ruggedly this massive unit weathers). The Sacandaga Fm. here has a northerly dip off the northern flank of the Piseco anticline and begins its descent into the southern limb of the Glens Falls syncline.

The pronounced flaggy layering in the Sacandaga Fm. is not of primary sedimentary or volcanic origin. Instead it is tectonic layering within a "straight" gneiss. Hand specimen and microscopic inspection of the light layers, particularly, reveals the existence of extreme grain size reduction and ductile flow. Long quartz rods consist of rectangular compartments of recovered quart and annealed feldspar grains occur throughout. The rock is clearly a mylonite with its mylonitic fabric parallel to compositional layering.

The chemistry of the light colored layers in the Sacandaga Fm. indicates that they are minimum melt granites. As one proceeds away from the core of the Piseco anticline, these granitic layers can be traced into less deformed sheets and veins of coarse granite and pegmatite. In the most illustrative cases the granitic material forms anastamozing sheets that get grain size reduced and drawn into parallelism as high strain zones are approached. The Sacandaga Fm. is interpreted as an end result of this process and represents a mylonitized migmatite envelope developed in metapelites where they were intruded by AMCG granites at ca. 1150 Ma and then intensely strained during the Ottawan Orogeny at ca. 1050 Ma. This interpretation is consistent with field relationships, the presence of spinel and sillimanite restites in the leucosomes, and with the fact that similar metapelitic rocks are crosscut by ca. 1300 Ma tonalites. The latter observation makes the Sacandaga Fm. protoliths older than the ca. 1150 Ma granitic rocks in the Piseco anticline and makes an intrusive relationship obligatory despite the conformable contact at the south end of the roadcut.

62.5-67.0	.5-4.5	All exposures are within the basal quartzo-feldspathic gneisses at the core of the Piseco anticline.
67.0	4.5	Re-enter the Sacandaga Fm. Dips are now southerly.
68.0	1.0	In long roadcuts of southerly dipping pink, quartzo-feldspathic gneisses with tectonic layering. The coarse grain size of the gneissic precursors can be seen in many layers.
70,4	2.4	Cross bridge over Sacandaga River.
74.4	4.0	Bridge crossing east corner of Sacandaga Reservoir into Northville, N.Y.

END LOG

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PATTERNS OF PHYLETIC EVOLUTION IN THE TRENTON GROUP

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INTRODUCTION

The Trenton Group has been the subject of 60 years of continuous biostratigraphic study since Marshall Kay began his work during the early 1930's. The unit displays a remarkably rich and diverse fossil assemblage which, along with its facies patterns, is now quite well known. The Trenton Group is thus very well suited for research directed at understanding patterns of evolution. This ongoing research program is currently focused on recognizing patterns of intraspecific clinal variation and relating these to facies change in order to develop comprehensive records of evolutionary events. Two case studies are now complete. Evolution within the crinoid genus *Ectenocrinus* has been described (Titus, 1989). A similar paper on the brachiopod genus Sowerbyella is pending (Titus, in preparation). This field guide will serve as an appendix to both of those papers.

PATTERNS OF EVOLUTION

The two broadest categories of evolutionary patterns are cladogenesis and anagenesis. Cladogenesis occurs when a population is divided and, from the two isolates, separate species are derived. This is also called allopatric speciation. Cladogenesis is multiplicative as two or more taxa are descended from a single ancestral form. Anagenesis, or phyletic evolution, is not. It consists of evolutionary change within an undivided lineage. It is driven by natural selection and thus conforms to Darwin's original view of evolution. Virtually all recent workers agree that both patterns do occur and much of the current debate centers over their relative frequencies and macroevolutionary significance.

The Trenton Group is well suited for the study of anagenesis. The unit is well exposed across an outcrop belt extending from Canajoharie to Ontario (Figs. 1, 2). Deposition of the unit spanned a period of at least 8 million years without any significant breaks in the record. There is a rich and diverse fossil fauna which is very well preserved. These assemblages are found within a diverse facies mosaic (Fig. 3). The author has spent 20 years studying the Trenton and has documented its facies patterns and the biostratigraphy of over 200 of its species (Titus, 1986, 1988). This provides a wealth

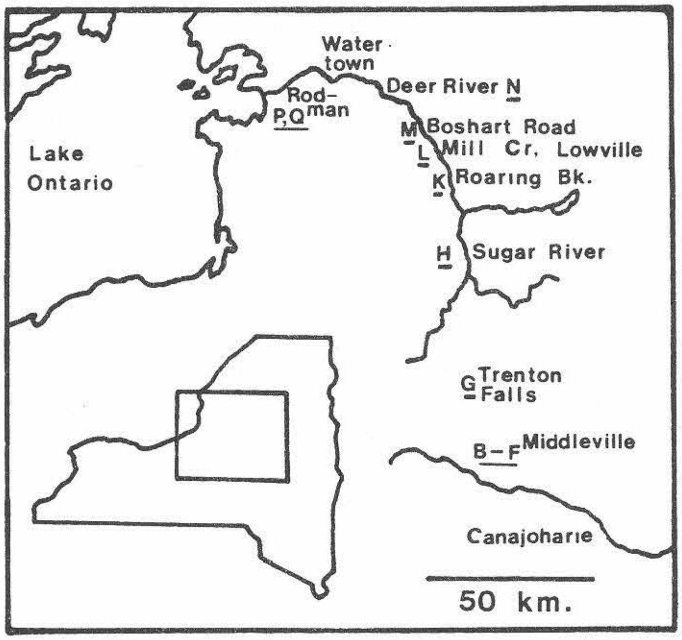


Figure 1. Map of the major locations of the Trenton Group.

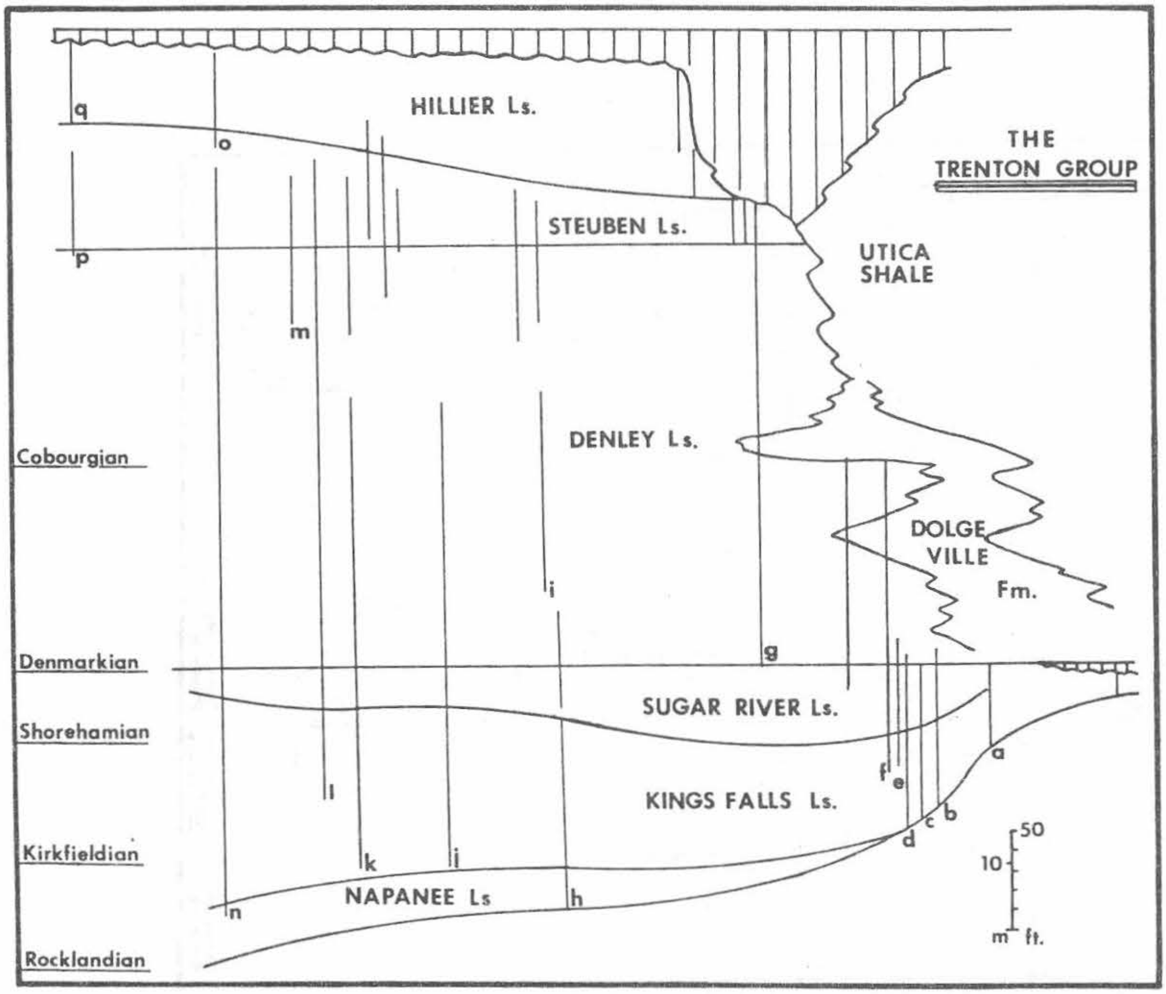


Figure 2. Stratigraphy of the Trenton Group.
Letters refer to Fig. 1.

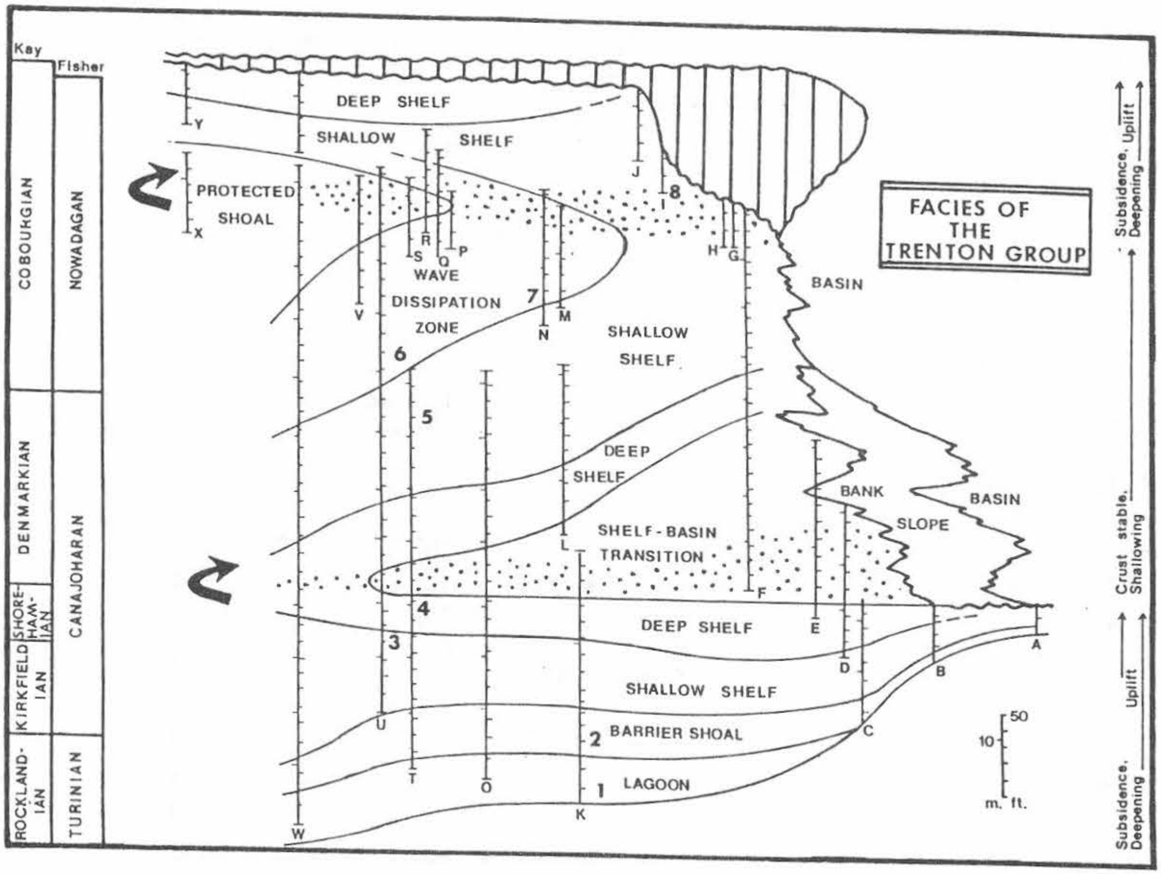


Figure 3. Facies of the Trenton Group. Arrows and stipples define the facies bottlenecks. Stippled areas display very few specimens of *Sowerbyella*. Numbers are trip stops. Stop six is approximate.

of background information. Thus any lineage can be knowledgeably traced through the entire Trenton Group and any phyletic evolutionary change can be observed.

The Trenton Group is not well suited for the documentation of cladogenesis. The carbonate platform is not likely to have any significant reproduction barriers and thus allopatric separation and cladogenesis is unlikely. It has not yet been observed. No examples of punctuated equilibrium have yet been demonstrated, although several possibilities are being investigated.

PHYLETIC EVOLUTION IN THE TRENTON GROUP

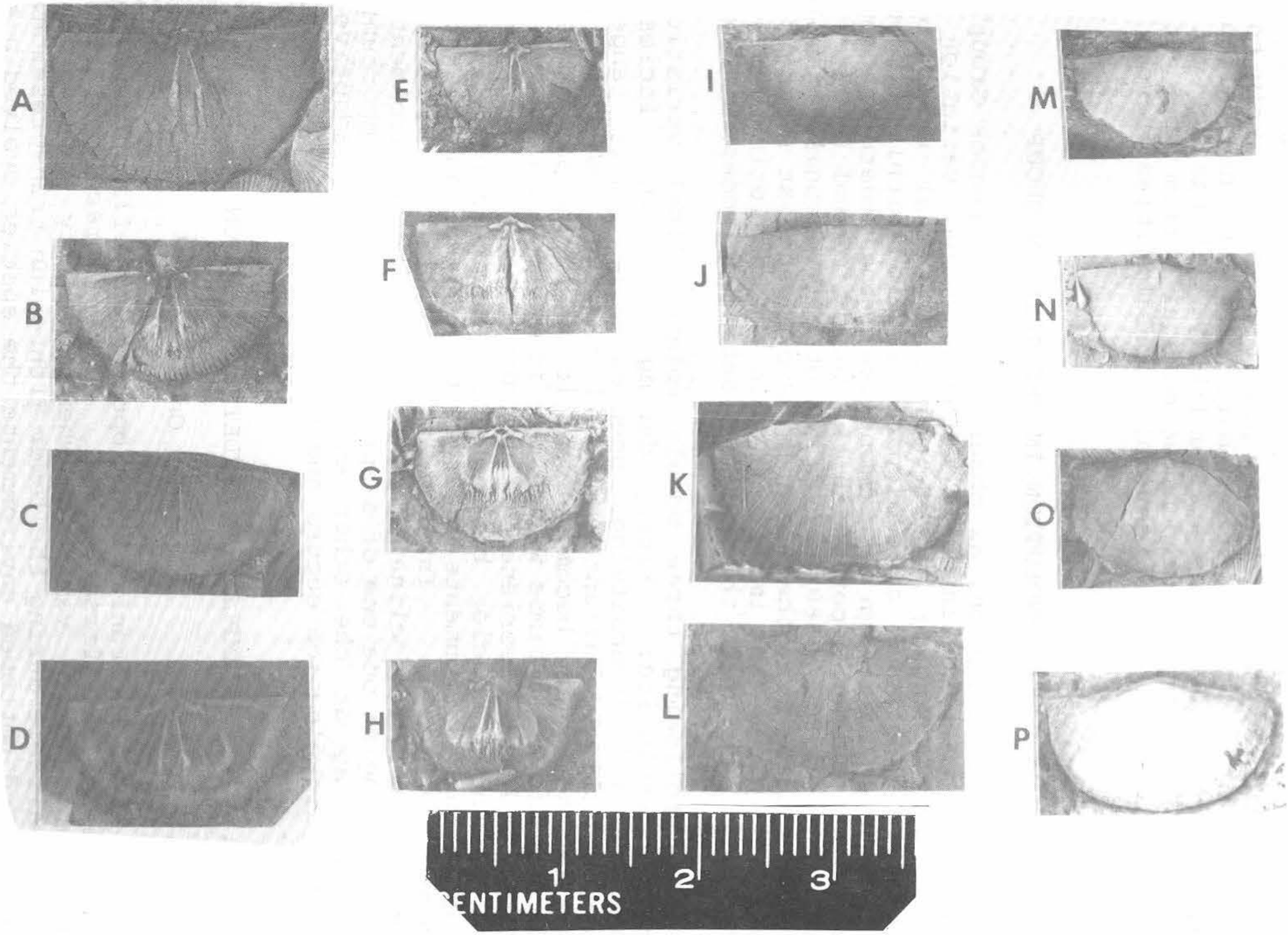
Phyletic evolution, as observed in the Trenton Group, can be described as continuously changing clinal variation in a continuously changing facies mosaic. Clinal variation is systematic geographic or ecologic variation within a species. Although such variation is commonplace among modern species, it has not often been reported in the fossil record (however, see Cisne et al. 1980a, 1980b & 1982 for other Trentonian examples). Recent studies in the Trenton Group suggest that clinal variation plays an important role in phyletic evolution: directional evolution through substantial episodes of time.

The following three models relate clinal variation to phyletic evolution: First, during times when facies are expanding and diversifying, species widen their ranges by evolving clinal variants adapted to the newly available environments. They become eurytopic by polymorphism. Second and conversely, at times when there is restriction of available facies, species experience cline sorting. Polytypic species pass through a facies bottleneck. Morphologies linked to disappearing environments are selected against and the species become stenotopic. In the third variation species can experience a form of clinal orthoselection when environments are deteriorating at one end of a species range while opportunities are expanding at the other end. Continuous, adaptive and directional selection occurs and the species evolves.

EXAMPLES OF PHYLETIC EVOLUTION

Model One

The lower Trentonian brachiopod *Sowerbyella curdsvillensis* illustrates the first model. This brachiopod evolved considerable clinal variation when a variety of environments became available during the lower Trentonian transgression. In quieter, Mud bottomed environments the species evolved a small morphology with a simple brachial valve interior (Fig. 4a). The pedicle exterior has a medial fold and alate corners (Fig. 4i).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CENTIMETERS

This form prevails in the nearshore lagoon and offshore deeper shelf facies (Fig. 3). In the adjacent wave swept, barrier bar facies *Sowerbyella* is much larger and its brachial interiors are quite ornate (Fig. 4d). They display septa, subperipheral thickenings, muscle scars and abundant and well developed vascular markings. The pedicle exteriors are often flattened or broadly rounded. These morphologies grade into each other in what is called a sloping cline. In the past this intergradation was not recognized and the plain, quiet water form was called *S. punctostriatus.*

During deposition of the middle Trenton Group a diversity of environments appeared in shallowing facies pattern (Fig. 3). Once again a model one pattern of clinal variation evolved. Plain brachial interiors occur in the quieter, muddier bottom environments while ornate brachial interiors are found in more agitated settings. This is a striking parallel to the lower Trentonian cline. There are differences, however. A new cline is seen in the morphology of the pedicle exteriors. In the middle Trentonian, deep and shallow shelf facies pedicles tend to have medial folds as was the case in similar lower Trentonian facies. In the barrier facies, however, pedicle exteriors tend to be more rounded and inflated. In the protected shoal facies there tends to be a broadly rounded medial fold. Thus, the overall structure of the middle Trentonian *Sowerbyella* cline is different and this cline should be recognized as a new species.

Model Two

Model Two occurs when there is cline sorting which accompanies times of facies restriction. Such an event would affect a species when facies change reduces its suitable habitats to a minimum. Perhaps a transgression would reduce suitable shallow water habitat, greatly restricting a species range. If polytypic forms pass through such a facies bottleneck, they are subject to intense natural selection. Inadaptive traits are selected against; these disappear and the species emerges from the facies bottleneck altered.

This is illustrated in the *Sowerbyella* populations of the middle Trentonian. Twice, there were times when the range of *Sowerbyella* was greatly restricted (stippled zones on Fig. 3). First, at the close of the lower Trentonian transgression, shallow water habitats had disappeared. The large forms with ornate brachial interiors were selected against and they are not seen above the facies bottleneck (Fig. 3). The middle Trentonian *Sowerbyella*, as noted above, closely parallels the lower Trentonian *S. curdsvillensis*. Both have plain, quiet

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Figure 4. a-d, *Sowerbyella curdsvillensis*, brachial interiors; e-h, *Sowerbyella*. n. sp., brachial interiors; i-l, *S. curdsvillensis*, pedicle exteriors; m-o, *S.* n. sp., pedicle exteriors; p, *S. subovalis*, pedicle exteriors.

water forms grading into ornate types found in agitated facies. However the middle Trentonian *Sowerbyella* is not as big as in the lower Trentonian form and it only very rarely displays well developed vascular markings (Figs. e-h). Large size and vascular markings are traits associated with shallow waters and selected against, within the deep water bottleneck, when the shallow facies disappeared.

A similar facies bottleneck is found at the top of the middle Trentonian Denley Limestone. During deposition of the Denley a well defined, east-to-west and deep-to-shallow *Sowerbyella* cline had evolved (see above). *Sowerbyella*, however, is virtually absent in the lower Steuben Limestone (stippled area, Fig. 3) except in the far west at Rodman. Lower Steuben depositional environments were evidently too agitated for *Sowerbyella* except in the quieter western protected shoal facies (Fig. 3). Only one clinal variant is found there. This form is recognized by the broadly rounded medial fold on the pedicle exterior (Fig. 4p) and plain brachial interior. This clinal variant made it through the western facies bottleneck (arrow on Fig. 3) and became a founding population for *Sowerbyella subovalis*, the most common species of *Sowerbyella* seen in the upper Trenton Group. The broadly rounded pedicle exterior is what most characterizes *S. subovalis*. The trait was, evidently, inherited from the bottleneck population.

Model Three

The columnals of the crinoid genus *Ectenocrinus* record an example of Model Three clinal variation. *Ectenocrinus* is first found in the deeper shelf facies of the Sugar River Limestone (Fig. 3). There it is commonly recognized by its nearly triangular columnals with triangular lumina (Fig. 5a). It passed, without effect, through the same facies bottleneck that generated the middle Trentonian species of Sowerbyella (arrow on Fig. 3). During the middle Trentonian, in a Model One case, it evolved a variety of clinal variants, which ranged in distribution from deep shelf to shallow water extremes (Fig. 5). The newly evolved shallow water clinal variants forms have round columnals with pentagonal lumina (Fig. 5z, aa). The middle Trentonian records a long episode of shallowing facies. This included the carbonate bank margin which was shallowing and steepening (Titus, 1986, Fig. 7). Thus while new shallow water facies were opening up in the west, the old, deeper shelf facies were being restricted in the east. The result was a form of clinal orthoselection. As deep water environments shrank, the triangular forms disappeared. At the same time shallowing facies favored the round columnal forms and they eventually were the survivors. The transition is between the species *E. triangulus* and *E. simplex*. The transition is gradual and both facies and clines are central to this event.

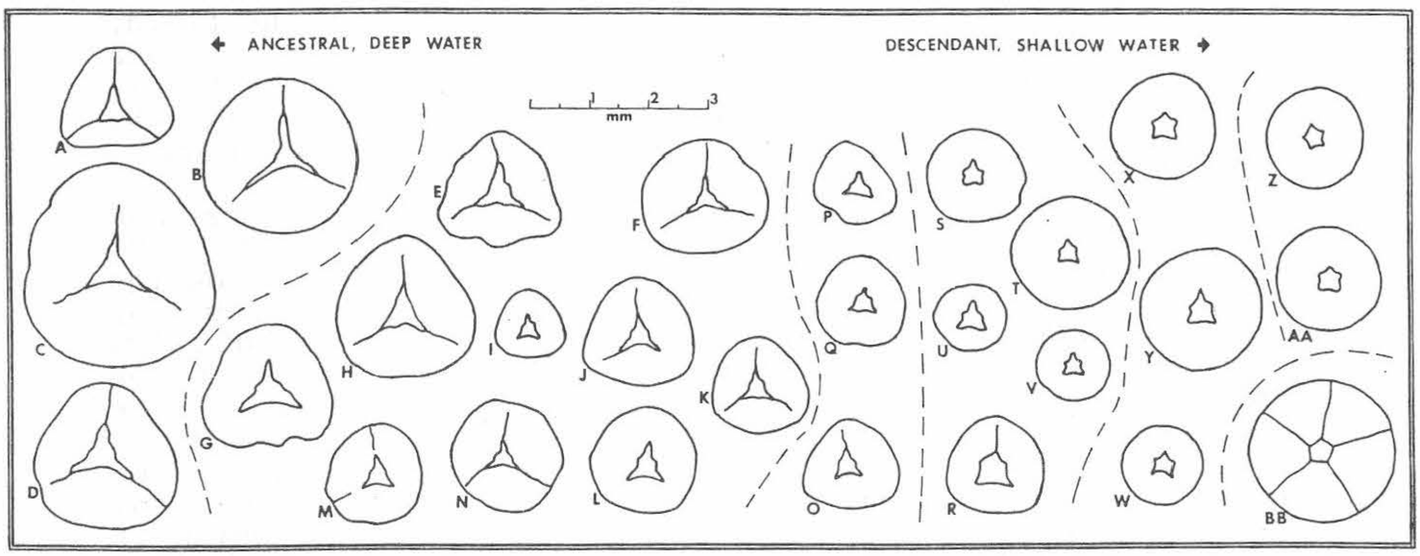

Figure 5. Columnals of *Ectenocrinus*. A-N, *Ectenocrinus triangulus*; O-W, intermediate forms; Z,AA, *Ectenocrinus simplex*; BB, a form of *Cupulocrinus* which is easily confused with *E. simplex*. *Cupulocrinus* is pentameric, not trimeric.

CONCLUSIONS

In the half century since clinal variation was first described by Huxley, paleontologists contributed little to our understanding of this concept. There needs to be a greater awareness of clinal variation in the definition of fossil species and as a factor in phyletic evolution. To date, biologists have done most of the work but they have viewed clinal variation only in terms of geography and ecology. Paleontologists can explore the temporal dimension. This is being done in the Trenton Group where we can observe clines through time as well as through space. We see in the Trenton Group that, by the standards of geologic time, clinal variation is immediately adaptive to changing environments. It is through clinal variation that species can track facies change through time. Finally it is the linkage of clinal variation and facies change through time which appears to define phyletic evolution.

FIELD DISCUSSION TOPICS

Our trip gives us an opportunity to examine, in the field, evidence pertinent to some of the most difficult issues confronting paleontologists as we try to understand the fossil record of evolution. At the end of this trip we might well take time to discuss some of them. Two issues are outlined below.

This field trip illustrates some of the basic problems that paleontologists face in recognizing fossil species. We have always relied too much on the typological approach to species recognition. The two lower Trentonian species of *Sowerbyella*, which have been traditionally recognized, are *S. punctostriatus* and *S. curdsvillensis*. As we have seen, they turn out to be intergrading clinal variants of a single species. The typological approach has failed us, but when we employ a polytypic approach to species recognition other difficulties appear. The polytypic middle Trentonian form, *Sowerbyella*. n. sp., is an example. This form has a clinal variant which is identical to the deep water forms of the lower Trentonian species. It also has a clinal variant which is identical to the upper Trentonian form, *S. subovalis*. What is unique about the middle Trentonian form is the *combination* of clinal variants. Is this a proper criteria for species definition? I think so. How do you react?

A number of workers vigorously deny that phyletic evolution can produce new species. Species, instead, are regarded as sharply bounded spatiotemporal entities. No matter how much evolution may occur within a phyletic lineage, they argue that only one species should be recognized. Adherents of this point of view would thus argue that only one species of *Sowerbyella* can be recognized in the entire Trentonian sequence. I believe that I am observing phyletic evolution producing change at the species level and thus that phyletic change does result in macroevolution. I suspect that species generally are not spatiotemporally bounded entities and therefore that species are emphemeral, evolving continuously through time. You have seen some of the evidence basic to my argument. How do you react to this issue?

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ROAD LOG

The first three stops are designed to display a model one pattern of clinal variation. The brachiopod *Sowerbyella curdsvillensis*, by evolving several different clinal variants, was able to occupy most of the environments seen in the lower Trenton Group. It became a polytypic and eurytopic form.

TOTAL MILES	MILES FROM LAST POINT	ROUTE DESCRIPTION
0.0		Trip meets and organizes at the prominent outcrop of the Napanee Limestone just west of the bridge where Rt. 12 crosses the Sugar River,

		3 miles west of Boonville. The group will walk up the west side of the stream and gather at the outcrop of the Napanee beneath the railroad bridge.

STOP 1 THE NAPANEE LIMESTONE

The strata here are primarily thick bedded, black micrites which have been interpreted as a nearshore lagoon facies. *Sowerbyella* is abundant. This form has traditionally been identified as *S. punctostriatus*. It is herein reinterpreted as the quiet water, mud bottom clinal variant of *S. curdsvillensis*. Examine the brachial interiors and see the preponderence of simple forms. Medial septa can be seen but muscle scars, vascular markings and subperipheral thickenings are uncommon. The pedicle exterior usually displays alate corners and a well developed medial fold giving this valve a peaked appearance.

0.0	0.0	Return downstream and cross the Rt. 12 bridge to the outcrop of the Kings Falls Limestone east of the road.

STOP 2 THE KINGS FALLS LIMESTONE

The lower strata here are primarily thick bedded, coarse grained biosparites which have been interpreted as a barrier shoal facies, lying offshore of the Napanee lagoon facies. Toward the top of the outcrop the facies grades into a shallow shelf facies. While, at this location, these strata overlie those of the lagoon facies, presumably there were contemporaneous lagoonal facies elsewhere. *Sowerbyella* is very abundant at the base of this outcrop. This form is larger than the one in the lagoon facies. Most of the brachial interiors are quite ornate. They have medial septa, muscle scars, vascular markings and subperipheral thickenings. The pedicle exteriors have blunt corners. They don't often show medial folds. Instead they tend to be flattened or broadly rounded. This form has traditionally been identified as S. *curdsvillensis*. It is herein reinterpreted as the agitated facies clinal variant of a more broadly defined *S. curdsvillensis*. By collecting enough brachial interiors at these two outcrops, an assemblage of forms can soon be put together that show the gradation between the lagoon and the shoal clinal variants.

21.1	21.1	Drive north on Rt. 12 to Lowville. Turn left in Lowville to remain on Rt. 12.
22.0	0.9	Drive north on Rt. 12. Cross Mill Creek bridge and park immediately. Climb down to creek and walk to exposure east of and about 50m below

the bridge.

STOP 3 THE SUGAR RIVER LIMESTONE

There is a prominent re-entrant below the bridge which can be traced downstream. This is a thick bentonite which lies 10 m (32 ft.) below the top of the Sugar River Limestone. *Sowerbyella* is rare in the deeper water facies above this bentonite, but common in the shallower facies below. Examine the strata several meters below the bentonite. The common form of *Sowerbyella* here is the mud bottom, quiet water form. Although some of the brachial interiors are ornate, most are plain. Medial septa are seen but other features are uncommon. The form is relatively small. Pedicle exteriors often display a medial fold and alate corners. This location records a return of both quiet water conditions and the quiet water clinal variant of *Sowerbyella*. Upstream into the highest strata of the Sugar River Limestone, and for a considerable distance into the overlying Denley Limestone, *Sowerbyella* is quite scarce. This zone is part of the first *Sowerbyella* bottleneck.

The next location is designed to display an example of model 3 clinal variation. The middle Trentonian crinoid genus *Ectenocrinus* occupied a shallow to deep shelf range of environments. Because of bank margin steepening, the deep end of its range was deteriorating while shallowing facies offered opportunities for *Ectenocrinus* in shallow water environments. The result is a kind of clinal orthoselection.

23.3	1.3	Turn around and return, on Rt. 12 through Lowville. Take the right fork onto Rt. 26.
26.7	3.4	Follow Rt. 26 through Martinsburg. Turn left onto Glendale Ave.
27.5	0.8	Turn left and enter Whitikers Falls Town Park. Park and follow one of the trails down to Roaring Brook and then climb down to the top of Whitikers Falls.

STOP 4 THE SUGAR RIVER AND DENLEY LIMESTONES

At the base of the falls a re-entrant can be seen. This is the same bentonite which was observed at Mill Creek in Lowville. Here it is also 10 m (32 ft.) below the top of the Sugar River Limestone. The bentonite is a marker bed which is found at the same level at all outcrops from Mill Creek, Lowville to Sugar River. It has not been found at Deer River or at any of the Mohawk Valley locations.

Examine the several meters of strata above the falls. *Ectenocrinus* is represented by moderately abundant columnals. Look carefully as these columnals are small. It is very helpful to bring a water container and pour water on columnal rich beds. This brings out the contrast between the columnals and their micritic groundmass. *Ectenocrinus* is easily recognized by its trimeric morphology. Each columnal is composed of three elements fused together along faintly visible sutures. The most common form is a triangle with very rounded corners. At first glance the lumina appear to be triangular, and many are. But you will soon notice that many also have very poorly developed 4th and 5th points, along with three long and attenuated points (Fig. 5i-5m). All specimens at this level belong to the species *E. triangulus*.

Climb to the level above where the trail intersects the stream. Here and for quite a distance upstream *Ectenocrinus* columnals are different. Now they are generally rounder and the lumina are composed of more equal sized points (Fig. 5s-5v). Forms at this level are intermediate between *E. triangulus* and its descendent *E. simplex*. Continue upstream and observe the increasingly abundant columnals of *Ectenocrinus*. The transition from *E. triangulus* to *E. simplex* is not just one of morphology. *E. triangulus* was a relatively stenotopic form which never became especially abundant. The descendent, *E. simplex*, was altogether different. It was eurytopic and one of the dominant forms in the upper Trenton Group, a "weed" crinoid. This evolutionary event records an ecological transition from the K-strategist ancestor to the r-strategist descendent.

The last locations are designed to illustrate model 2 examples of clinal variation. *Sowerbyella* passed through two facies bottlenecks, one at the base of the middle Trentonian and one at the base of the upper Trentonian (Fig. 3, arrows). The best place to study the *Sowerbyella* of the middle Trentonian is at Mill Creek, Lowville, but as that outcrop extends for such a long distance along the stream, it is not practical to include this location in a one-day trip. We will examine middle Trentonian forms at other locations, stops 5 and 6.

		Return to the cars and drive back to the park entrance.
27.7	0.2	Turn left and travel to bridge where Glendale Road crosses Roaring Brook. Climb to the exposure immediately above the bridge.

STOP 5 THE DENLEY LIMESTONE

At this level the transition to *Ectenocrinus simplex* is complete. Virtually all *Ectenocrinus* columnals, upstream from the bridge, are round with pentagonal lumina (Fig. 5 z-5aa). *Sowerbyella* is not common here but the specimens which can be seen are of interest. They are nearly identical to the quiet water lower Trentonian forms. They have plain brachial interiors; the pedicle exteriors have medial ridges and alate corners.

28.9	1.2	Turn around and travel back towards Martinsburg. Turn right on Rt. 26.
30.8	1.9	Turn left onto B. Arthur Rd.
32.7	1.9	Turn left onto W. Martinsburg Rd.
32.8	0.1	Stop at large road outcrop on the right.

STOP 6 THE DENLEY LIMESTONE

The exposure here is of the middle Denley Limestone. The facies are the shallow shelf possibly grading into the barrier (Fig. 3). Several types of *Sowerbyella* can be recognized from the brachial interior structure. Many have plain interiors. While medial septa are present, none of the other internal structures are seen. The other type has an ornate brachial interior. Medial septa, muscle scars and subperipheral thickenings can be seen, some or all, on the same interiors. The plain forms found at this outcrop are typical of the Denley's deeper shelf facies while the ornate forms, although never very common, are typical of the more shallow and agitated facies. Significantly, vascular markings are quite unusual and these ornate forms are never as large as their lower Trentonian equivalents. Pedicle exteriors are more inflated and rounded. This is *Sowerbyella* n. sp.

33.9	1.1	Continue south on West Martinsburg Road. Turn left onto West Road.
51.4	17.7	Head east on West Road. At Whetstone Gulf State Park bear right onto Rt. 26. Continue east on Rt. 26 past its junction with Rt. 12D. Continue on 12D until the bridge over Sugar River at Talcottsville. Park, descend dirt path to outcrop.

STOP 7 THE UPPER DENLEY LIMESTONE

The best location to see *Sowerbyella* next is at the outcrop of the lower Steuben Limestone on Rt. 177 just west of Rodman, but this is too far away for our trip today. The Rodman exposure shows *Sowerbyella* within the second bottleneck (Fig. 3, arrow). Specimens there are small and have a well developed, broadly rounded medial fold. At Talcottsville we can see

Sowerbyella in the barrier facies just below this second bottleneck. The form is abundant up to the base of the falls and then quite a bit less common in the bottlenecking beds above. The specimens here are, for the most part, characterized by well inflated pedicle exteriors which sometimes display the broadly rounded medial folds typical of the second bottleneck. Brachial interiors are generally plain. Notice the magnificent display of large symmetrical ripple marks (pararipples) here.

55.4	4.0	Continue south on Rt. 12D. Enter Boonville; enter Rt. 46 at its junction with Rt. 12D (no turn).
69.7	14.3	Follow Rt. 46 south to Frenchville. Turn left onto Rt. 274.
70.8	1.1	Park just west of the bridge over Big Brook. Descend to outcrop east of the brook.

STOP 8 THE HILLIER LIMESTONE

The Hillier Limestone is exposed all along Wells Creek and Big Brook. The environment of deposition was quiet deep shelf and the deposits are mostly biomicrites. This location is above the second facies bottleneck and a third species of *Sowerbyella* is exposed. The form, *S. subovalis*, is characterized by a greatly inflated pedicle exterior with a broadly rounded medial fold. Brachial interiors are mostly plain. Medial septa are seen, and sometimes poorly developed subperipheral folds can be found, but muscle scars and vascular markings are absent. These traits are apparently inherited from the clinal variants of S. n. sp. which made it through the second bottleneck. The ones on Big Brook are larger than their middle Trentonian ancestors.

END OF FIELD TRIP

75.4	4.2	Continue east on Rt. 274. Turn left onto the road to Remsen.
79.0	3.6	Turn right onto the road to Remsen.
79.5	4.1	Turn right onto Rt. 12.

To get to Oneonta for the rest of the meeting, follow Rt. 12 through Utica. South of Utica, exit onto Rt. 8 south. Follow Rt. 8 until New Berlin. There take County Road 13 to Morris. At Morris, pick up Rt. 23 to Oneonta.

PAC STRATIGRAPHY OF THE HELDERBERG GROUP: CYCLE DEFINITION, ALLOGENIC SURFACES, HIERARCHY, CORRELATION AND RELATIONSHIP TO "VAIL" SEQUENCES

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ABSTRACT

Sections at Thacher Park and Schoharie demonstrate our method of defining 6th order rock cycles (PACs) on the basis of facies patterns (as opposed to key lithologies). Cycle boundaries are placed at surfaces where deeper facies abruptly overlie shallower facies in a disjunct relationship. Several genetically distinct kinds of stratigraphic surfaces are well illustrated in these sections, including PAC boundaries, intracycle surfaces (sea-level fall surfaces) and cryptic unconformities. Criteria for correlation of cycles (PACs) in the Manlius Formation between the two localities include: degree of facies change at PAC boundaries, tracing distinctive litho and bio facies, matching patterns of facies change through several cycles and matching the hierarchic pattern of cycles. In the Manlius Formation alone 14 or more 6th order cycles, as many as six 5th order cycles, parts of two 4th order cycles and an incomplete 3rd order B cycle are represented. The Manlius-Coeymans formational boundary is a cryptic unconformity that represents a 3rd order B "Vail" sequence boundary. The stratigraphic section at Schoharie from this unconformity to the Oriskany unconformity represents a nearly complete 3rd order B sequence and is tentatively divisible into a hierarchy of three 4th order and numerous 5th and 6th order sequences (rock cycles). In terms of systems tract analysis the section equivalent to the Coeymans and Kalkberg Formations is identified as shelf margin wedge, the lowermost New Scotland (equivalent) rocks are the transgressive systems tract and the remaining New Scotland and Becraft Formations are the highstand systems tract.

INTRODUCTION

Our purpose on this trip is to introduce participants to field application of allogenic, hierarchic cyclic stratigraphy. Participants are asked to begin with the assumption that the internal structure of the stratigraphic record is a product of a hierarchic set of cyclic processes that caused sea-level fluctuations. If this assumption is correct then the significant components of the stratigraphic record may include rock cycles arranged in a hierarchy, a variety of allogenic surfaces both within cycles and at their boundaries and a systematically arranged set of unconformities. We will attempt to provide a basis for discriminating the various types of cycles and surfaces and in doing so make the case that field analysis is dictated by assumed models. We would like to emphasize that our focus on these stratigraphic components of the rock record is different from the traditional focus on grains, laminations, beds and formations. New stratigraphic models therefore require new field methods and a different approach to recording data (see Anderson and Goodwin, 1990; Goodwin and Anderson, 1988 and Goodwin et al., 1986).

In this field exercise we will be applying a hierarchic stratigraphic model consisting of 5 orders of allogenic cycles:

3RD ORDER A (Supersequence)	8-10 million years
3RD ORDER B (Sequence)	1-2 million years
4TH ORDER (eccentricity)	400 thousand years
5TH ORDER (eccentricity)	100 thousand years
6TH ORDER (precession)	20 thousand years

Choice of this particular hierarchy is justified by the large number of sequence stratigraphic studies documenting the existence of the two "third order" natural groupings and by the growing number of studies confirming the presence of Milankovitch cyclicity throughout the stratigraphic record.

STOP 1, THACHER PARK

This locality has been selected to illustrate the fundamental, 6th order, meter-scale, allogenic cycle (fig. 1). This cycle is thought to be produced by the 20ky, precessional Milankovitch signal. At this locality these rock cycles range from less than one meter to almost three meters in thickness. Some of them are totally subtidal and others (e.g. PACs 9 and 12) have tidal flat facies at their tops. Cycle boundaries are placed at surfaces where deeper water facies abruptly overlie shallower water facies independent of the occurrence of specific facies. **Cycles are pattern-defined and do not require the presence of tidal flat facies or any other recurrent lithology.**

Five partially complete fifth order cycles are recognized at Thacher Park in the interval represented by the Rondout and Manlius Formations (PAC 3 by itself; PACs 5-6; PACs 7-9; PACs 10-12; PACs 13-14). In that fifth order cycles are thought to be the product of the 100ky eccentricity cycle this section at Thacher Park is incomplete in terms of the number of 6th order cycles within each 5th order cycle. Reasons that could explain this incompleteness include hiatus, vacuity, failure to recognize 6th order cycles and incorrectly designated 5th order boundaries. PACs 5-14 are potentially the record of a 4th order cycle, the product of the 400ky eccentricity cycle.

The final purpose of using this section is to illustrate our approach to correlation at the level of 6th order cycles. For example, distinctive widespread lithologies are introduced at the bases of two PACs in the sequence of PACs at Thacher Park. A subtidal well-sorted calcarenite appears in PAC 7 at all localities (fig. 1) and at all localities PAC 11 is thick bedded and contains well developed stromatoporoids. These two PACs in which relatively large facies changes occur represent the response to larger deepening events in the basin. PAC 7 is the first 6th order cycle in a 5th order sequence of cycles and PAC 11 contains the deepest facies in its 5th order sequence. In summary distinctive lithologies in certain cycles, the magnitude of facies change at cycle boundaries and the sequential pattern of cycles are all keys to correlation.

STOP 2, SCHOHARIE A, I-88

This locality illustrates a record of 6th order cycles in a different, slightly more offshore set of Manlius facies than those seen at Thacher Park. For example, PACs 9 and 12, in which high tidal-flat facies were developed at Thacher Park, are here represented by nearly all subtidal facies. Second the record of 6th order cycles associated with unconformities is more complete. PAC 2 (actually a 5th order sequence), PAC 6b and PAC 15, recognized here at Schoharie, are missing in the more onshore Thacher area. PAC 2 and 6b are missing owing to hiatus (non-deposition) at Thacher; PAC 15 is missing due to erosion (i.e. vacuity).

In addition to PAC boundaries and unconformable surfaces, the section at Schoharie contains good examples of intracycle allogenic surfaces. Whereas PAC boundary surfaces are thought to be produced by the eustatic sea-level rise forced by the precessional (20ky) cycle, intracycle allogenic surfaces are produced by the complementary sea-level fall on the precessional eustasy curve. Thus each 6th order cycle potentially may be composed of a set of sea-level highstand facies overlain abruptly by a set of lowstand facies. **The surface between the two facies sets is thus an intracycle allogenic stratigraphic surface.** Such surfaces are well illustrated in PACs 9, 11 and 12, in which massive ostracod bearing calcarenites sharply overlie thin bedded ribbon calcarenites.

Finally the Manlius-Coeymans boundary is an erosional unconformity. At Syracuse, 100km to the west, an additional two 6th order cycles are preserved below this erosion surface (PACs 16 and 17) while at Kingston, 100km to the southeast, PACs 14 and 15, present at Schoharie, have been removed by erosion (fig. 1). This major erosion surface is the basis for defining this surface as a third order "Vail" sequence boundary. The section from the Brayman-Cobleskill unconformity to this unconformity represents the second 3rd order B sequence in the Helderberg

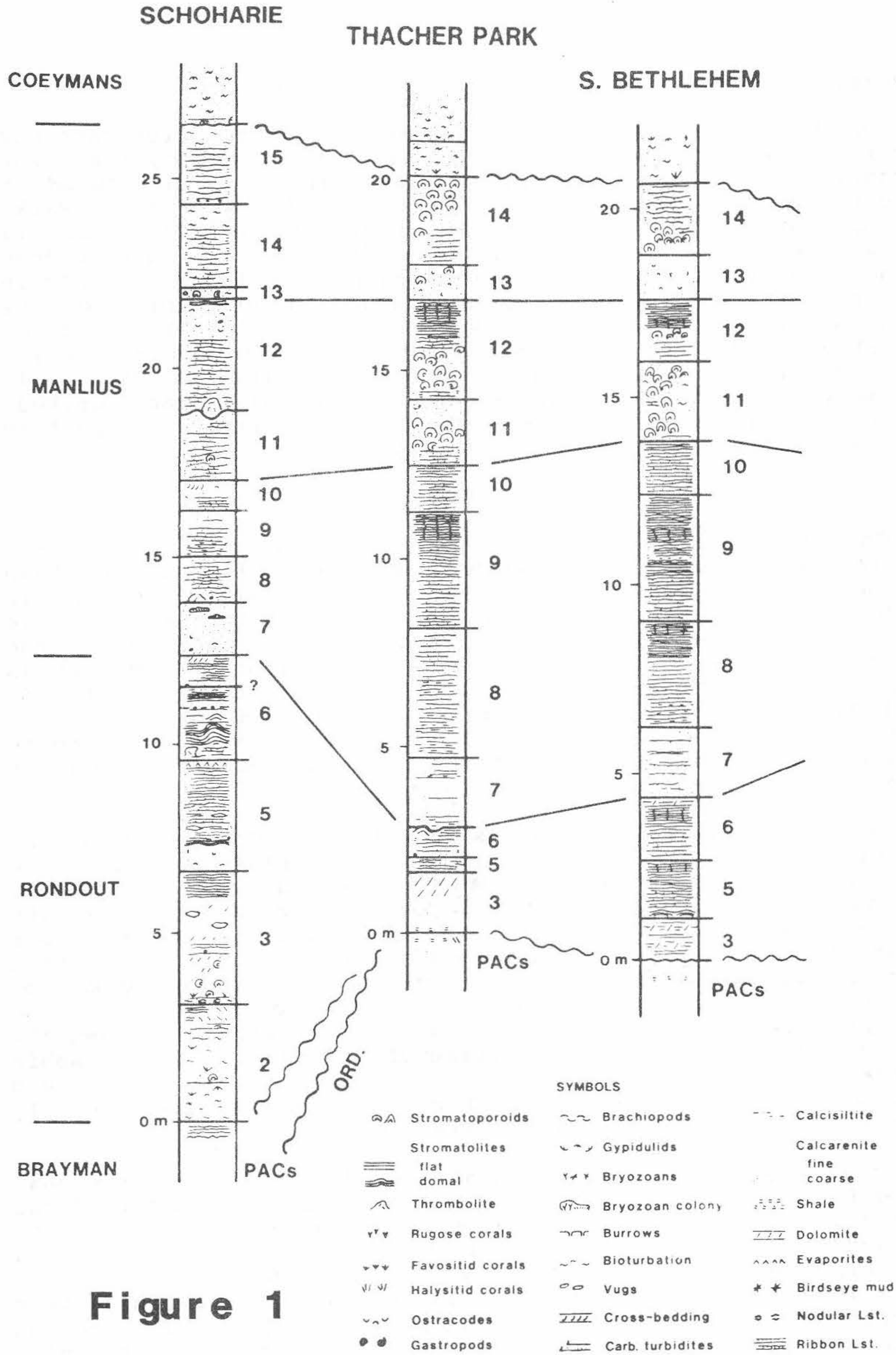
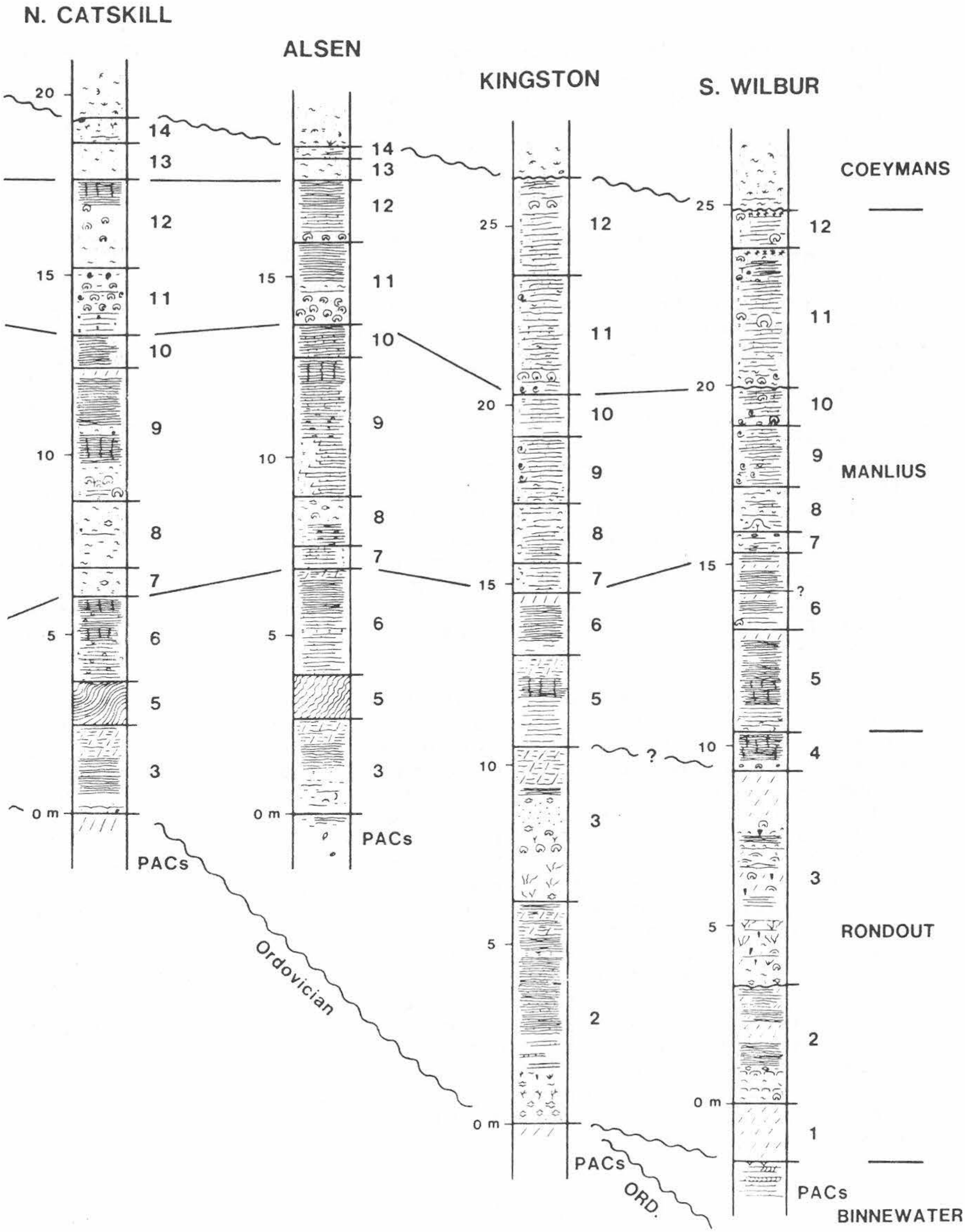

SCHOHARIE
THACHER PARK
S. BETHLEHEM
COEYMANS
MANLIUS
RONDOUT
BRAYMAN
PACs
ORD.
SYMBOLS
Stromatoporoids
Stromatolites
flat
domal
Thrombolite
Rugose corals
Favositid corals
Halysitid corals
Ostracodes
Gastropods
Brachiopods
Gypidulids
Bryozoans
Bryozoan colony
Burrows
Bioturbation
Vugs
Cross-bedding
Carb. turbidites
Calcisiltite
Calcarenite
fine
coarse
Shale
Dolomite
Evaporites
Birdseye mud
Nodular Lst.
Ribbon Lst.

Figure 1


N. CATSKILL
ALSEN
KINGSTON
S. WILBUR
COEYMANS
MANLIUS
RONDOUT
BINNEWATER
PACs
Ordovician
ORD.
0 m
5
10
15
20
25
1
2
3
4
5
6
7
8
9
10
11
12
13
14
?

supersequence. The section from this surface up to the top of the Becraft Formation (seen in the next two stops) is the third 3rd order B sequence. Erosion associated with the Oriskany unconformity (last stop) removes most of a fourth 3rd order B sequence in the Schoharie area.

STOP 3, SCHOHARIE B, I-88

This final stop is in three parts. The first and stratigraphically lowest part is on the north side of I-88 two miles west of STOP 2. The lower 25 feet of this exposure are equivalent to the upper half of the Hannacroix Member of the Kalkberg Formation in the Kingston-Catskill area and the upper five feet are equivalent to the lower part of the Broncks Lake Member (fig. 2). The base of this section is 2-3 feet stratigraphically above the top of rocks exposed in the section seen at STOP 2. The section continues in the large road cut on the south side of I-88. There is a seven foot covered interval between the two exposures. The thickness of these two covered intervals can be exactly determined because the entire stratigraphic interval is continuously exposed three miles to the east in the Schoharie Quarry and all exposed cycles and bounding surfaces can be correlated between the two localities. The Becraft and Oriskany Formations are exposed at the top of the large roadcut; however a more accessible exposure of the upper Becraft and Oriskany rocks occurs in I-88 roadcuts one mile farther west.

Our purpose in looking at the stratigraphic interval between the Manlius-Coeymans unconformity and the top of the Becraft is to lay out a tentative interpretation of the hierarchic cyclic structure of a fairly complete 3rd order B sequence representing 1-2 million years of accumulation. The entire section is in subtidal shallow to deep ramp facies. It is divisible into three 4th order sequences associated with the 400ky eccentricity cycle. These sequences in turn are each divided into four 5th order cycles related to 100ky eccentricity. Examples of 6th order cycles (PACs) in these facies will be pointed out, but analysis at this level is incomplete. Analysis of 4th and 5th order cycles is based on pattern and magnitude of facies changes and depends on correlation of those patterns to other localities. In this case larger scale cycle boundaries have been correlated to the Cherry Valley area to the west and to localities at Broncks Lake, Catskill and Kingston in the Hudson Valley. In short 3rd, 4th and 5th order cycle boundaries in the Schoharie sections have been correlated with prominent surfaces in sections over 100km away.

<u>4th Order Cycle I</u>

This cycle begins at the Manlius-Coeymans unconformity, also a 3rd order B boundary, and ends at the 161 foot mark on the log (fig. 2). The upper boundary is a surface between massive calcarenite and chert bearing calcisiltite (near the base of the large cut on the south side of I-88. A prominent shale occurs on the surface. The same surface is found on top of the shallowest facies in the Broncks Lake Member in the Hudson Valley (e.g. 16 feet below the New Scotland at Kingston). In systems tract analysis this 4th order cycle represents the shelf margin wedge of the third order B sequence. Fourth order cycle I can be divided into four 5th order cycles:

5th Order Cycle 1 ends at 108.5 feet on the log at the shallowest point in Coeymans facies. This surface is the Coeymans-Kalkberg Boundary at Kingston and Catskill.

5th Order Cycle 2 ends at 135 feet on the log at a surface separating coarser favositid rich calcarenites from gypidulid bearing, more argillaceous calcarenites.

5th Order Cycle 3 ends at 140 feet on the log at a surface just below a one foot thick shale interval with lime turbidite beds. This surface is the base of the Broncks Lake Member in the Hudson Valley.

5th Order Cycle 4 ends at the 4th order boundary.

4th Order Cycle II

This cycle contains the deepest facies in the third order sequence and at Schoharie its thickness is highly condensed (32 feet versus 66 feet at Kingston). The lower boundary (described above) is at 161 feet; the upper boundary, at 193 feet on the log, is the surface at the top of a six-foot-thick massive calcisiltite unit. In systems tract analysis of the third order B sequence the lower two 5th order cycles in this 4th order cycle represent the transgressive systems tract and the 5th order boundary at 181 feet would be the maximum flooding surface. Fourth order cycle II is divisible into four 5th order cycles:

5th Order Cycle 1 is 10.5 feet thick and in the lower part is charcterized by several amalgamated lime turbidites with cherts. These amalgamated units are thought to represent 6th order cycles. The upper 6th order unit is represented by 2.5 feet of calcisiltites in dicontinuous lenses. The top surface of this first 5th order cycle is correlated to the Kalkberg-New Scotland formational boundary in the Hudson Valley.

5th Order Cycle 2 ends at 181 feet on the log; a concentration of chert and a thick shale occur at this boundary.

5th Order Cycle 3 ends at 187 feet on the log; beds representing condensed 6th order cycles thicken up through this 5th order cycle.

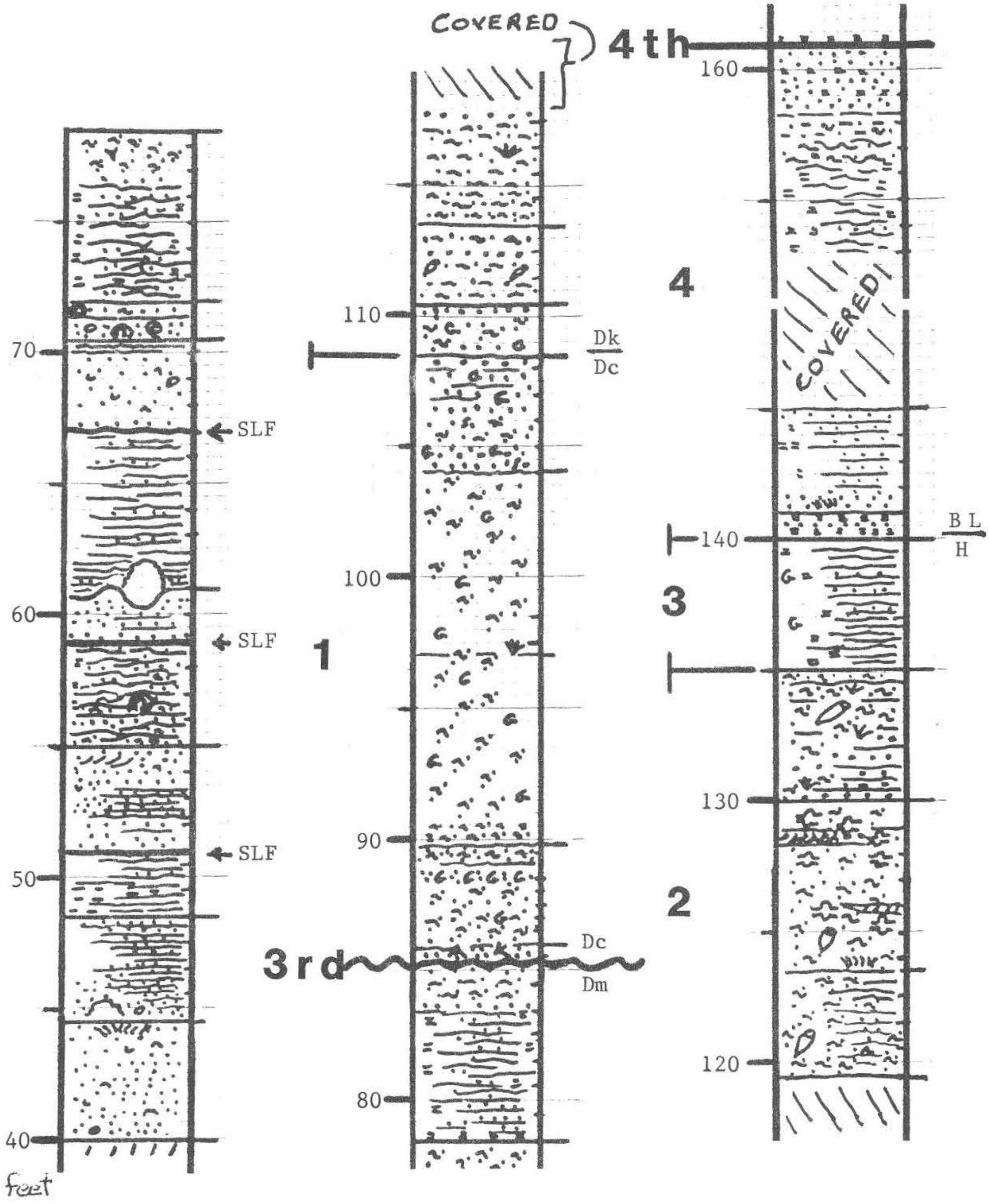
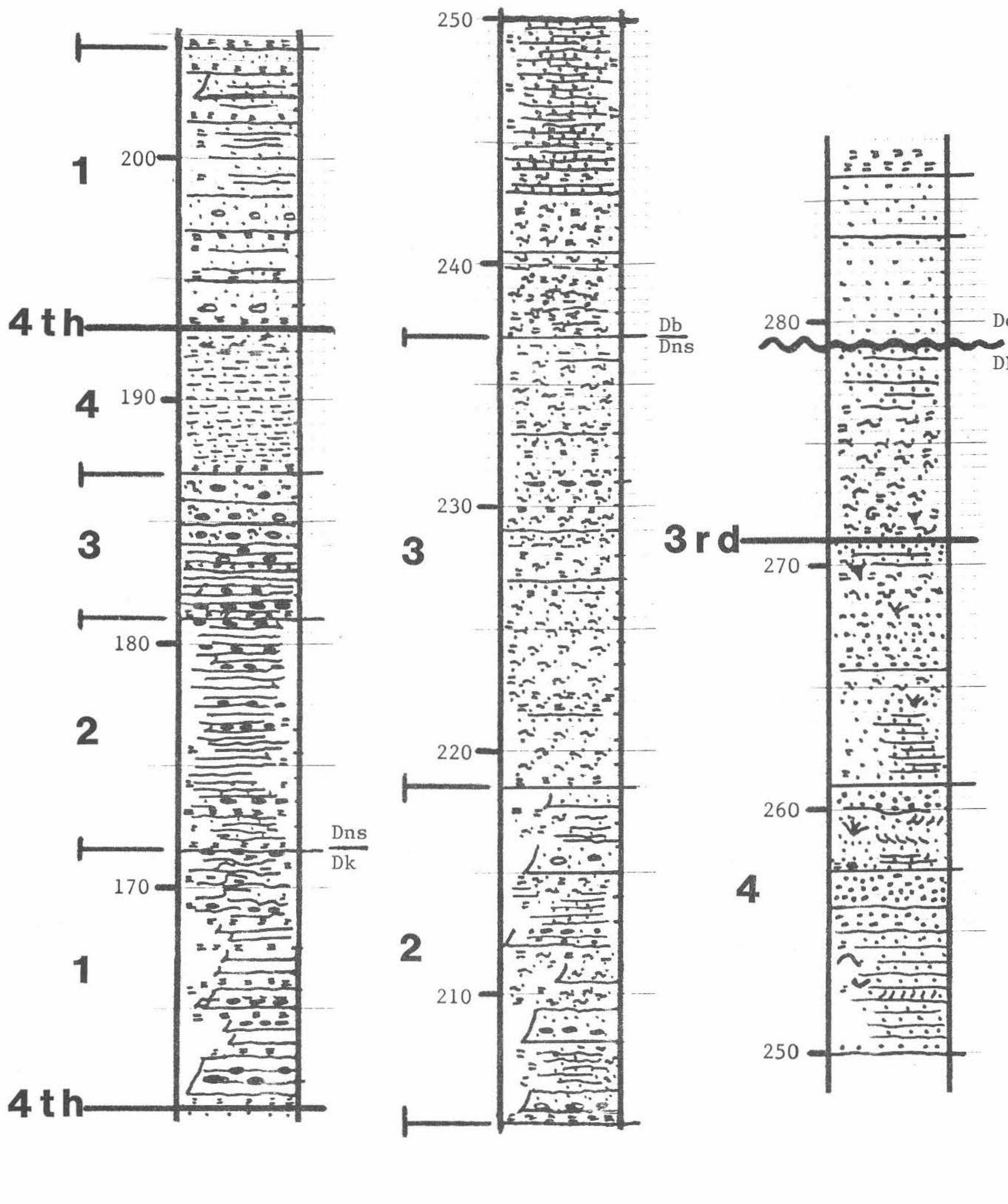


Figure 2 Log I-88 Schoharie

SLF = Sea level fall Dm = Manlius Fm. Dc = Coeymans Fm.

Dk = Kalkberg Fm. H = Hannacroix Mbr. BL = Bronk's Lake Mbr.

Dns = New Scotland Fm. Db = Becraft Fm. Do = Oriskany Fm.


1
200
4th
4
190
3
180
2
Dns
Dk
170
1
4th
250
240
Db
Dns
230
3
220
2
210
280
Do
Db
3rd
270
260
4
250

<u>5th Order Cycle 4</u> is the thick calcisiltite unit.

<u>4th Order Cycle III</u>

This is the thickest of the three 4th order sequences at Schoharie. It is 66.5 feet thick and in general a regressive sequence of 5th order cycles, beginning in deep ramp and ending in shoreface facies. This unit is equivalent to the highstand systems tract of sequence stratigraphy. The top of the cycle is placed at a surface above the coarsest calcarenite facies in the Becraft, at 271 feet on the log. This surface is also a third order B sequence boundary. The Oriskany unconformity, 8 feet above this surface, has eliminated most of the last 3rd order sequence of the Helderberg supersequence (third order A). At Kingston the upper 3 feet of the Becraft and the Alsen-Port Ewen Formations comprise this 3rd order sequence. Fourth order cycle III consists of four 5th order cycles:

<u>5th Order Cycle 1</u> ends at 204.5 feet on the log and is an 11.5 foot thick set of amalgamated lime turbidites.

<u>5th Order Cycle 2</u> ends at 218.5 feet on the log just below a 4-6 inch shale and like the cycle just below, is a 14 foot thick set of amalgamated turbidites.

<u>5th Order Cycle 3</u> ends at 237 feet on the log just above a massive calcisiltite bed. Rocks above this surface are coarser and would generally be described as Becraft.

<u>5th Order Cycle 4</u> extends to the top of the 3rd order sequence and consists of five coarsening upward, calcarenite shallow ramp cycles and is the thickest of the 5th order cycles in the set.

The last part of STOP 3 is one mile farther west on I-88. Here the surface between Becraft/Alsen calcarenite and fossiliferous quartz sandstone of the Oriskany represents superimposed 2nd (i.e. Sloss sequences), 3rd, 4th, 5th and 6th order boundaries. There are two 6th order cycles in the Oriskany followed by a major deepening into basinal Esopus facies.

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Lengthy reference lists including other peoples work on the Helderberg Group and on cyclic stratigraphy are presented in the above papers.

ROAD LOG

CUMULATIVE MILEAGE	MILES FROM LAST POINT	ROUTE DESCRIPTION
0.0	0.0	Start at the central Oneonta entrance to I-88
39.0	39.0	Cobleskill Exit I-88
39.3	.3	Oriskany/Esopus, STOP 3B
40.0	.7	Kalkberg to Becraft, STOP 3A
42.5	2.5	Rondout to Coeymans, STOP 2
44.0	1.5	Schoharie Exit I-88, turn south on Route 30A
44.9	.9	End Route 30A, continue south on Route 30
46.2	1.3	Intersection with Route 443, turn east on Route 443
50.2	4.0	The village of Gallupville, continue east on Route 443

CUMMULATIVE MILEAGE	MILES FROM LAST POINT	ROUTE DESCRIPTION
51.4	1.2	Quarry in the Manlius-Coeymans on the south side of road (Rickard Locality 67)
53.8	2.4	The village of West Berne
56.6	2.8	The village of Berne, continue easat on Route 443
60.1	3.5	The village of East Berne, turn left on Route 157A
60.5	.4	Warner Lake, turn east on Route 157A toward Thacher Park
62.3	2.3	Intersection with Route 157, Thompson Lake,Continue straight northeast on Route 157
63.8	1.5	Thacher State Park, pool and recreation area on left
64.4	.6	Turn left into parking lot for Mine Lot Falls, STOP 1

Return by same route to I-88 Schoharie, go to STOP 2 and then STOP 3 as located above on the interstate.

Middle Devonian near-shore marine, coastal and alluvial deposits, Schoharie Valley, central New York State

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INTRODUCTION

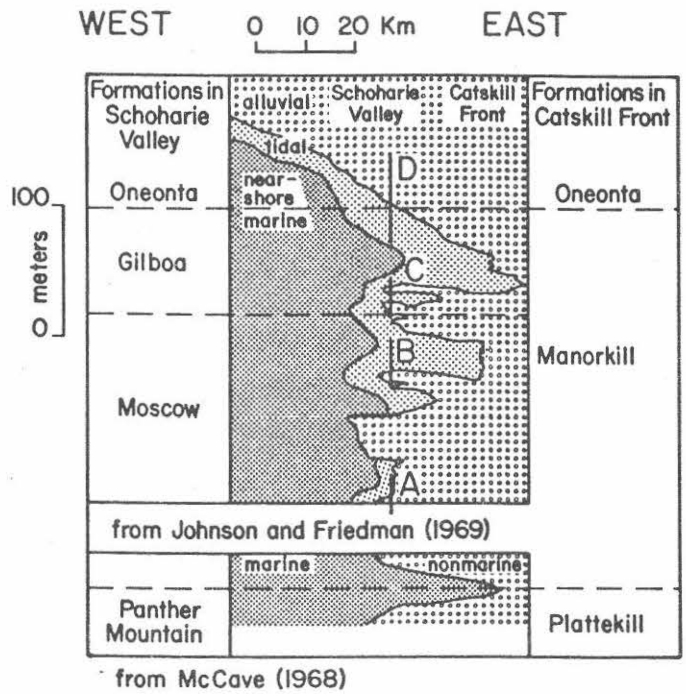
The mid-Devonian part of the Catskill clastic wedge in New York State is composed of fluvial deposits to the east along the Catskill Front and relatively thinner marine rocks to the west. Deposits of the Mid-Devonian shoreline occur in Schoharie Valley, central New York State, and provide an important record of changing marine to nonmarine depositional environments. The paleontology, stratigraphy and sedimentology of these rocks have been studied by many over a long period of time (eg. Goldring, 1924, 1927; Cooper, 1930, 1933, 1934; Cooper and Williams, 1935; McCave, 1968, 1969, 1973; Johnson and Friedman, 1969; Banks et al., 1972, 1985; Bonamo, 1977; Miller, 1979; Grierson and Bonamo, 1979; Miller and Johnson, 1981; Shear et al., 1984; Dannenhoffer and Bonamo, 1984). Sedimentologic variations within these deposits have recently been reevaluated with the goal of providing more detailed paleoenvironmental interpretations (Bridge and Willis, in press; Miller and Woodrow; in press).

The purpose of this field trip is to visit five outcrops which display the diversity of rock types exposed in Schoharie Valley. These deposits record a wave- and tide-influenced deltaic shoreline where the tidal range and degree of wave influence varied in space and time. During this trip we will focus on the relative influence of wave, tidal and fluvial currents in shaping this ancient coastline and nearshore sea bed.

STRATIGRAPHIC SETTING

Geologists have examined rocks in the Schoharie Valley since early in the nineteenth century. The current stratigraphic subdivision and correlation of the rocks (Rickard, 1975; Fig. 1) is due mainly to the work of Cooper (1930, 1933, 1934) and Cooper and Williams (1935). Formations were defined biostratigraphically in the absence of certain key species. As a result, formations are not well defined nor distinctive lithologically. This casts doubt upon lithostratigraphic correlations with the Catskill red beds to the east (Fig. 1).

Locations of outcrops and cores studied by Bridge and Willis (in press) are shown on Fig. 2. We will visit five of these outcrops on this trip. Outcrop positions projected onto a north-south line of section are given in Fig. 3. Stratigraphic dips in this area are generally to the south at approximately 1.5°. Such dip determinations allowed approximate lithostratigraphic correlations among the outcrops



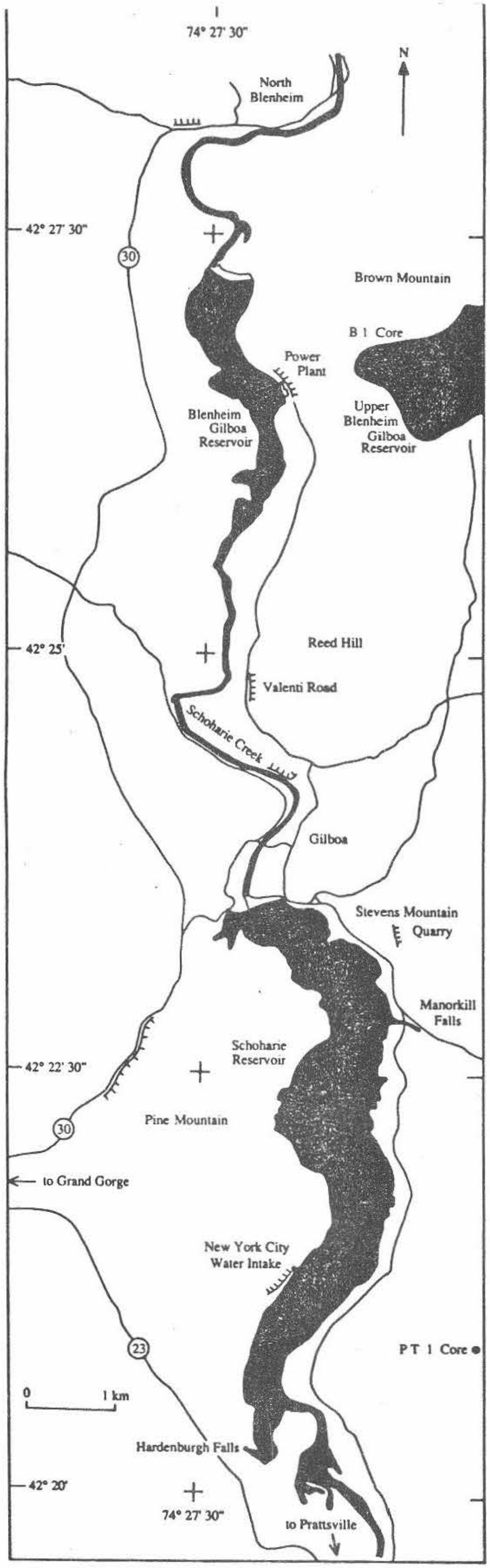


Figure 1. Stratigraphic section oriented E-W showing formations in Schoharie Valley, and their correlation with those at the Catskill Front. Interpretations of depositional environment are from Johnson and Friedman (1989) and McCave (1968). Lettered vertical sections refer to the stratigraphic positions of: (A) Schoharie Creek, near Gilboa; (B) Manorkill Falls, Schoharie Reservoir, (C) NY City Water Inlet, Schoharie Reservoir; (D) Stevens Mountain Quarry, Hardenburg Falls, and Route 30 N. of Grand Gorge. See Figure 2 for location map.

Figure 2. Location map of study area, showing position of outcrops and cores studied by Bridge and Willis (in press). Drawn from USGS 1:24,000 maps of Gilboa and Prattsville, N.Y.

and cores shown on the dip-parallel line of section (Fig. 3). Distinctive lithostratigraphic units on the order of tens of meters thick appear to correlate over distances of at least 10 kilometers. However, complex lateral lithologic variations inhibit detailed correlations within many stratigraphic intervals. The outcrops and cores include the lower part of the Oneonta Formation, the Gilboa and Moscow (Cooperstown) Formations, and part of the Panther Mountain Formation. The definition and characteristics of these formations are discussed in Bridge and Willis (in press). We have slightly revised the existing lithostratigraphic correlations with the thick fluvial successions at the Catskill Front, based on consistent vertical facies changes over 100's of meters of strata (Fig. 4).

Brachiopods from the Moscow and Panther Mountain Formations are typical Hamilton Group faunas (McGhee, pers. comm.). However, the top of the Hamilton Group (top Moscow, base Gilboa Formations) is difficult to define because characteristic fossils are absent. Preliminary palynological age determinations (Richardson, pers. comm.) suggest the Panther Mountain and Moscow Formations cover most of the *lemurata-magnificus* zones of Richardson and McGregor (1986), which corresponds to the upper *ensensis* to middle *varcus* conodont zones (early to mid Givetain). The Catskill Front alluvial sequence (which mainly includes the Moscow, Gilboa and Oneonta Formation equivalents) extends from the mid *lemurata-magnificus* to the mid *optivus-triangulatus* zones (mid to late Givetain) (Traverse et al., 1984, 1987). This is in general agreement with Rickard's (1975) biostratigraphic-lithostratigraphic chart. More detailed and extensive palynostratigraphic studies are in progress.

SEDIMENTOLOGICAL DESCRIPTION AND INTERPRETATION OF FIELD STOPS

There are complex vertical and lateral changes in the sedimentological properties of the rocks within each outcrop studied, making it difficult to describe them in a rigid facies framework. Although there are broad similarities among some of the outcrops, paleoenvironmental interpretation of these deposits rests critically upon integrating vertical and lateral sedimentological variations at individual outcrops. On this trip five particularly well exposed outcrops will be visited. Together they provide a representative view of depositional environments associated with the ancient Catskill shoreline.

STOP 1: Hardenburg Falls

Description

The lowest 6 meters of the exposure at Hardenburg Falls (Fig. 5) contains interbedded gray sandstones, dark-gray siltstones and shales. Lenticular-wavy bedded parts contain very-fine grained sandstone beds which fine up to mudstone. These sandstone beds are wave-ripple cross-laminated and have wave-ripple forms on their tops. Thicker sandstone beds (typically 0.1 to 0.5 meters-thick) have sharp, erosional bases overlain by intraformational shale fragments and disarticulated shelly fossils. They are fine to very-fine grained and fine upwards. Internal structures change upwards from

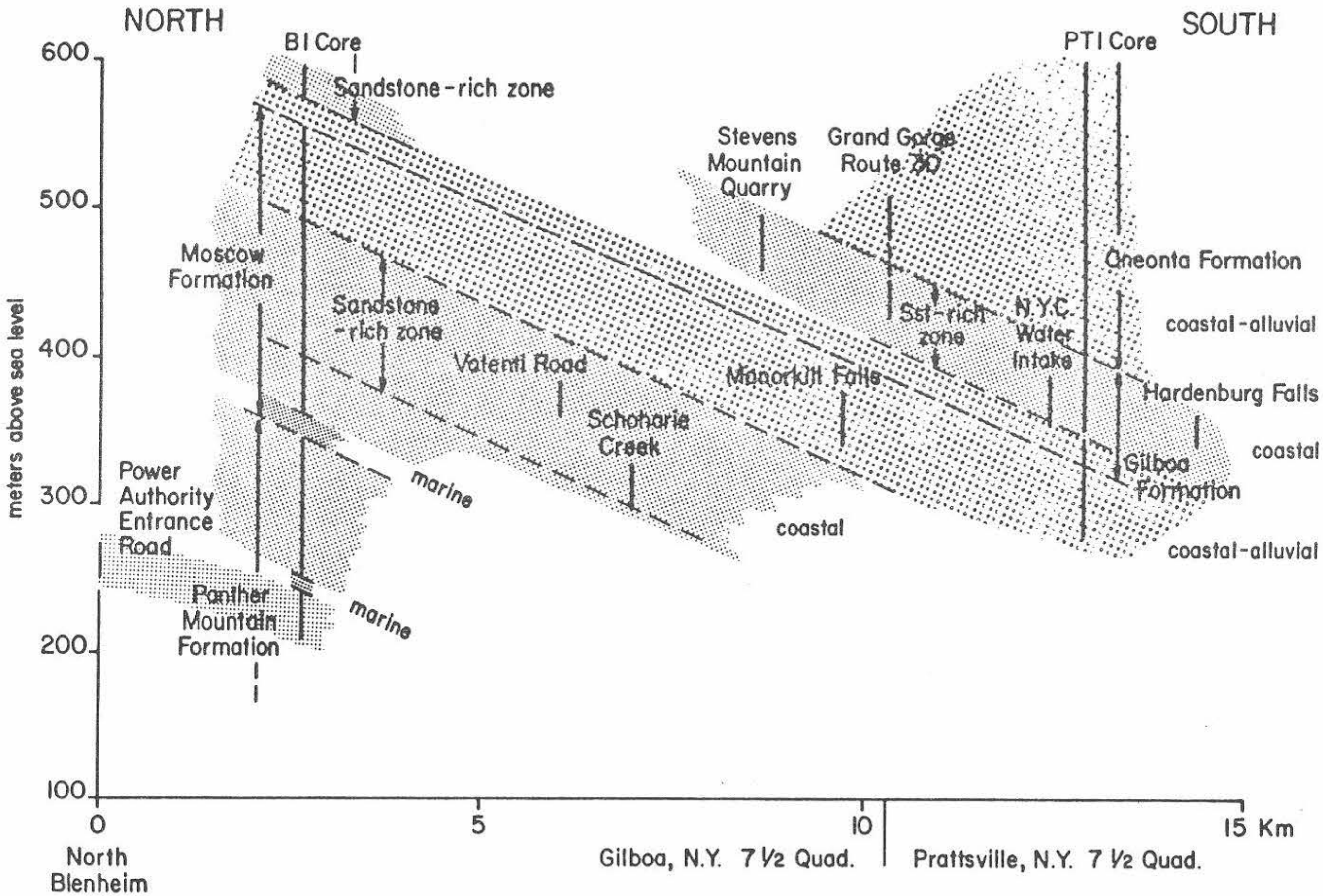


Figure 3. Stratigraphic section oriented N-S showing positions of outcrops and cores, and correlation of formations and other distinctive lithostratigraphic units within Schoharie Valley. Line of section approximately follows Blenheim-Gilboa and Schoharie Reservoirs (Fig. 2). Correlations were made based on a regional dip to the South at 1.5°. Broad paleoenvironmental interpretations are shown. The wide spacing of outcrops and cores precludes more detailed interpretations.

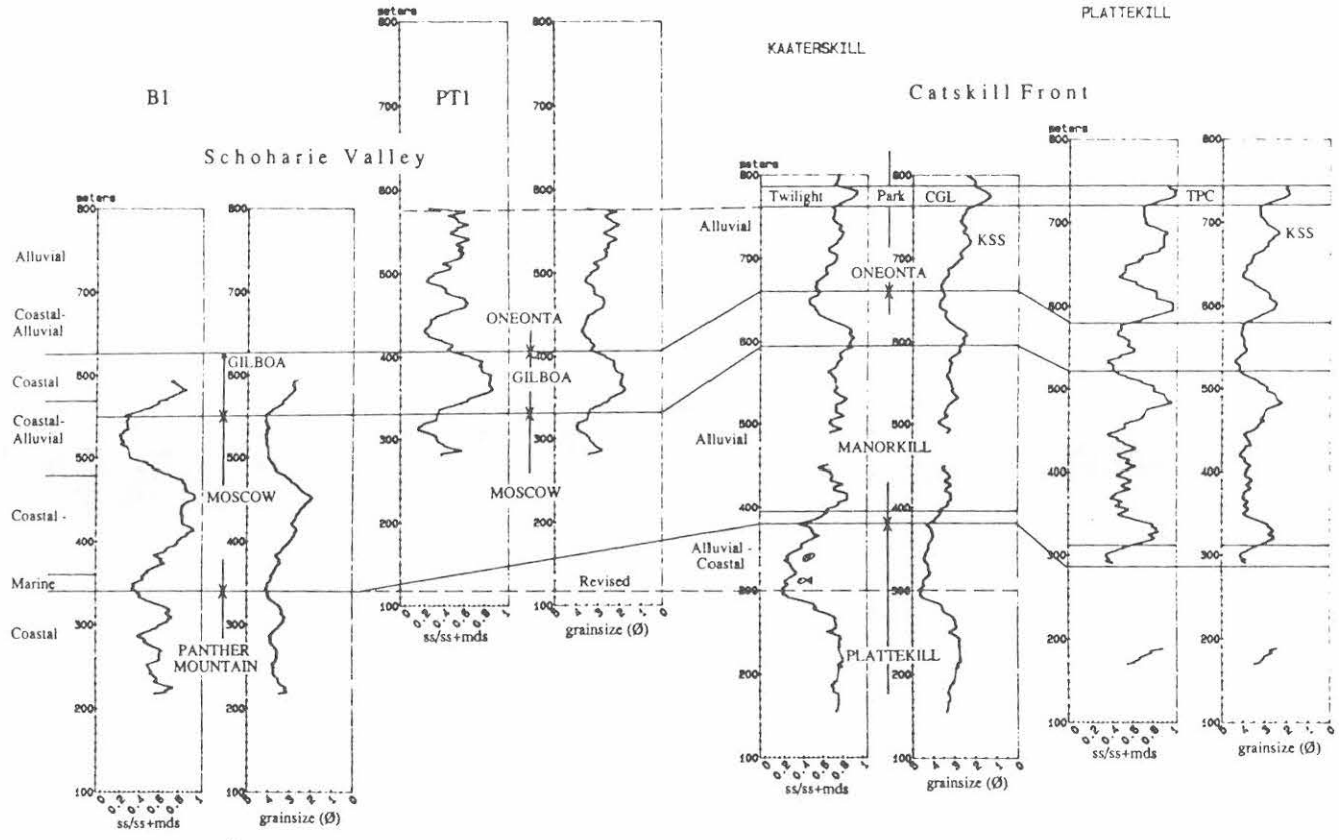
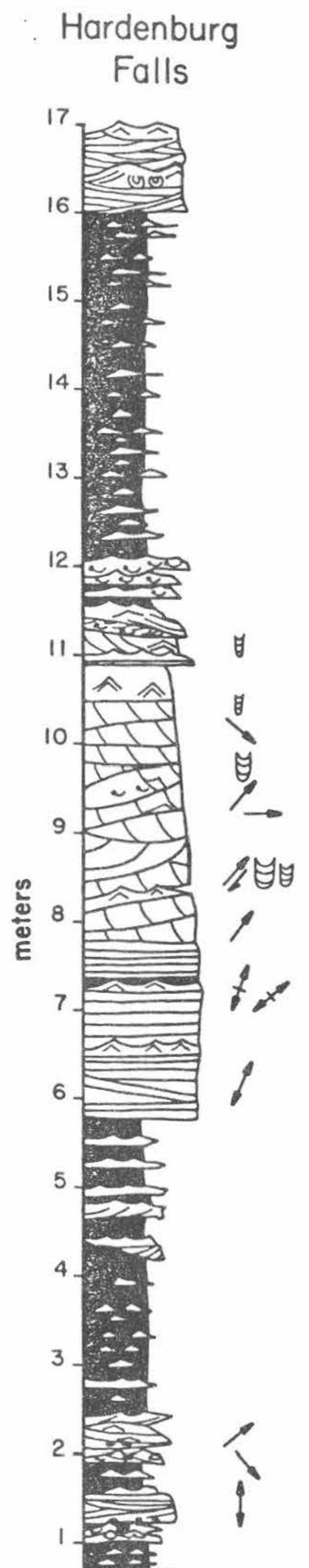
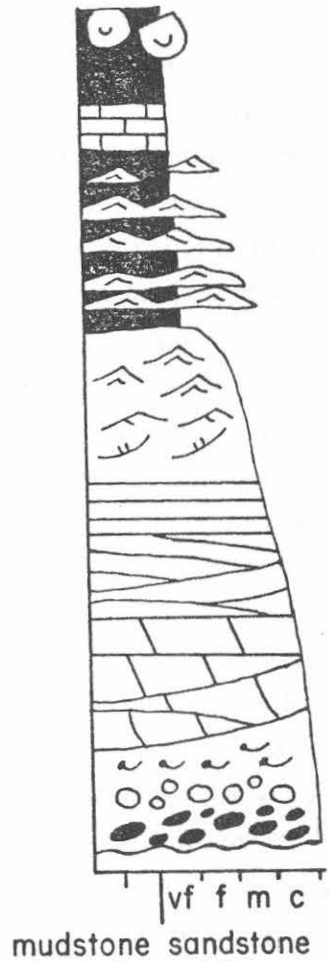


Figure 4. Revised lithostratigraphic correlations between Schoharie Valley and the Catskill Front, based on the proportion of sandstone and mean grain size in long cores and continuous stream sections. Existing formation boundaries and their correlations are solid lines. Schoharie Valley data are from this paper. Catskill Front data are from Willis and Bridge (1988). TPC = Twilight Park Conglomerate. KSS = Kaaterskill Sandstones.

Figure 5. Sedimentological log of Hardenburg Falls and key to sedimentological logs.



LEGEND



Load structures

Mudstone

Calcareous mudstone

Lenticular/wavy bedding with wave or current ripple marks

Flaser bedding

Small-scale cross-strata with wave ripples

Small-scale cross-strata with current ripples

Planar-strata

Hummocky/swaley cross-strata

Large-scale planar cross-strata

Large-scale trough cross-strata

Shell breccia
Extraformational pebbles
Intraformational breccia
Erosion surface

Fish fragments

Vertical and horizontal burrows

Meniscate burrows of various sizes

Chondrites

Spirophyton

U-shaped burrows

Branching burrows

Tree trunk cast

Plant fragments

Root casts

Rhizoconcretions

Calcareous nodules

mudcracks

= Fissile

Blocky texture

Disrupted

Pseudoanticlines

Unidirectional paleocurrent direction (N to top of page)

Orientation of parting lineation

Oscillatory paleocurrent directions

hummocky cross-strata or planar-strata to wave-ripple cross-lamination. Tops of beds are covered with wave-ripple marks (3-D, 2-D, and interfering types). Two-dimensional wave ripples indicate SW-NE paleocurrent oscillation. Some low-angle, hummocky cross-strata with superimposed wave-ripples and abundant shell fragments show a preferred paleocurrent direction to the NE. Rarely, such relatively coarse grained, shell-rich strata contain angle of repose cross-strata with SE paleocurrents. At the base of the section, amalgamated hummocky cross-stratified beds fill N-S oriented channels up to decimeters deep and meters wide. These beds contain rounded extraformational chert and quartzite pebbles, and intraformational shale clasts within a medium sandstone matrix.

Shell accumulations include the brachiopods *Spinocyrtia, Mucrospirifer, Mediospirifer,Orthospirifer, Cupularostrum*, the bivalves *Goniophora, Actinopteria, Palaeoneilo*, and crinoid ossicles. Most of the disarticulated shell fragments lie concave down along stratification surfaces and show little evidence of breakage or abrasion of fine detail. Some shell concentrations are markedly monospecific (e.g. *Cupularostrum*). Small, centimeter diameter vertical burrows occur in the tops of sandstone beds.

The 5 to 6 meter thick sandstone body in the middle of this section has a sharp base which is overlain by 2 meters of planar-stratified and swaley cross-stratified, fine to medium grained sandstone. These beds contain horizons of symmetrical and asymmetrical wave ripples with shale drapes. Parting lineation indicates NNE-SSW paleocurrents and asymmetrical wave ripples indicate a NE-SW oscillation with a preferred migration to the SW. These beds are overlain by about 2 meters of trough cross-stratified sandstone. Paleocurrents are towards the NE mainly, with the exception of a rare, oppositely directed cross-set. Wave and current ripples are superimposed on the large-scale cross-strata. One set of small-scale cross-strata indicates an easterly paleocurrent. Large-scale planar cross-strata occur above the trough cross-strata and climb up low-angle lateral-accretion surfaces; that is, inclined bedding surfaces dip to the NW (normal to the underlying trough cross-strata) and the planar cross-strata within beds dip to the SE. Vertical, meniscate burrows of varying diameter occur in this part of the sandstone body. The upper part of the sandstone body is wave-ripple laminated, very-fine grained sandstone, which contains sparse shell fragments. This is overlain by several beds of hummocky cross-stratified and wave-rippled, fine-grained sandstone with abundant shell fragments.

Interpretation

Interbedded shales and sandstones containing hummocky cross-strata, planar-strata, wave-ripple cross-laminae, wave-ripple marks, and abundant shelly-fossil concentrations are interpreted as nearshore marine deposits formed below fair-weather wave base (cf. Dott and Bourgeois, 1982; Harms et al., 1982; Hunter and Clifton, 1982; Swift et al., 1983). The sandstones are generally interpreted to be formed during storms by combined wave and unidirectional currents, whereas the shales generally represent fair-weather deposits or those formed below storm wave base. Oscillatory flows are generally directed alongshore, whereas the stronger undirectional currents represented by low-angle to angle-of-repose cross-strata are directed onshore (easterly direction). Although the shelly fossils in sandstone beds are clearly transported, their preserved detail precludes long periods or distances of transport. McGhee and Sutton (1985) believe that such shelly-fossil concentrations reflect the diversity of the indigenous infauna and epifauna (in this case typically nearshore marine), although their relative abundances do not reflect original population densities. Monospecific shell concentrations may well reflect only slightly disturbed original communities, as it is unlikely that hydraulic sorting could so perfectly separate shells of one species from other shells with similar hydrodynamic properties.

The thick sandstone body in the middle of this exposure is interpreted as deposits of a channel mouth bar associated with a tidal channel which prograded into a storm-wave dominated sea. Swaley-hummocky cross-strata, planar-strata and wave-ripples in lower beds but angle-of-repose cross-strata in upper beds, are evidence for variation in wave and unidirectional currents upwards. Lower down within the sandstone, amalgamated hummocky-swaley cross-stratified beds and wave-ripples overlain by shale drapes indicate periodic variation in combined-flow strength. Oscillatory currents associated with wave-ripples are generally to the NE and SW (alongshore). Upper beds with trough cross-strata directed alongshore (NE) to landward (E) (but rarely in the opposite direction) are most likely associated with sinuous crested dunes formed by strongly asymmetrical tidal flood currents. The SE-directed planar cross-strata may represent wave-formed swash bars or straight-crested tide formed dunes (eg. Hayes and Kana, 1976; Boothroyd, 1978; Fitzgerald, 1984). Superimposed wave- and current-ripples indicate periodic variation in current strength, and the presence of weak wave currents. Inclined bedding surfaces exposed near the top of the sandstone testify to lateral migration of the upper bar surface. The sparse fauna and low degree of bioturbation is typical of sandy deposits in coastal areas (Gould, 1970; Howard and Reineck, 1981).

Fossiliferous sandstones and mudstones above the channel mouth bar deposit indicate rapid return to marine conditions. Depositional environments at the shoreline probably depended on the position of tide-influenced channels which distributed sediment to the shoreline and near-shore areas. Avulsion of the channel from this area abruptly reduced local deposition rates, allowing the sea to transgress locally over this location.

Stop 2: Stevens Mountain Quarry

This quarry exposure is important because it provides continuous exposures from wave-dominated marine deposits to tide-dominated channel deposits. The extensive exposures also provide a better record of lateral variations than were exhibited at the last stop.

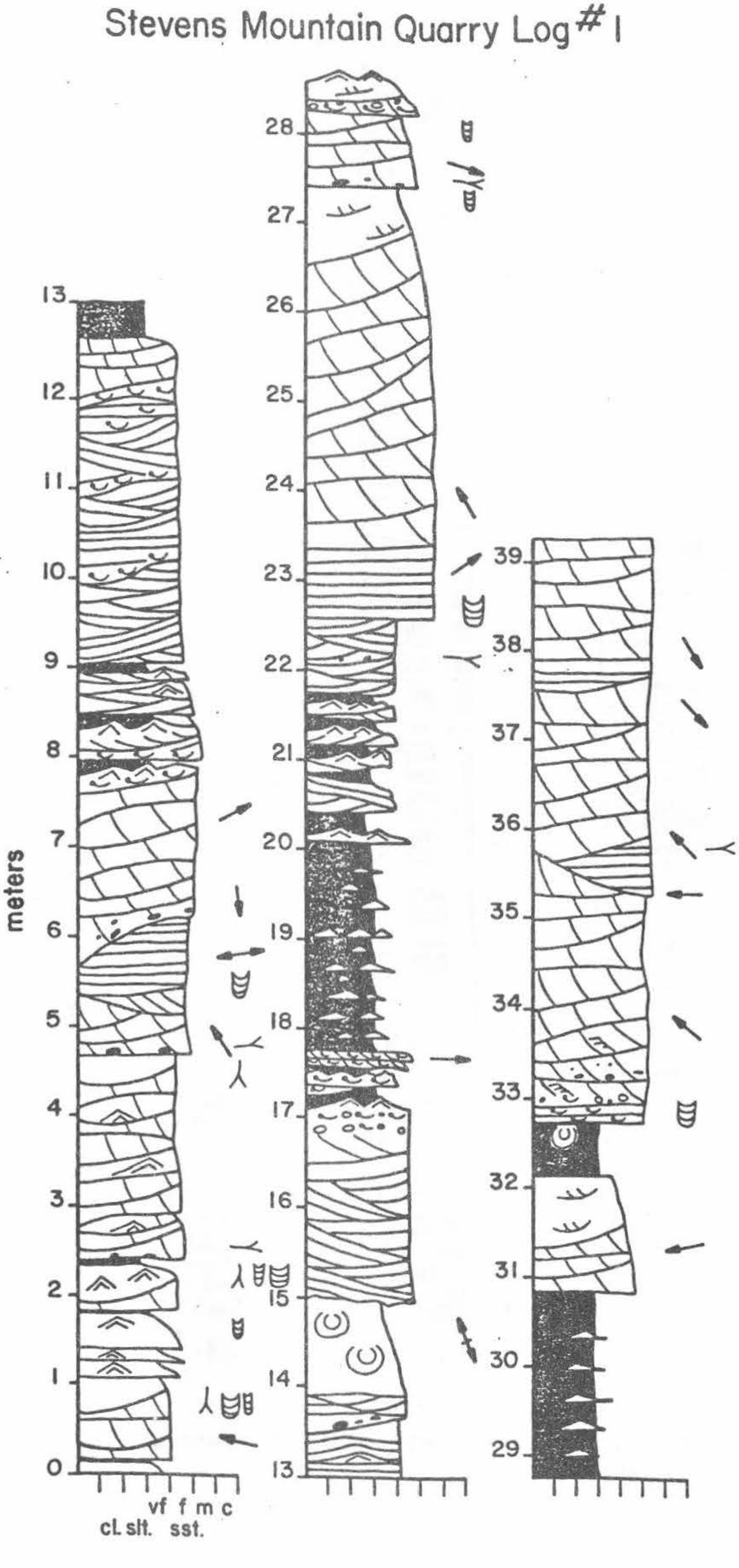
Description

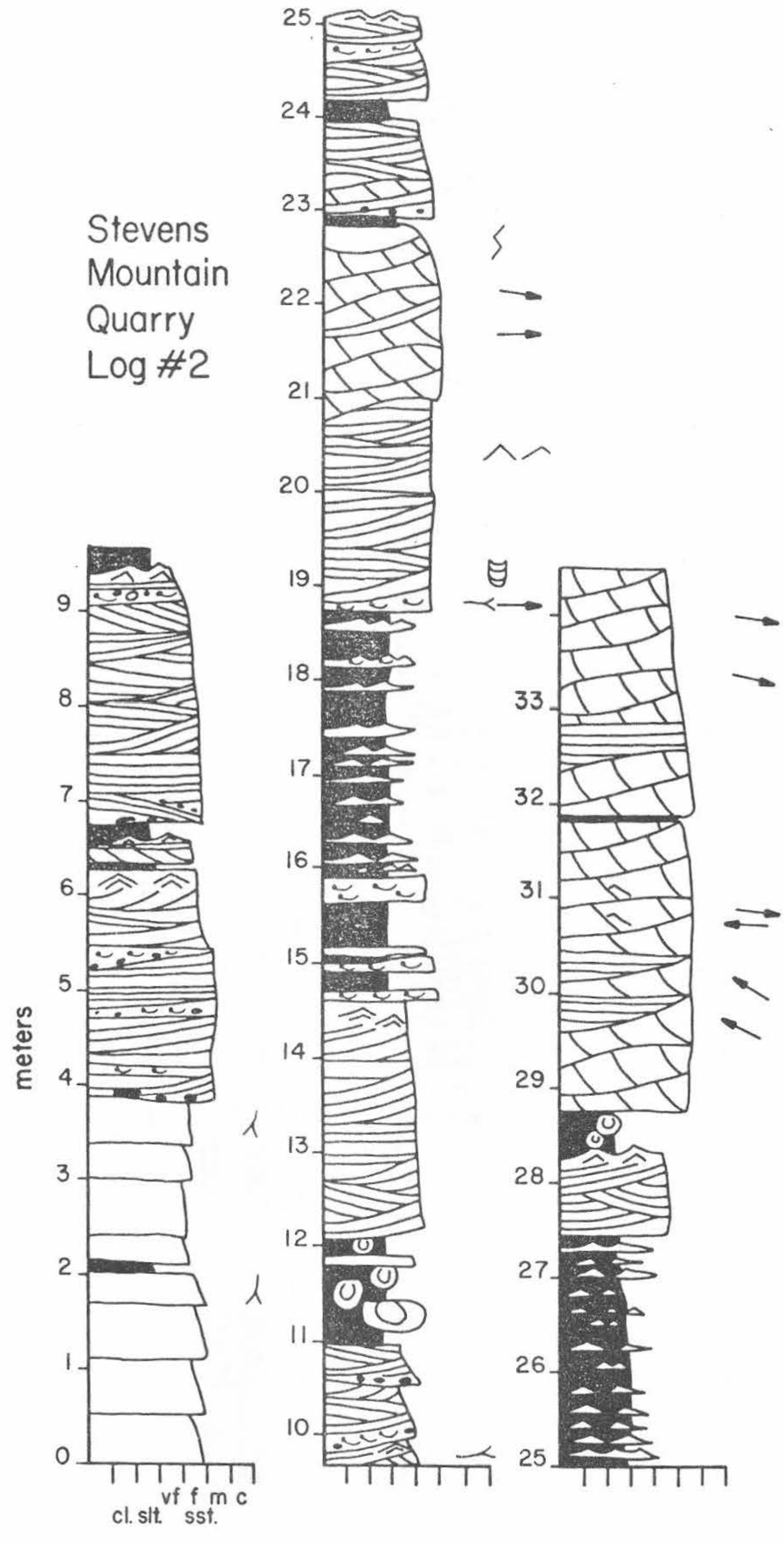
Much of the quarry exposure is composed of 2 to 6 meter thick, gray sandstone bodies separated vertically by mudstone dominated intervals of dark-gray shales interbedded with centimeter- to decimeter-thick gray sandstone beds (Fig. 6). The lowest 4 to 5 meters exposed above the quarry floor contains decimeter-thick beds of fine-grained sandstone with some centimeter-thick shale horizons. The sandstone beds have sharp bases overlain in places by drifted plant remains and shale chips. Internal structures are difficult to discern because of intense bioturbation by roots and burrows. However, large-scale cross-stratification (westerly paleocurrents) occur, and tops of beds have small-scale cross-strata and wave-ripple marks. Burrows are mainly vertical, sand-filled tubes which range in diameter from 2 to 10 centimeters: some have meniscate fills.

Within the overlying 11 to 12 meters of section (Fig. 6a, meters 5 to 17) occur several sharp-based, 2 to 3 meter-thick, fine-grained sandstone bodies separated by shale-dominated intervals with thinner sandstone beds. Bases of sandstone bodies show tool marks, and are commonly lined with plant remains, shell fragments and shale chips. The sandstone bodies are composed of decimeter- to meter-thick beds containing hummocky-swaley cross-stratification and planar-stratification. Individual beds in such amalgamated sequences are recognizable by their erosional bases overlain by shell concentrations of transported brachiopods, bivalves and *Tentaculites*. The upper beds within these sandstones are commonly shell-rich and contain intraformational shale chips and extraformational quartz pebbles. Angle of repose cross-stratification also is common in these upper beds, normally indicating an E to NE paleocurrent orientation. Wave ripples cap most sandstone bodies. The sandstone bodies vary in thickness and sediment type laterally. This change is most dramatic in the lowermost sandstone body in this succession. To the north, the sandstone body is about 3 meters thick and contains mainly large-scale cross-stratification and planar-stratification, indicating a range of paleocurrent directions (Fig. 6a, meter 5-8). To the south, this body thins to about 2 meters and contains mainly swaley cross-strata (Fig. 6b, meter 6 to 8). Shale-rich zones in the sequence contain decimeter-thick hummocky cross-stratified sandstone beds with wave-rippled tops. Load casts occur in distinct, laterally extensive horizons, particularly immediately beneath sandstone bodies.

The next beds contain 4 to 5 meters of interbedded shale and sandstone (Fig. 6a, meters 17-22). The lowermost decimeter-thick sandstone beds in this sequence (immediately above a sandstone body) are markedly lenticular, and rich in shells, intraformational breccia and extraformational quartzite pebbles. They contain hummocky cross-strata and, in places, angle of repose cross-strata with easterly paleocurrents. Wave ripple marks cap these sandstone beds. Most of the remainder of this interval is wavy and lenticular bedded, with very-fine grained sandstone occurring in centimeter-thick, wave-rippled beds. In the north end of the quarry, these lenticular-wavy beds coarsen and thicken upwards to decimeter-thick, hummocky cross-stratified, very-fine grained sandstone beds with wave-rippled tops, and

Figure 6. Sedimentological logs of the north (A) and south (B) ends of Stevens Mountain Quarry.




Stevens
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to amalgamated hummocky cross-stratified, fine grained sandstone beds with abundant plant remains and intraformational breccia.

These beds are sharply overlain by a 5 to 6 meter-thick fine to medium grained sandstone body (Fig. 6a, meters 23 to 28). It is dominated by large-scale trough cross-stratification (NW paleocurrent), with the exception of planar stratified beds above the erosional base and some small-scale cross-strata in the finer beds near the top. The top most decimeter-thick beds have abundant drifted plant remains, shells, intraformational shale and extraformational pebbles in their bases. Paleocurrents in these upper beds are to the SE. Large vertical meniscate burrows (5-10 centimeter diameter) occur in the lower parts of the sandstone body, but smaller (1-2 centimeter diameter) vertical burrows occur in the upper most beds. In the south end of the quarry this sandstone body is thinner (Fig. 6b, meters 19 to 23). Lower parts of the sandstone body are replaced by amalgamated swaley cross-stratified, fine grained sandstone beds, and the upper two meters are replaced by hummocky cross-stratified sandstone beds with wave-rippled tops and shale interbeds. Underlying beds are finer grained than to the north. The basal erosion surface contains flutes (E paleocurrent) and grooves, and is overlain by bivalve and brachiopod shells and plant remains. Paleocurrents in the central, coarsest part of the sandstone body are to the east.

Approximately 4 meters of interbedded sandstone and shale overlie this sandstone body (Fig. 6a, meters 29 to 33). The sequence is mainly wavy-lenticular bedded with centimeter-thick wave-rippled, very-fine grained sandstone beds. Within this succession to the south is a 0.9 meter-thick, hummocky cross-stratified sandstone bed with an erosional base and wave-rippled top (fig. 6b, meter 18). It occurs at the top of a coarsening/bed-thickening upward sequence. To the north, this bed changes laterally to a 1.3 meter-thick, fine to very-fine sandstone bed with large-scale cross-strata at the base (westerly paleocurrent) and small-scale cross-strata at the top (Fig. 6a, meter 31 to 32). This bed is in turn cut out laterally by a fine-grained channel fill. At the top of this sequence, immediately below the top most sandstone body, is an extensive zone of load casts.

The top most sandstone body in the quarry is approximately 6 meters thick (Fig. 6a, meters 33 to 39). The erosional base is overlain by intraformational breccia (shale blocks up to decimeters across) with brachiopod and bivalve shells in a medium sandstone matrix. The sandstone body is medium to fine grained and fines upwards in places. It is mainly large-scale trough cross-stratified, with some lenses of planar-stratification. Reactivation surfaces in the cross-strata of the lower part of the sandstone body are mainly erosional. Between laterally adjacent reactivation surfaces, cross-strata vary in dip. In places oppositely directed, asymmetrical ripple marks are preserved on foresets and reactivation surfaces, and shale drapes may also occur. Large vertical meniscate burrows also occur in these lower beds. Paleocurrents associated with large-scale cross-strata are towards the WNW in the lower half of the sandstone body, but towards the ESE in the upper half. Low angle, lateral-accretion surfaces extend through the sandstone body, and a sandstone filled channel cuts through the upper one third.

Interpretation

The sandstone body at the top of this exposure is interpreted as the deposits of a laterally-migrating tidal channel (see also Johnson and Friedman, 1969). River channels preserved in coeval Catskill facies to the east have generally north-westerly paleocurrents, but range from northerly to southerly (Willis and Bridge, 1988). Therefore paleocurrents with a westerly component are in the tidal ebb and river flow direction, whereas those with an easterly component are in the tidal flood direction. Variation in ebb paleocurrent orientation in various examples of this type of sandstone body suggest that channels had a range of orientations due to local channel curvature and/or a distributive channel pattern. Dominance of ebb-directed paleocurrents lower in the sandstone bodies, with flood currents and channel fills in upper parts is typical of the strongly asymmetrical tidal currents in tidal inlets, estuaries, and tide-influenced deltaic distributaries.

Variable dip angles of large-scale cross-strata, erosional reactivation surfaces and superimposed current ripples are further evidence of strong tidal-current asymmetry. They indicate growth of sinuous-crested dunes to 'full vortex' stage during the dominant tidal current, slackening of the current and modification of dune geometry, then erosion of the dune by the subordinate tide. Such features are commonly reported from mesotidal and macrotidal estuaries (eg Terwindt, 1981; Boersma and Terwindt, 1981; Dalrymple, 1984; DeMowbray and Visser, 1984). General absence of tidal bundle sequences and rarity of current ripples on reactivation surfaces suggests an erosional or nondepositional subordinate flood-tidal current and a dominant ebb-tidal current that may be reinforced by a fluvial current. Bioturbation is characteristically rare in such coastal sandstone bodies (eg. Howard and others, 1975). Large, vertical meniscate burrows are similar to those reported in fluvial and coastal channel deposits elsewhere (Thoms and Berg, 1985; Miller, 1979; Bridge, Gordon and Titus, 1986) and may be due to upward escape of bivalves. Brachiopod-bivalve fauna sparsely distributed through the sandstone bodies is also typical of nearshore communities (McGhee and Sutton, 1985; Sutton and McGhee 1985).

Sandstone bodies just below this top most sandstone body, with swaley-hummocky cross-strata, planar-strata and wave-ripples in lower parts, but angle-of-repose cross-strata in upper parts, are comparable to the sandstone body exposed in Hardenburg Falls (Stop 1). They are also interpreted to record evidence for combined wave and unidirectional currents of varying importance and intensity. Detailed interpretation of these deposits hinges critically on their lateral transition to thicker sandstone bodies dominated by angle-of-repose cross-strata and thinner sandstone bodies dominated by amalgamated hummocky-swaley cross-strata. It is also important that their bases are sharp in many places, but in others they occur at the top of coarsening-upward sequences with interbedded hummocky cross-stratified sandstones and shales immediately beneath. Furthermore, any evidence of channel fills or lateral-accretion bedding is restricted to their upper parts. These features suggest deposition on channel mouth bars which were prograding rapidly into a marine, storm-wave-dominated area. Such bars may be on the seaward side of tidal inlets (i.e. ebb-tidal deltas) associated with estuaries or barrier-beach shorelines, or associated with tide-influenced deltaic distributaries. Landward-directed (E, NE) trough cross-strata are most likely associated with sinuous crested dunes formed by strongly-asymmetrical tidal flood currents. Sets of planar cross-strata migrating to the NE, E or SE may represent wave formed swash bars or straight-crested tide-formed dunes (e.g. Hayes and Kana, 1976; Boothroyd, 1978; Fitzgerald, 1984). Wave and current ripples on dune/bar slipfaces and troughs indicate periods of tidal slack water and decreased tidal current strength, respectively. Sparse fauna and low degree of bioturbation (but dominance of the *Skolithos* ichnofacies) is typical of sandy deposits in coastal areas (Gould, 1970; Howard and Reineck, 1981).

Similar sandstone bodies in somewhat younger Frasnian rocks in the Binghamton (NY.) region were also interpreted as channel mouth bar deposits by Halperin and Bridge (1988). However, similar sequences reported from the Cretaceous Western Interior Seaway have been interpreted as due to deposition on prograding strandplains cut by estuaries (McCrory and Walker, 1986). McCrory and Walker (1986) and Hamblin and Walker (1979) interpreted amalgamated swaley-hummocky cross-stratified beds as beach and shoreface deposits, even though they rest sharply on interbedded hummocky cross-stratified sandstones and shale interpreted as offshore storm-dominated deposits. They interpreted this sharp contact as due to rapid progradation of the shoreface. Similar sandstone bodies occurring at the tops of gradational coarsening upward sequences have been interpreted as due to progradation of a wave-dominated deltaic shoreline (Chan and Dott, 1986; Swift et al, 1987). Swift et al.(1987) interpreted the amalgamated hummocky cross-strata as middle shoreface deposits, and angle-of-repose cross-strata as deposits on the upper shoreface by combined currents with an alongshore unidirectional component. A beach face origin for the sandstone bodies studied here is considered unlikely because of a lack of characteristic seaward-dipping planar-laminae with heavy mineral layers; alongshore directed cross-strata overlain by landward-directed cross-strata and planar-strata typical of ridge and runnel systems; and eolian cross-strata. The sharp base to many (but not all) of the sandstone bodies is hard to explain by rapid progradation of a beach face, but may be readily explained by rapid progradation of a storm-wave modified channel mouth bar which formed rapidly as a result of a channel diversion. Coleman and Prior (1982) show examples of distributary mouth bar sands that lie abruptly on offshore silts, and state that bases of sandstone bodies become sharper closer to the distributary channel.

Relatively thin sandstone bodies dominated by amalgamated swaley-hummocky cross-stratified beds are interpreted as more distal parts of channel mouth bars, which are completely dominated by storm waves (combined currents). The lack of mud suggests deposition above fair weather wave base. The common occurrence of load casts near their bases suggests rapid deposition of sand on mud. Such soft-sediment deformation features are common offshore from Mississippi delta distributaries (eg. Coleman and Prior, 1982). Paleocurrent indicators in the base of the sandstone bodies suggest a dominantly offshore directed unidirectional current, as is common in similar Frasnian rocks in the

Catskill region (Craft and Bridge, 1987; Halperin and Bridge, 1988). Lenticular sandstone beds on top of these sandstone bodies, with abundant shelly fossils, intraformational breccia, extraformational pebbles, and local channelling, are evidence of relatively strong unidirectional currents directed to the east and southeast (i.e. shoreward) superimposed upon wave currents. As there is no evidence of offshore directed currents, the unidirectional currents are not considered to be tidal flood currents but rather storm-wave associated currents. As these sandstone bodies are sharply overlain by interbedded sandstones and shales interpreted as deeper water nearshore deposits, these uppermost shelly sandstone beds are apparently associated with rapid abandonment of the channel mouth bar and reduction in sand supply. Johnson and Friedman (1969) interpreted these sandstone bodies as nearshore subtidal bar deposits, and Miller and Woodrow (in press) assigned an estuary mouth shoal origin.

Similar sandstone bodies elsewhere have been interpreted as wave-formed offshore bars or shoals (eg. De Raaf et al, 1977; Cotter, 1985). Wright and Walker (1981) have interpreted thin pebble layers at the tops of such sequences as transported by density currents. This is considered unlikely here. If such sandstone bodies do represent distal parts of storm-wave modified channel mouth bars, then their occurrence within sequences of deeper water sandstone and shales indicates channel switching (see also Swift et al, 1987). The overall sequence at Stevens Mountain Quarry indicates sandstone body deposition in progressively more landward parts of the channel-mouth bar complexes. Thus, this is an overall regressive sequence with superimposed variations in deposition related to local channel avulsion.

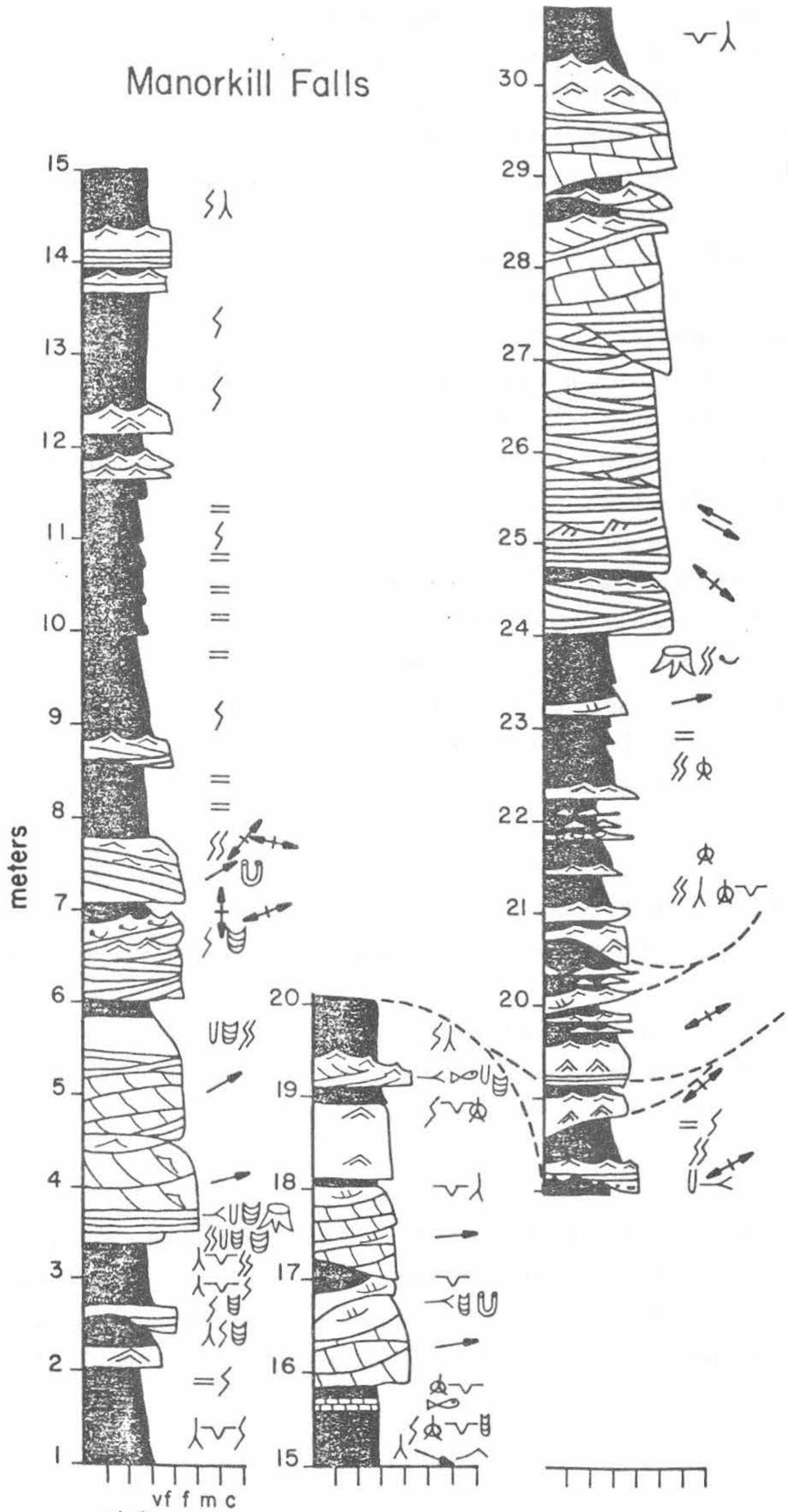
As implied above, interbedded shales and sandstones with hummocky cross-strata, planar-strata, wave-ripple cross-laminae wave-ripple marks, and abundant shelly-fossil concentrations are interpreted as nearshore storm-dominated marine deposits. These are interpreted in a similar way to those exposed within Hardenberg Falls (Stop 1). Sandstone beds near the base of the quarry were deposited mainly by N-NW directed unidirectional currents moving sand as dunes. Wave ripples in the upper parts of sandstone beds indicate deposition in standing water with weak wave currents. These characteristics, plus the abundance of burrows and roots, and the absence of marine fossils, suggest deposition by fluvial or ebb-tidal currents on a near coastal plain.

Stop 3: Manorkill Falls, Schoharie Reservoir

Description

Most of this section is composed of gray sandstone interbedded with red, green and gray siltstone and mudstone (Fig. 7). The lower part of this section contains two 10 meter-scale fining upward sequences (Fig. 7, meters 1 to 24). The lower 4 meter-thick interval within the first sequence has a high proportion of sandstone which sharply overlies intensely disrupted mudstones. These sandstone beds generally thin and fine upwards. Individual sandstone beds are sharp-based, sheetlike to lenticular, and centimeters to a meter thick. Thinner sandstones are very-fine grained, and small-scale cross-stratified with associated current ripples or (mainly) wave ripples. Thicker sandstone beds are

Figure 7. Sedimentological log of Manorkill Falls section, east side of Schoharie Reservoir.



mainly fine grained, and contain large-scale trough cross-strata, planar-strata, or (unusually) hummocky cross-strata passing upwards to small-scale cross-strata and associated current or wave ripples. Some thicker sandstone beds lower in the sequence show low-angle (<10°) inclined bedding surfaces extending throughout most of the thickness of the bed which dip in the direction the bed thins (i.e. to the east). Upper parts of these inclined surfaces are covered with straight crested and interfering wave-ripple marks, indicating oscillatory paleocurrents NE-SW and E-W. Oscillatory paleocurrent directions are quite variable throughout the sequence; however, a NE-SW orientation is common. Large- and small-scale cross-strata and current ripples consistently indicate an easterly paleocurrent.

Sandstone beds contain vertical burrows (including *Skolithos*), which range in diameter from a few millimeters to 7 centimeters; larger ones commonly showing meniscus structure. Horizontal meniscate burrows occur in places in the tops of sandstone beds. *Arenicolites* burrows also occur. Drifted plant remains are common in sandstones, and root casts occur in the top of some sandstone beds. There are two important occurrences of tree-trunk casts. Shelly fossils are rare. One shell-rich layer near the base of the section occurs near the top of a hummocky cross-stratified sandstone bed. This layer contains bivalves, brachiopods (eg. *Spinocyrtia*, *Allanella tullius*) juvenile crinoid fragments and bryozoa. Fish fragments are found in places.

Mudstones are either relatively unbioturbated and fissile, or bioturbated with abundant rootcasts, small vertical and horizontal burrows, and desiccation cracks. A nodular carbonate bed occurs near the top of the first sequence where calcareous rhizoconcretions are particularly abundant. This bed is similar to the one occurring at the New York City Water intake building.

The base of the second fining upward sequence is another sandstone-dominant interval (about 3 meters-thick), in this case dominated by large-and small-scale cross-strata (Fig. 7, meter 16 to 19). This interval is in turn overlain by an interval of overlapping channel fills. These channels are oriented E-W and are filled with alternating mudstone and sandstone beds. The sandstone beds are dominated by wave-ripple lamination, with rare planar and hummocky strata. Mudstones occurring adjacent to and above these channel fills are highly disrupted by mudcracks, burrows, rootcasts and in one location a tree trunk cast.

The sandstone body at the top of this section is similar to those occurring near the top of Stevens Mountain Quarry and the lower Grand Gorge route 30 section, and shows lateral-accretion surfaces. Paleocurrents near the base of the sandstone body are NW-SE.

Interpretation

In the Manorkill Falls section, desiccated, bioturbated siltstones with abundant root casts, rare tree trunk casts, and horizons of calcareous concretions, clearly represent floodbasins with a well developed flora which were periodically exposed and subjected to soil-forming processes. Sandstone beds overlying such deposits near the base of the section are distinctive and show evidence of easterly progradation of wedges of sand into at least temporarily standing water in which wave currents were active. The lowest bed buried and preserved the trunk casts. Two origins for this bed is considered possible: (1) a fluvial overbank sand splay, or (2) a back barrier washover deposit. Paleocurrent evidence supports the storm washover origin and deposits show many features in common with modern examples (e.g. Schwartz, 1982). However, there is little evidence in these rocks of the presence of beaches, so barrier islands seem unlikely. Although lateral migration of tidal inlets or other types of coastal channels may have removed the evidence of beaches, an overbank sand splay from a major channel is probably the most viable explanation.

Associated with the sandstone-dominated interval near the base of the section are sandstone beds showing evidence of strong wave currents, containing marine shelly fauna, and with a diverse assemblage of burrowing organisms. These deposits testify to the close proximity of the sea, and the onshore transport of marine fossils by storm waves. Furthermore, the overlying fissile siltstones with isolated hummocky cross stratified and wave-rippled sandstone beds indicate deposition of mud in quiet, standing water which was periodically agitated by waves. All of this evidence together suggest an upward transition from fluvial floodbasin/lacustrine conditions to a brackish interdistributary bay with access to fully marine waters.

A return to subaerial floodbasin conditions is signalled by desiccated mudstones with root casts, rhizoconcretions, fish fragments and a nodular carbonate-rich horizon (beds directly under the bridge; Fig. 7, meters 15-16). The overlying sandstone beds at the base of the second 10 meter-scale sequence record periodic deposition by temporally-waning unidirectional currents, probably associated with river floods. Desiccation cracks in the siltstones capping these sandstone beds and the lack of wave ripples, suggests subaerial exposure immediately following deposition. This sequence of interbedded sandstones and siltstones is interpreted as a crevasse-splay complex, but it is hard to dismiss flood tidal currents (storm enhanced?) as a depositional agent in view of the overall coastal setting.

The overlying sequence of interbedded sandstones and shales also has features indicative of mud deposition in an area of shallow water (but with periodic subaerial exposure) in which weak wave currents periodically deposited sand, and transported marine fossils into the area. The east-west oriented channel in this part of the section were apparently filled mainly by weak wave currents. The lack of evidence of reversing unidirectional sedimentary structures precludes a tidal-channel origin, and suggests instead a fluvial channel cutting across a floodbasin/interdistributary bay. Indeed, deposits typical of intertidal flats in mesotidal or macrotidal settings are notably absent throughout this outcrop. The sandstone body at the top of this section is similar to others viewed at previous stops and are interpreted as a channel-mouth bar deposits. Johnson and Friedman (1969) interpreted these deposits as interfingering alluvial marsh, tidal flat and channel deposits. They are also somewhat similar to the 'muddy shoreline' deposits of Walker and Harms (1971).

The 10 meter-scale sequences exposed at Manorkill Falls reflect relative sea level changes. Mudstone beds immediately below sandstone-rich intervals reflect floodplain emergence. Abrupt onset of sand deposition following fining indicates an abrupt change in depositional environment and the onset of a deepening trend. Such sequences may be related to channel switching which locally varied deposition rate at the shoreline.

Stop 4: Schoharie Creek outcrop

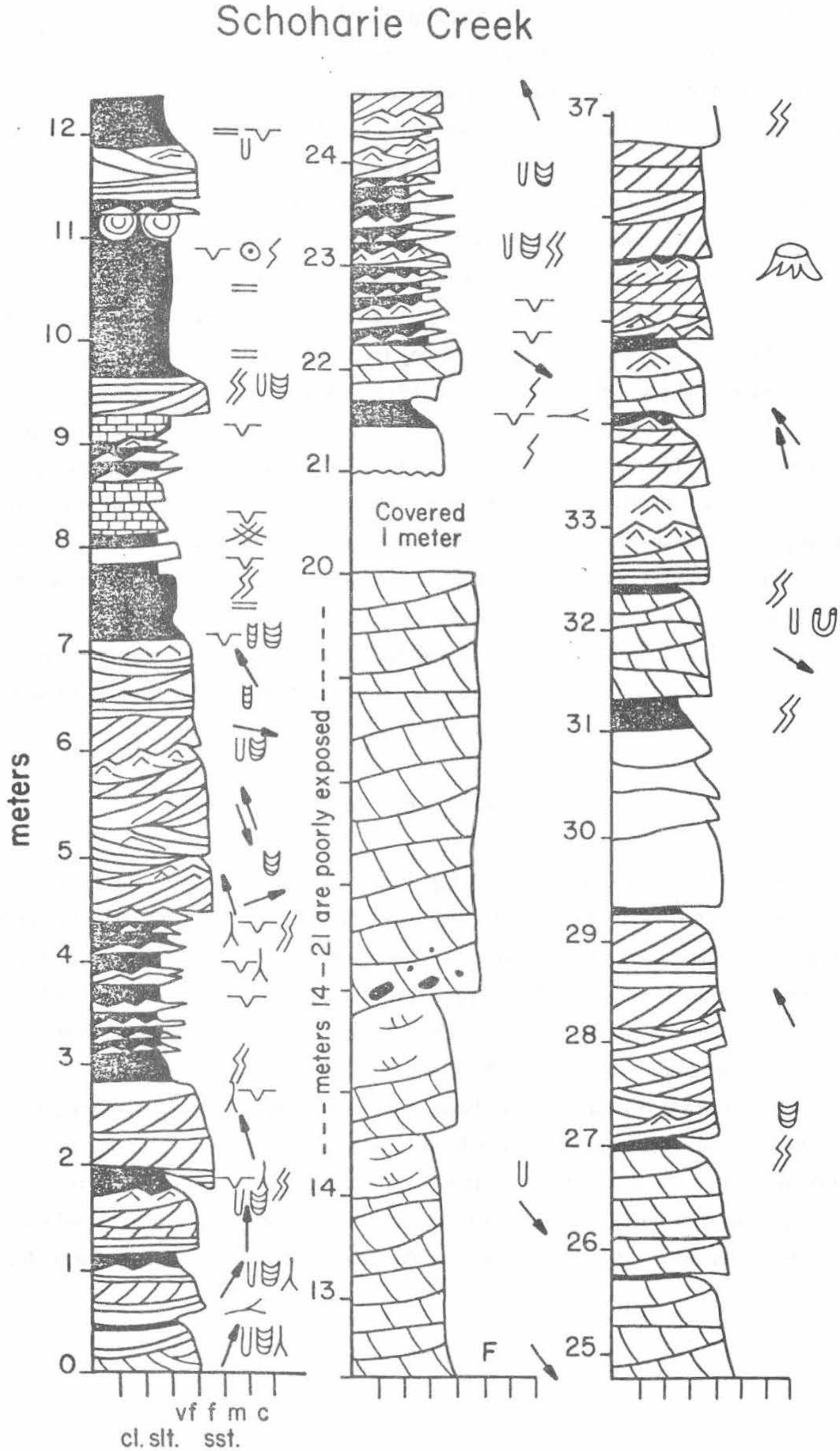
Description

The lowermost 11 meters of interbedded sandstone and mudstone in this stream and hillside section is, with the exception of the 2 to 3 meter thick sandstone body (Fig. 8, meters 4.5 to 7), similar to the lower parts of Manorkill Falls section (Fig. 8). Centimeter-thick, very-fine grained sandstone beds are mainly wave-rippled, whereas the decimeter-thick, fine-grained sandstone beds may also contain large-scale cross-strata (paleocurrents to N), planar-strata, and hummocky cross-strata. Tops of sandstone beds are normally wave-rippled and bioturbated by root casts and/or vertical, horizontal and U-shaped burrows of various sizes (millimeters to centimeters in diameter). Gray siltstones are bioturbated by roots and small burrows, and desiccation cracks are common. Calcareous concretions occur locally, and buff calcareous shale commonly fills desiccation cracks. There are several carbonate-rich mudstone beds near the top of this interval, in one case associated with pseudoanticlines. Some siltstones are relatively undisturbed and fissile.

The 2 to 3 meter thick, fine-grained sandstone body near the base of the section is somewhat unusual. It is dominantly large-scale trough cross-stratified, with some large-scale planar cross-strata and planar-strata. Large-scale cross-strata indicate paleocurrents to NNW and in a generally easterly direction. Wave-and current-ripples occur on reactivation surfaces, the current-ripples oriented in the opposite direction to the adjacent large-scale cross-strata. Ripple-marked surfaces throughout the sandstone body show vertical and horizontal meniscate burrows of various sizes centimeters in diameter).

Above poorly exposed sandstone bodies between meters 11 and 19 (Fig. 8) there occurs a 3 meter thick interval of sandstone and mudstone beds that are similar to those lower down. Above this interval is approximately 12 meters of section dominated by decimeter-thick fine to very-fine grained sandstone beds, with only small amounts of gray mudstone. Sandstone beds have erosional bases and are mainly large-scale cross-stratified, with minor amounts of planar and low-angle cross-strata. Upper parts of the beds are commonly wave-ripple laminated, with wave rippled tops. Paleocurrents from large-scale cross-strata are mainly to the NNW, but some are to the SE. Beds are intensely bioturbated with vertical and U-shaped burrows. A tree trunk cast occurs near the top of the sequence.

Figure 8. Sedimentological log of Schoharie Creek and hillside section just down stream of Gilboa bridge.



Interpretation

The lower part of the Schoharie Creek outcrop (Fig. 8, meters 1 to 11) is interpreted in a similar way to deposits exposed in the Manorkill Falls outcrop; that is, fine grained deposition in semi-permanent, near-coastal lakes or bays, with periodic emergence, plant growth and soil formation. Sand was introduced by unidirectional and/or wave currents and reworked by waves. Although trace fossils are abundant and diverse, marine shelly fossils are notably absent.

Evidence for strong tidal currents is generally lacking, with the exception of the 2 to 3 meter thick sandstone body near the base of the section. In this sandstone body bimodal paleocurrent directions in large-scale cross-strata, and wave-and current-ripples on reactivation surfaces all indicate deposition by tidal currents. However, the topmost part of the sandstone body shows more wave influence. Its stratigraphic position between coastal bay or floodbasin deposits suggests deposition in a flood-tidal delta and/or tidal channel (c.f. Boothroyd, 1978). Wave-rippled and extensively bioturbated sandstones lower down in the section along Schoharie Creek are probably sandy tidal-flat deposits.

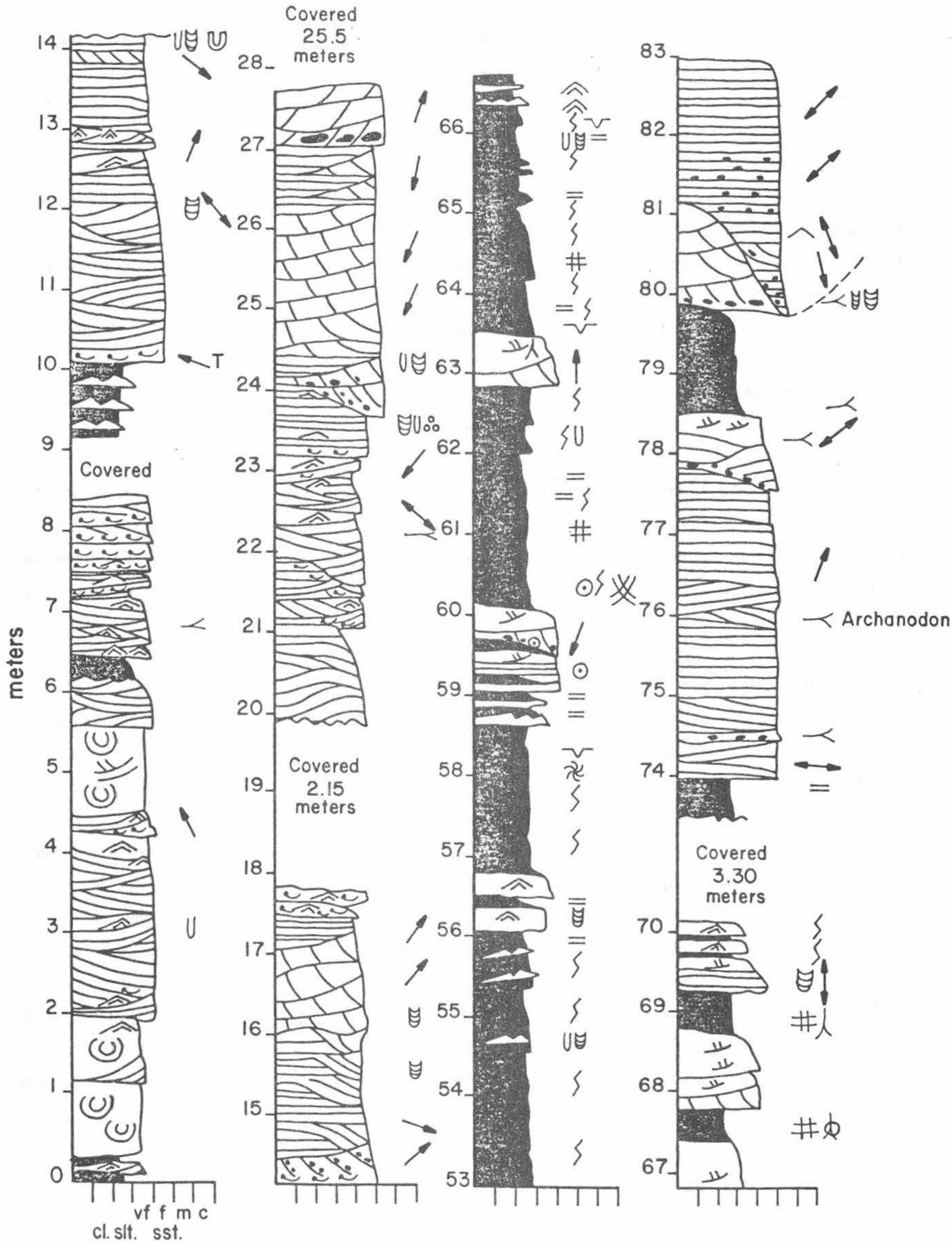
Sandstone beds in the upper 12 meters of section (Fig. 8) were deposited mainly by N-NW directed unidirectional currents moving sand as dunes. Wave-ripples in upper parts of many sandstone beds indicate deposition in standing water with weak wave currents. Stronger wave currents are indicated by swaley cross-strata near the base of some sandstone beds. These characteristics, plus the abundance of burrowing, lack of roots (except at the top of the section), and absence of marine fauna, suggests deposition by fluvial sheet-floods in a coastal body of standing water. Despite rare southeasterly paleocurrents, there is little evidence to suggest tidal currents. Johnson and Friedman (1969) assigned these rocks to the boundary between tidal and alluvial facies.

Stop 5: Grand Gorge Route 30

Description

The lowermost 28 meters of this section (Fig. 9) is closely comparable to the upper 30 meters or so of Stevens Mountain Quarry. In particular, sandstone bodies increase in thickness upwards and contain more angle-of-repose cross-strata relative to swaley cross-strata and planar-strata. There are also lateral transitions within the lower parts of the upper two sandstone bodies between angle-of-repose cross-strata, swaley cross-strata and planar-strata. Large-scale cross-strata in the upper parts of sandstone bodies indicate paleocurrents to the NNE and NE, whereas those lower down indicate S to SSW paleocurrents. However, solitary sets of planar cross-strata within the planar/swaley parts of these sandstone bodies indicated paleocurrents ranging from NNE to SE. Oscillatory current directions are commonly NW-SE. Large (5-10 centimeter-diameter) vertical meniscate burrows occur commonly in sandstone bodies, along with millimeter-wide *Skolithos*, *Arenicolites* and *Chondrites*. Load casts occur beneath sandstone bodies low in the section. Beds at the base of the section contain marine fossils.

Figure 9. Sedimentological log of Route 30 section, N of Grand Gorge.



Further up section, above a 25-26 meter covered interval, a 17 meter-thick sequence of red mudstones and red-gray very-fine to fine-grained sandstones are exposed (Fig. 9, meters 53 to 70). Mudstones are either fissile, or intensely disrupted and blocky-weathering with slickensided clay-lined surfaces between blocks. Desiccation cracks are common. Horizons of calcareous concretions (rhizocretions) occur rarely, as do pseudoanticlines. Bioturbation is pervasive and includes root casts and vertical burrows. One burrow type is 1-2 centimeters in diameter and has a meniscate fill. Others are 3 to 5 millimeters in diameter. The sandstone beds are sheetlike to lenticular, and range in thickness from centimeters to decimeters. Thinner beds are small-scale cross-stratified and have wave- or current-rippled tops. Thicker beds may be dominantly small-scale cross-stratified but can have large-scale cross-strata or planar-strata immediately above their erosional base. Sparse paleocurrent data indicate flow to the N or SSW. Large vertical, meniscate burrows (centimeters in diameter) occur in these sandstones but bioturbation is dominated by smaller burrows (1 centimeter diameter). Upper parts of sandstone beds may have root casts and rare calcareous concretions. Meter-thick upward fining and coarsening sequences occur in places (eg. Fig. 9, meters 58 to 60).

The upper part of the Grand Gorge road cut contains two erosional-based, fine-to medium-grained sandstone bodies. The lower one is dominated by low-angle cross-strata and planar-strata, but also contains angle of repose cross-strata. Along erosion surfaces occur drifted plant remains and a rare occurrence of the nonmarine bivalve *Archanodon* (internal mold, transported). The upper sandstone body shows lateral-accretion bedding passing laterally into a channel fill. Paleocurrents lower in the sandstone body are to the south. Vertical burrows of varying diameter occur in these lower parts.

Interpretation

The lower succession exposed along route 30 is comparable to the top of Stevens Mountain quarry, and is interpreted similarly. Red beds in the middle of the Grand Gorge Route 30 roadcut indicate a fluvial overbank sequence (see also Miller and Woodrow, in press). Desiccated, bioturbated mudstones represent floodbasin deposits. Blocky weathering mudstones with local calcareous concretions and pseudoanticlines indicate repeated wetting and drying, illuviation, and formation of ped structure typical of calcareous vertisols. Fissile mudstones indicate relatively minor influence of bioturbation and repeated exposure, and may represent relatively high deposition rates in perennially ponded areas. Wave-rippled sandstone lenses indicate periodic wave activity on such ponded areas. Sandstone beds with unidirectional sedimentary structures indicate periodic sand deposition during floods, either as individual sheet flood deposits within the floodbasin, or associated with progradation of levees or crevasse splays. The coarsening-upward sequence at meters 58 to 60 (Fig. 9) suggests progradation of a levee or crevasse splay into a floodbasin. Calcareous concretions and pseudoanticlines in overlying beds indicated subsequent rapid abandonment, reduction in deposition rate, and soil formation. The overall vertical sequence at Grand Gorge, Route 30 therefore is generally regressive, representing an upward transition from nearshore marine through coastal and fluvial deposits.

DISCUSSION OF DEPOSITIONAL ENVIRONMENTS

The Givetian rocks of Schoharie Valley show evidence of: 1) tide-influenced channels (estuaries of deltaic distributaries) with mouth bars and associated shallow-marine shoals; 2) muddy and sandy tidal flats and bays; and 3) river channels, floodplains and lakes (see also Johnson and Friedman, 1969; Miller and Woodrow, in press). The marine shelf was clearly storm-wave dominated. There is no evidence for beaches, intertidal sand-mud channel deposits with lateral-accretion bedding, nor well-developed tidal bundle sequences in cross-stratified sands. Therefore, the North Sea tidal flats may not be particularly good analogues for these deposits, as implied by previous workers (eg. McCave, 1968; Johnson and Friedman, 1969). Strongly-asymmetrical, ebb-dominated currents, possibly due to dominance of river flow during deposition, may explain the rarity of tidal-bundle sequences. Reactivation surfaces in cross-stratified sandstones generally indicate a nondepositional or erosional subordinate tide (c.f. Dalrymple, 1984; DeMowbray and Visser, 1984). Tidal range is difficult to estimate because tide-formed sedimentary features are not precise indicators of tidal range, and because tidal ranges and currents may have been enhanced by 'storm tides' or river floods during deposition. Furthermore, different parts of the coast may experience different tidal ranges, as is clearly evident in the North Sea and the Atlantic coast of the USA. This lack of beaches argues against a microtidal range, and many of the tide-formed sedimentary features in Schoharie Valley indicate higher tidal ranges (possibly mesotidal; see Slingerland, 1986). The lack of beaches along a coastline that clearly experienced strong storm waves may be the result of high rates of deposition of sand and mud near the mouths of a complex of channels. Switching (avulsion) of distributary channels was clearly an important process and controlled local deposition rates at the shoreline. It is recorded by the vertical alternations of channel sandstone bodies and the sandstone/mudstone successions. This implies the existence of a tide-influenced delta.

Storm-wave dominated marine shelf environments with submarine sand bars have also been reconstructed in the slightly younger Frasnian part of the 'Catskill clastic wedge' in New York and Pennsylvania (Craft and Bridge, 1987; Halperin and Bridge, 1988; Slingerland and Loule, 1988). Although wave-ripple crest orientations are variable, modal oscillation directions appear to be WNW-ESE (normal to the paleoshoreline) and NNE-SSW (alongshore), as observed in the Schoharie Valley. In Pennsylvania, Slingerland and Loule (1988) believe there is longshore drift to the SW. These workers all agree that tidal currents were only important depositional agents at the coast, but reconstructions of the nature of the paleocoastline differ.

It is generally agreed that beaches were not important in the Catskill clastic wedge (eg. Walker and Harms, 1971; Woodrow, 1985; Halperin and Bridge, 1988; Slingerland and Loule, 1988). Frasnian coastal deposits in central New York show evidence of distributary channels, mouth bars and interdistributary bays with little evidence of strong tidal currents (Sutton et al, 1970; Halperin and Bridge, 1988), but also of estuarine? channel bars and tidal flats with evidence for strong tidal currents (Bridge and Droser, 1985). Slingerland and Loule's (1988) reconstruction of the Frasnian shoreline in Pennsylvania is one of estuaries, tidal flats and shoals. Tidal ranges were considered to be high mesotidal, with strong flood domination. They saw no evidence for channel mouth bars, nor did they record tidal-bundle sequences in their cross-stratified estuarine sandstones. In contrast, Walker and Harms (1971) interpreted a Frasnian shoreline in Pennsylvania as being muddy with no shoreface sands, and only a few channel sandstone bodies. They took this to indicate low wave energy and low-tidal range (microtidal).

This diversity of different types of Givetian and Frasnian shoreline deposits should not necessarily be taken to reflect markedly different shoreline physiographies and sedimentary processes in different regions at different times. The common threads linking all of these deposits are: the presence of sandy channels with varying degrees tidal influence; the presence of extensive shallow bays and tidal flats where mud and sand were deposited; the rarity of beaches; and the storm-wave domination of the shelf. Much of the variability could be explained within the context of a wave-and tide-influenced deltaic coastline where location of distributary channels, tidal range, and wave activity varied in space and time.

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Apendix 1: Road log for Schoharie Valley field trip.

Start in Oneonta. Travel underneath Interstate 88 at exit 15 and turn left onto NY route 23, heading east. Approximantly 3 miles past Grand Gorge, turn left down a dirt road towards Hardenburgh Falls. Cross the bridge over Bear kill and park. Walk east down track on north side of the creek to exposures adjacent to the bridge.

Stop 1: Hardenburgh Falls

Return to NY route 23 and turn left (east) towards Prattsville. Travel for approximately 1.5 miles until the bridge over Schoharie Creek. Turn sharply left (north) immediatly after crossing the bridge along the road that follows the east side of Schoharie reservoir. Travel approximantly 0.6 miles as far as Gilboa- West Conesville Central School. Turn right immediately past the school, travel to the back of the school and park near the playing fields. Walk Southeast to the far corner of the playing fields and find the trail leading up the hillside to the quarry (about a 10 to 15 minite walk).

Stop 2: Stevens Mountain Quarry

Return to the main road by the school and travel south for approximately one mile to the bridge over Manorkill. Park on the south side of the bridge and find the trail leading down to the reservoir.

Stop 3: Manorkill Falls

Travel north approximately 1.5 miles (past the school again) and cross Schoharie Creek at Gilboa. Turn right immediately, along the left bank of Schoharie Creek, and travel approximately .5 miles. The outcrops to be viewed are rock ledges along the opposite side of Schoharie Creek.

Stop 4: Schoharie Creek

Return to Gilboa and turn right (do not cross the bridge again). Several Devonian tree trunk casts are displayed in a fenced area by this turn. This road joins NY route 30 after 1 mile. Turn left (south) on NY route 30 and travel approximately 1 mile until the road starts to climb up the west side of Pine Mountain. Park at the base of the hill near the first outcrops. Be careful of traffic by these road cuts.

Stop 5: Route 30, north of Grand Gorge

Continue on NY route 30 to Grand Gorge, and turn right (west) onto NY route 23, heading back to Oneonta.

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

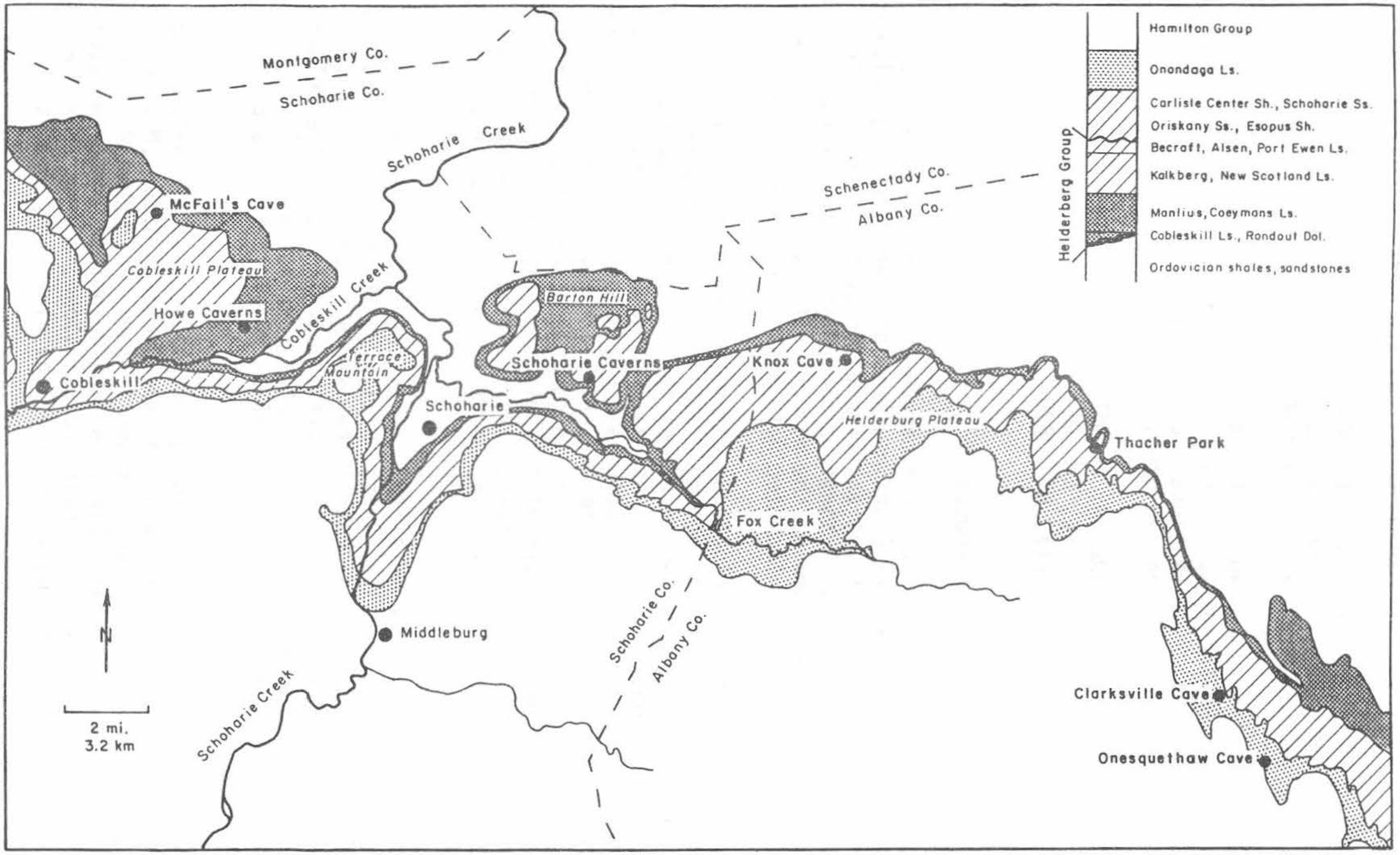

Figure 1: Geologic map of the field-trip area, showing location of stops (from Palmer, et al., 1991).

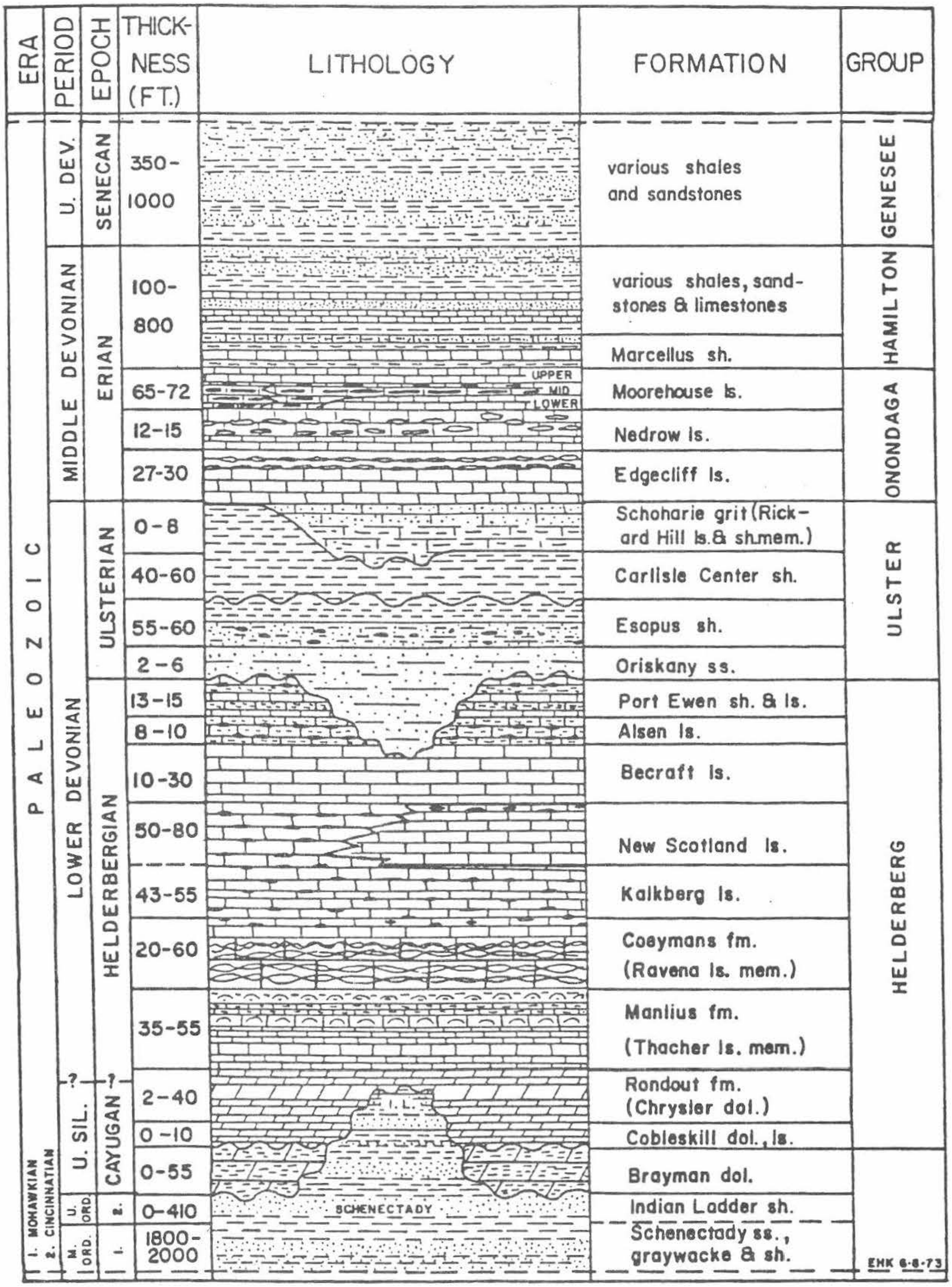

Figure 2: Stratigraphic section in the field-trip area (Kastning, 1975).

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.

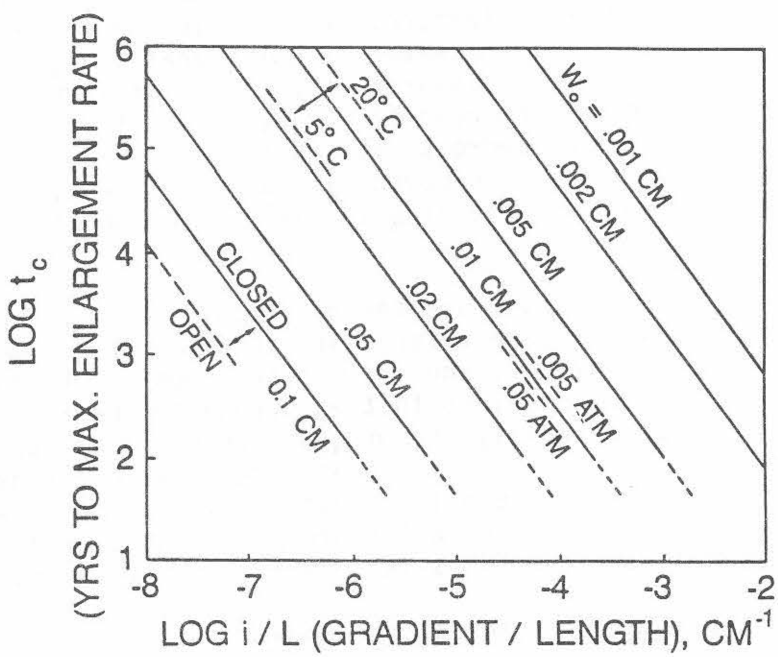

Figure 3: Time (t_{max}) required for a fissure in limestone to reach its maximum rate of enlargement (Palmer, 1991). L = length of fissure (cm), i = hydraulic gradient (dimensionless), W_0 = initial fissure width. The effects of difference in temperature, carbon dioxide partial pressure, and open vs.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

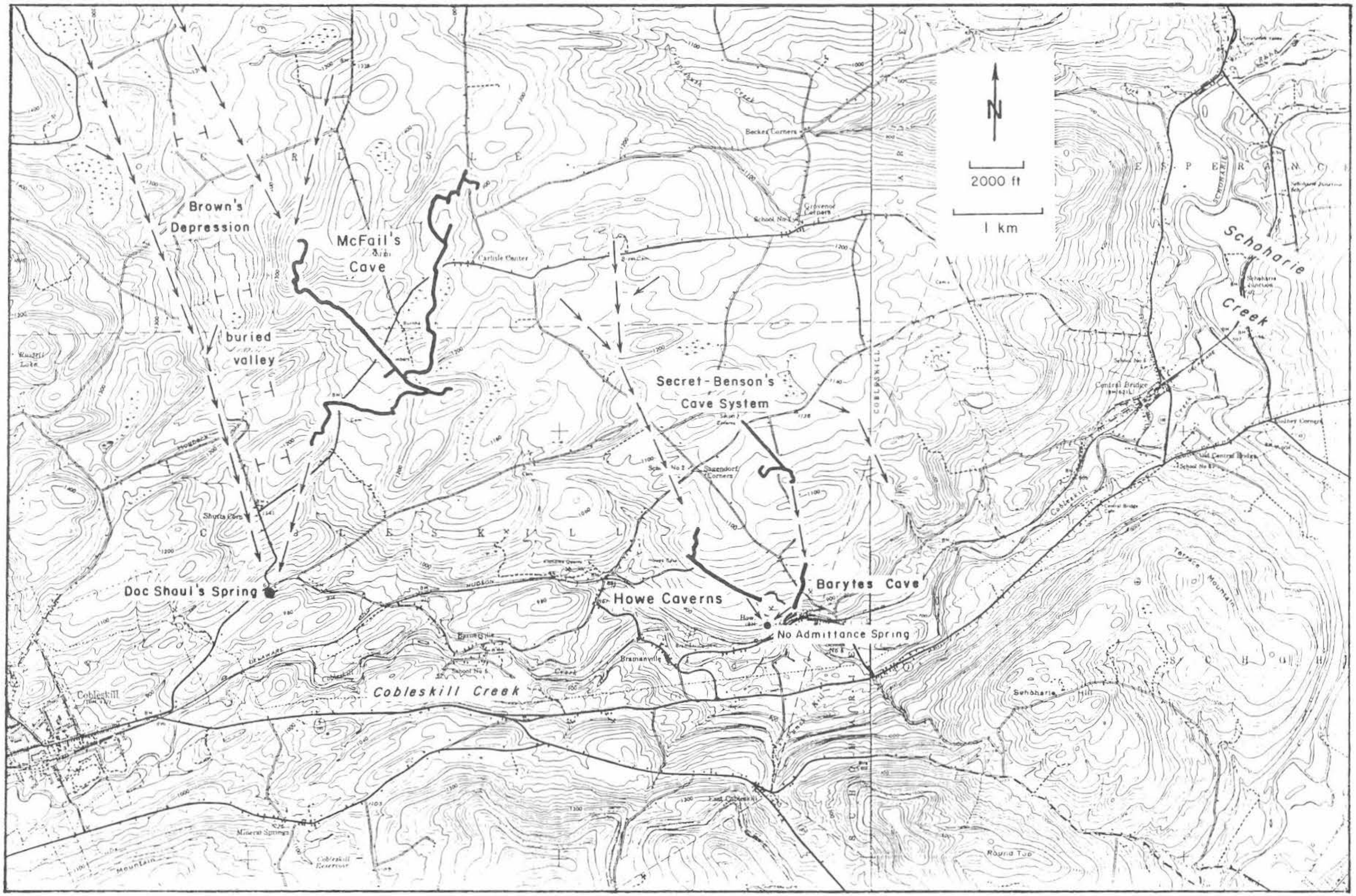

Figure 4: Topographic map of the Cobleskill Plateau. Arrows = dye traces (Mylroie, 1977). Dark lines = caves.

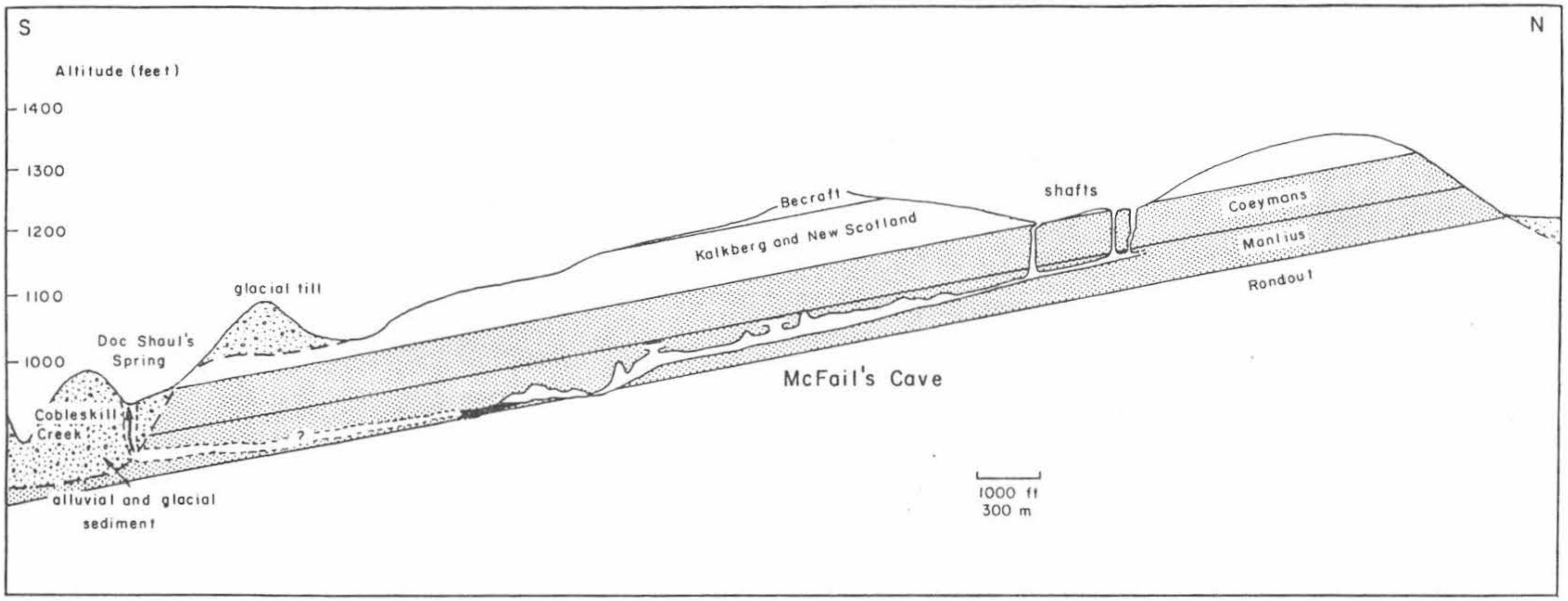

Figure 5: Geologic profile through the Cobleskill Plateau, showing the relationship of the Cobleskill valley and glacial deposits to the pattern of underground drainage.

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^{o} E and S 70^{o} 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).

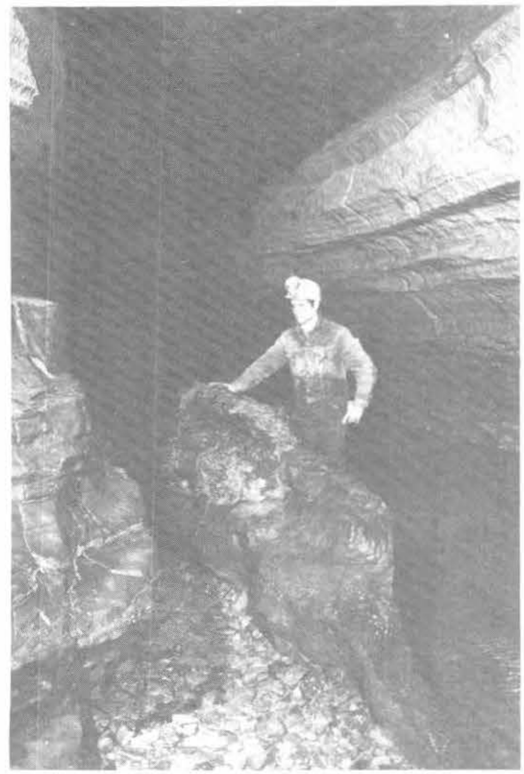

Figure 7: Down-dip vadose canyon passage in McFail's Cave. Rock pedestals in this area show evidence for two stages of down-cutting interrupted by a sediment fill stage. Remnants of cobble fill and glacial lake clay are visible in the foreground.

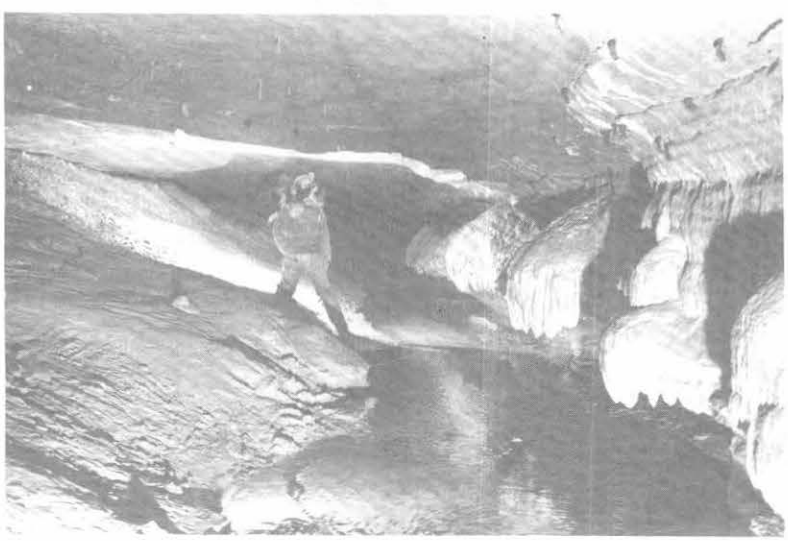

Figure 8: Strike-oriented tubular passage in McFail's Cave. Travertine has accumulated where water seeps into the passage along a gently dipping thrust fault.

History, Economy, and Geology of the Bluestone Industry in New York State

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Albany, NY 12230

ABSTRACT

Upper Devonian sandstone units that occur in southeastern New York, known as *bluestone*, were a source of dimension stone for the construction industry of southern New York and eastern Pennsylvania long before their descriptions were reported in the geological literature in 1886. Various parts of the Sonyea and West Falls Groups have been quarried. However, the economic significance far outweighed the geological importance of the units, so precise relationships between the minable stone and its geologic unit or environment of deposition were and are largely unknown. The extraction of bluestone, like the exploitation of other low value per volume commodities, has been driven principally by market demand rather than by geologic availability. Descriptions of the properties of the stone are primarily cosmetic rather than petrographic.

The purpose of this trip is to demonstrate the association between the geologic and mining aspects of bluestone quarrying. It will explore the connection between the geology and the mines and examine historic and modern quarrying operations. Participants will learn of the technologies used to locate, identify, and extract currently recognized bluestone resources. In addition, changes in the style of mining and in parameters of the product that could be economically recovered throughout the history of the industry will be illustrated.

INTRODUCTION

Bluestone has been historically quarried in the Hudson Valley, central and southern New York, and northern Pennsylvania and continues to be quarried in some of these areas. Bluestone, or flagstone, is the trade, or commercial name for this type of dimension stone. The rock can be generally classified as a well-cemented, angular, medium- to fine-grained sandstone with its most outstanding characteristic being the bedding. Even-bedded rock that can be split in parallel layers is marketable as bluestone, regardless of its grain size, texture, color, or composition. In general, this bedding characteristic and its resistance to wear and weathering, the result of being composed primarily of quartz, give the stone its economic value. The "blue" of the bluestone name came from the color of the stone quarried in Ulster County during the 19th century, and is no longer an accurate description. The color of bluestone varies from the more common blue and gray colorations to include green

through red-purple varieties. The stone is used for sidewalks, veneer, stair treads, curbing, and in other places where durability and a non-skid surface are required.

STRATIGRAPHY

The rocks that compose the bluestone resource are generally of Upper Devonian age. The bluestone quarries in the Deposit-Hancock region are generally developed in rocks of the Upper Walton and Slide Mountain Formations. These are middle Late Devonian units of Frasnian age assigned to the West Falls Group (Rickard, 1975). The Walton Formation is the Catskill-facies equivalent of the Rhinestreet Formation. Whereas the Rhinestreet Formation was deposited in a marine basin environment, the Walton Formation consists of non-marine sandstone and shale. The Walton Formation is approximately 1900 feet thick and contains greenish gray sandstone, green shale and red sandstone and shale. The red beds are relatively rare, constituting about 12% of the stratigraphic section (Sutton, 1963). The Rhinestreet Formation is divided into several dark or black shale members in south-central New York. These units, traced eastward, interfinger with tongues of Catskill facies rocks of the upper Walton Formation. This marine-non-marine change occurs in the area of the Deposit-Nineveh Quadrangles. Sutton (1963) suggested that several dark gray shale units in the upper Walton can be correlated with the black shale units of the Rhinestreet Formation to the west. The Slide Mountain Formation a coarse, non-marine green sandstone, is the equivalent of the uppermost members of the Rhinestreet, Gardeau, and Nunda Formations to the west. In New York, the Upper Walton Formation of the West Falls Group is the predominant source of bluestone in this area. This group corresponds to the New Milford Formation from which the majority of northern Pennsylvanian stone is quarried (Krajewski and Williams, 1971).

PETROLOGY and DEPOSITIONAL ENVIRONMENT

In addition to its characteristic parallel bedding planes the petrology of bluestone contributes to its utility as a building stone. In the most general case, bluestone can be described as mineralogically stable, having a great resistance to ordinary weathering, and being almost as durable as quartzite (Wadeson, 1986). Historically the composition of bluestone, as described in Dickinson (1903), is of angular quartz grains, two feldspars, one decomposed and the other very fresh, cemented with an acid resistant material, probably silica.

Lithologically, bluestone is a non-marine sub-graywacke according to the classification system of Pettijohn (1957) and a low-rank graywacke in the usage of Krynine (1948). Figures 1a and 1b show the composition of bluestones.

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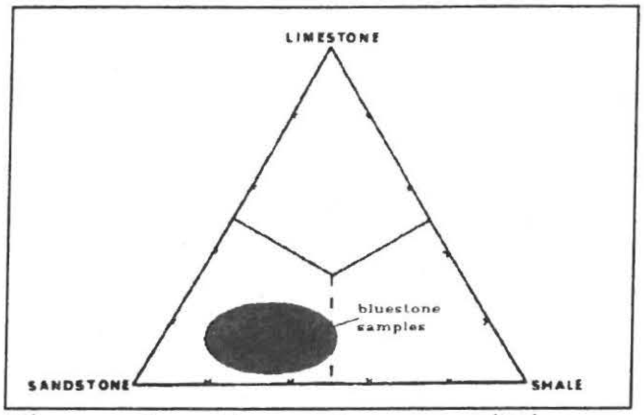

Figure 1a. General composition of bluestones.

The physical properties of Pennsylvania bluestone are very likely to be similar to those of the rocks quarried in the Deposit-Hancock area. The specific gravity of the Pennsylvania stone is 2.48. Percent H_2O absorption (porosity) is 5.3%. Porosity in other commercial sandstone is 2-15%. Permeability is 48.9 millidarcies, considered to be in the "good" permeability range. Weight loss through 16 sodium sulfate test cycles was 13.8% (Krajewski & Williams, 1971). The durability of the stone is better than that of other commercial sandstone, such as the Crab Apple Sandstone quarried in Tennessee.

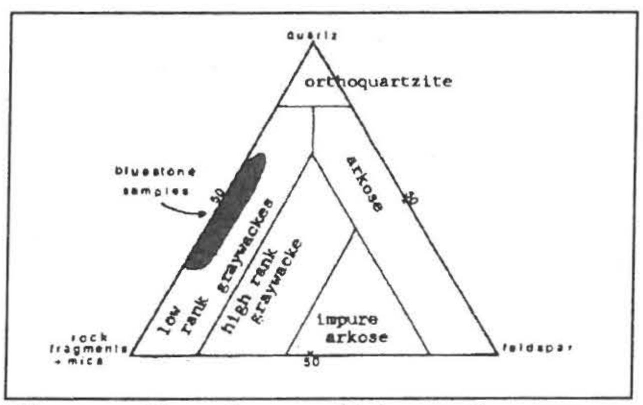

Figure 1b. Composition of bluestone from northeastern Pennsylvania (Krajewski & Williams, 1971).

The clastics of the Catskill delta complex were derived from both sedimentary and low-grade metamorphic terrain to the east and southeast of the delta complex. These sediments, including those that comprise the bluestone resource, were deposited in a shoreline and non-marine coastal alluvial plain environment characterized by lagoons, tidal flats, and low gradient

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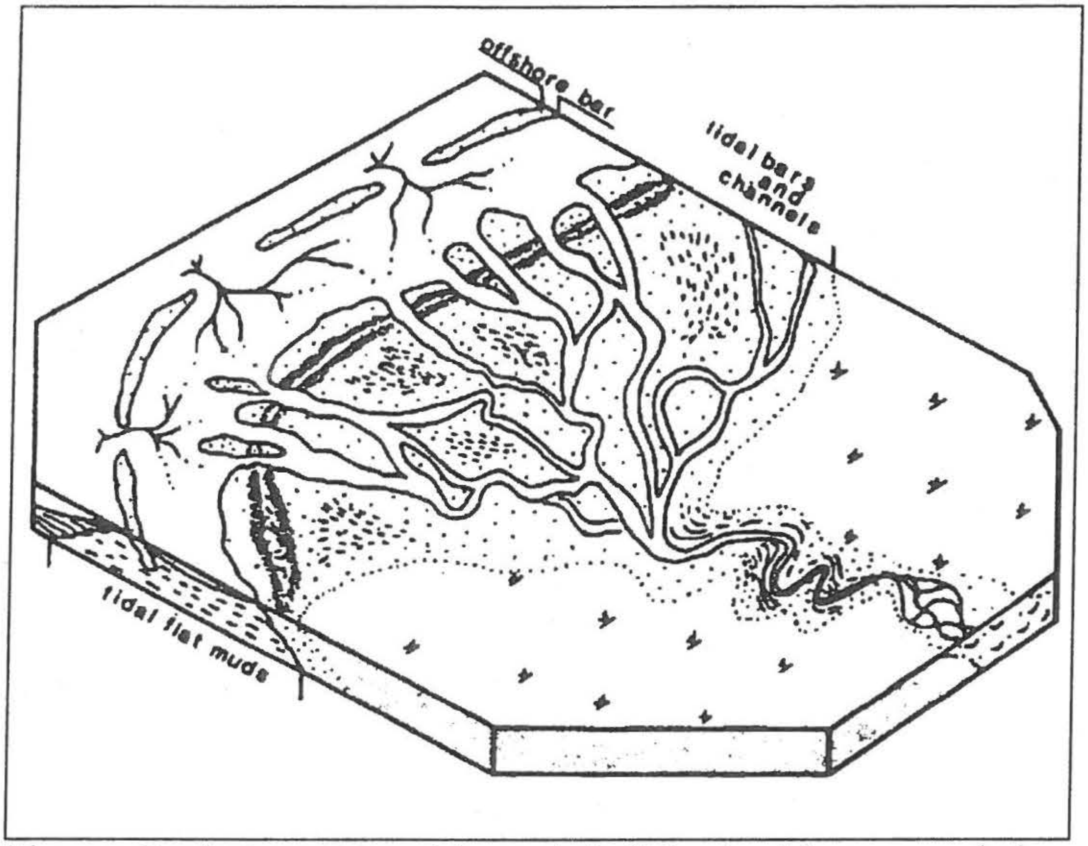

Figure 2. Paleogeographic model illustrating an origin of bluestone deposits (after Krajewski & Williams, 1971).

Descriptions of the paleogeography, sedimentary processes and non-marine facies of the Catskill delta are given by Woodrow (1985) and Sevon (1985).

Krajewski and Williams (1971) conducted an investigation of 875 bluestone quarries in northeastern Pennsylvania the results of which also apply to the of bluestone in New York. Most of the quarries studied, 811, were located in Susquehanna County, with the remainder in Bradford, Wyoming, Lackawanna, and Wayne Counties. Mining in this region was conducted since the 1880's. Krajewski & Williams attempted to classify the quarries by type of sedimentary depositional environment. They distinguished among beaches, offshore bars and interchannel bars. Geologically, the source of the bluestone in the Pennsylvania region is the New Milford Formation, which varies between 400 and 500 feet thick. Here, as in New York, the bluestone is a dense, compact sandstone deposited in a fluctuating marine and freshwater environment. Individual layers are 1 to 20 feet thick and commonly wedge out laterally. A parting lineation, defined by ridges and grooves on the bedding surfaces establishes the paleocurrent direction. It has been suggested that the alignment of elongate grains that help to define the parting lineation impart a zone of weakness along which the stone can be split (Krajewski & Williams, 1971).

A modern lithological analysis of bluestone indicates it is composed of grains of quartz, rock fragments, feldspar, and other minor components. High-grade metamorphic minerals are lacking. The matrix is composed of detrital and recrystallized clay minerals. The rock is cemented largely by quartz with lesser amounts of calcite. The rocks are mostly fine to very fine sandstone, moderately well sorted, with elongate mineral grains. Modal analyses of bluestone samples are given in Table 1.

Table 1. Mean mineral composition of bluestone from three environments of deposition: 1 = offshore bar, 2 = beach, 3 = interchannel bar (Krajewski & Williams, 1971). Abbreviations: Qtz - quartz, RxF -rock fragments, Mic - micas, Op - opaques, Mtx - matrix, Cc - calcite.

	GRAINS					MATRIX	CEMENT	
	Qtz	RxF	Mic	Fsp	Op	Mtx	Qtz	Cc
Type 1	31.5	22.1	3.8	3.1	2.6	26.7	8.7	1.5
Type 2	38.5	26.5	4.2	5.1	1.2	14.1	9.1	4.2
Type 3	35.9	21.0	2.7	3.4	1.5	23.3	10.3	2.1
Total(n=26)	34.9	22.9	3.8	3.5	1.8	22.1	9.4	2.4

From east to west in their study area, Krajewski & Williams (1971) noted an overall decrease in grain size, increase in matrix content, and decrease in cement. To the west, quarries in offshore-bar environments were the more prevalent type. Eastward, quarries in beach type and interchannel-bar type environments were found to be dominant. This observation is consistent with the paleogeography of the Catskill delta complex.

The color of bluestone is imparted by the matrix materials (Krajewski & Williams, 1971). Red or lilac stone has hematite in the matrix. Green stone has green iron-bearing matrix minerals such as chlorite. The matrix of brown stone contains limonite and organic material, while "blue" bluestone is characterized by unaltered iron minerals in the matrix and an absence of hematite. They reported that the blue stone was recovered deeper in a quarry block while the other colors were found around the edges. Chemical analysis reported by Krajewski & Williams and shown in Table 2 indicates that there is an inverse relationship between ferrous and ferric iron in red and blue bluestone.

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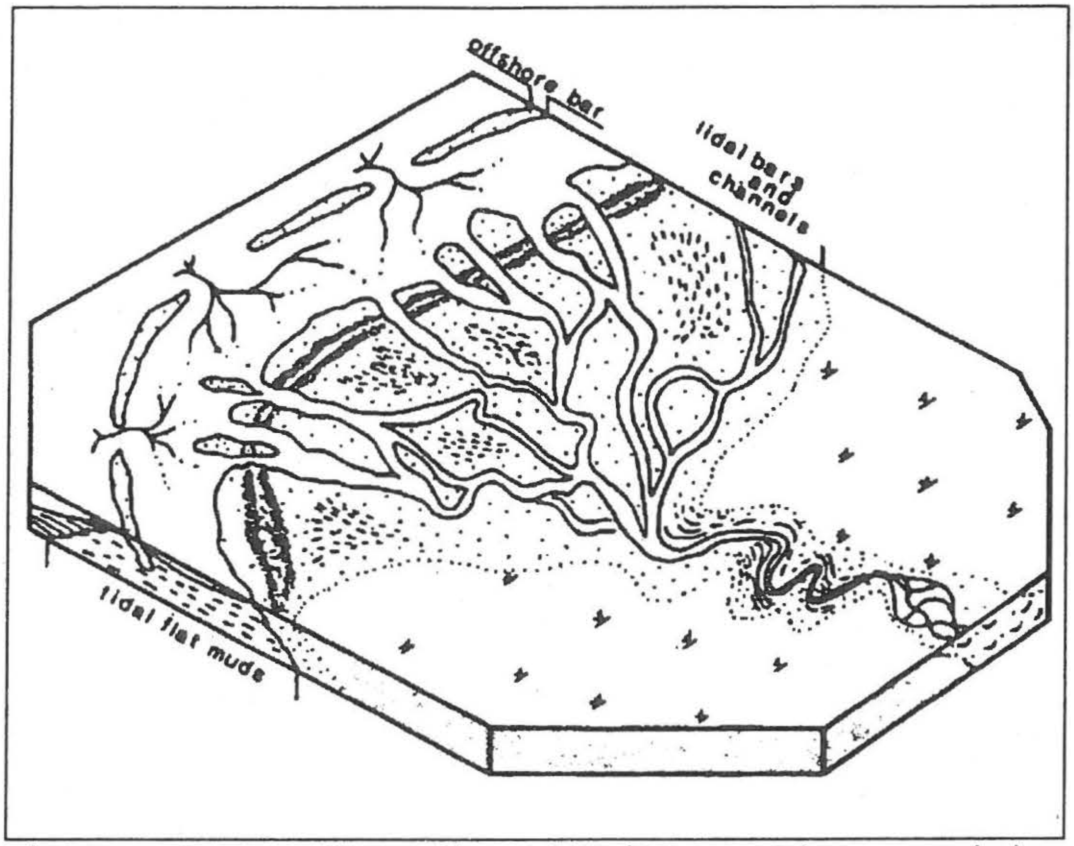

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Table 2. Chemical composition of bluestone of various color (Krajewski & Williams, 1971).

	Red	Brown	Green	Blue
SiO_2	71.50	76.60	69.60	73.70
TiO_2	1.03	1.15	0.81	1.66
Al_2O_3	13.50	11.60	12.50	11.40
Fe_2O_3	3.22	1.82	1.70	0.94
FeO	1.68	1.87	2.99	3.58
MnO	0.06	0.03	0.06	0.07
MgO	1.19	1.02	1.34	1.31
CaO	0.75	0.22	0.45	1.38
Na_2O	1.54	0.57	0.59	0.69
K_2O	2.80	2.24	2.58	1.88
H_2O	2.13	2.17	2.89	2.26
	99.40	99.29	95.81	98.87

HISTORY and QUARRYING TECHNIQUE

The quarrying of bluestone as a dimension stone industry began during the mid 1800's in the area from southern Albany County southward through Greene, Ulster, Sullivan, and Delaware Counties, and into Broome county. The stone from the Hudson River counties, Albany, Greene, and Ulster, was shipped by barge to New York City, while that of the other counties was shipped by rail to Philadelphia and inland locations.

Quarries were (and are) originally located by finding outcrops of sandstone that appeared to be suitable for the production of bluestone. This process of prospecting for bluestone is usually random. Productive quarries have been found by hunters walking through the woods on a hunting expedition. The evaluation of a prospective quarry is based on the experience of the quarryman, the location of existing local quarries and luck (Wadeson, 1986). Figure 3 shows the locations of operating and inactive bluestone quarries of southeastern New York, centered in the Delaware County area of this field trip, as determined by airphoto interpretation.

The major expense of opening a quarry was in the past and is now the stripping of overburden to expose the rock. The overburden consists of soil, glacial deposits, and rock that may not be suitable for stone. This material is dug and blasted away and usually disposed of close to the potential quarry. In many modern quarries, the stripping procedure is one of removing the waste material from a previous quarrying operation when sites are reoccupied in response to demand for particular types of stone changes.

Once the overburden and other materials are cleared, the quarrying of the stone begins. The stone occurs in blocks usually bounded by two sets of vertical joints that trend generally

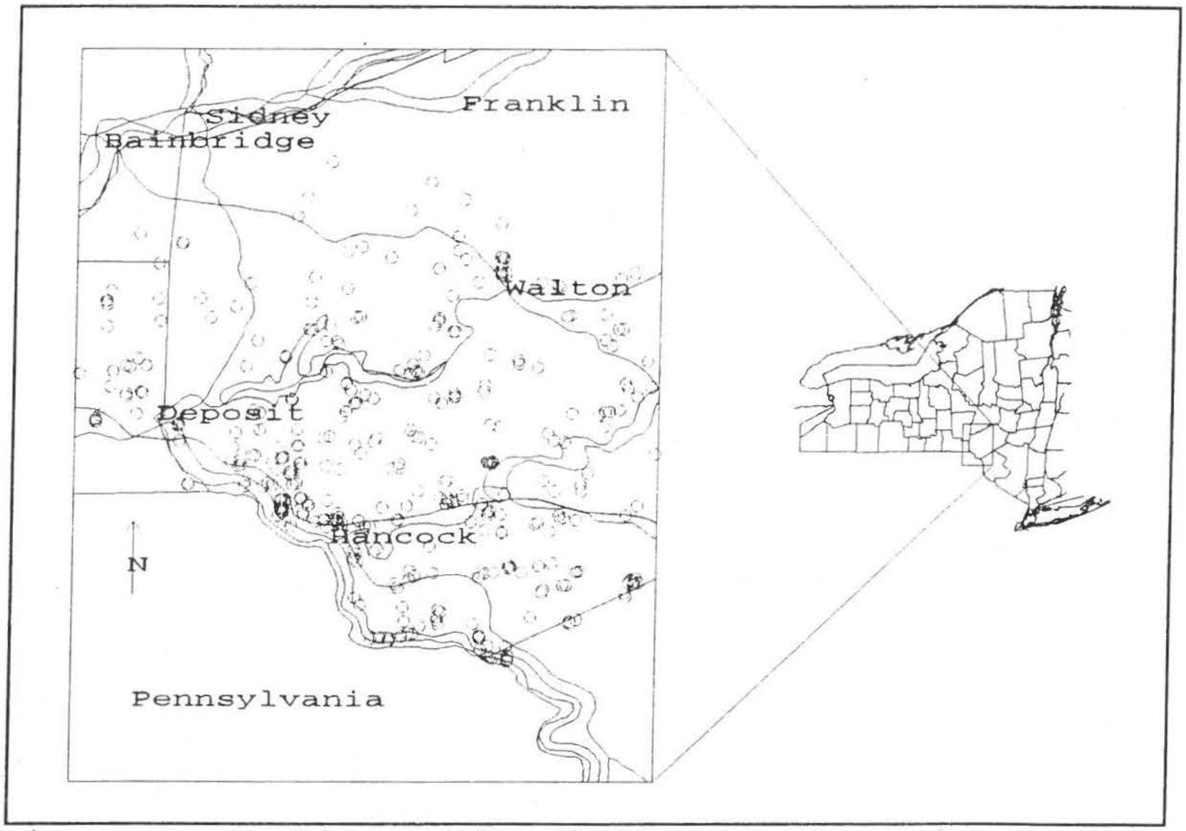

Figure 3. Operating and inactive bluestone quarries located by airphoto interpretation in and around Delaware County.

north-south and east-west. The stone is split horizontally within each block. Depending on the character of the stone, the thickness of each split layer, referred to as a lift, ranges from one inch to several feet. Each lift is extended horizontally as far as possible within the block. If the lift cannot be followed to the edge of the block, the rock must be cut. Bluestone can be cut by scoring a line on the surface and then snapping it, much the same as glass is cut. Hand drilling and plug-and-feather methods for cutting the stone have given way to diamond blade saws, but the basic method is the same. In this way, stone of saleable dimensions-length, width, and thickness-is produced.

Bluestone is sold in seven standard classes, based on its dimensions. Pattern or flag stones are those that range in thickness from 1/2" to 3" and has a minimum surface dimension of 1 square foot, increasing in multiples of 36 square inches. Thicker stone, 1" to 4", that is longer than it is wide (10" to 24" by 3' to 8') is sold as tread for stairways. Similar in dimension to tread, but of random lengths, is coping stone, used as a topping for walls, and sill stone, for windows. Large pieces of stone, with more than about 16 square feet of surface, are referred to as slab stone. Thinner stone is cut into randomly sized parts for use as veneer. Bluestone with at least one

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MnO	0.06	0.03	0.06	0.07
MgO	1.19	1.02	1.34	1.31
CaO	0.75	0.22	0.45	1.38
Na_2O	1.54	0.57	0.59	0.69
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The type of edge on the stone is also used in its classification for market. Bluestone can be snapped (producing a smooth, natural edge), pitched (formed by snapping the stone at a angle not perpendicular to its surface), machined, cut by a guillotine-type cutter, or sawn. Sawing of bluestone, a relatively modern method of preparing it, is done with an abrasive circular or wire saw. This produces a very smooth, straight cut, which is not always desirable. Stone cut in this way is often flamed, a processes that uses a gas torch on the wetted surface to spall the sawed surface, thereby simulating a snapped edge.

The final method of classifying bluestone is by its coloration. The most common color classes are; purple, blue-green, gray-green, tan, and red. Color classification is the most subjective feature and often varies from stone to stone from the same quarry.

ECONOMICS

In quarryman's jargon, "rock" is the material that is thrown over the side as waste and "stone" is the material that can be sold. Bluestone, defined on this basis, like other economic commodities, is a material that can be removed from the ground and sold for a profit. The bluestone industry can be subdivided into three major parts, quarrying, distribution, and marketing. Bluestone quarries have historically been two or three person seasonal operations (Dickinson, 1903), and this tradition generally continues. There are interdependent reasons for this tradition. The quarries are usually small because they are bound geologically by joint patterns, and the quantity of marketable stone that can be produced from each is therefore limited. This situation results in a low value per quarry making the operation profitable to a limited number of workers. The quarries also tend to be relatively remote, and the topography of the area makes access difficult, especially considering that bluestone is a heavy and fragile product. The distributors, who buy the stone from the quarrymen and ship it to the markets, are generally located near major transportation routes for ease in shipping. Quarry operations are generally independent, but there is a growing number of operator-distributor partnerships. These partnerships have advantages for both the operator and the distributor. The quarryman benefits when the distributor contributes to the high start-up cost involved in opening a new quarry, the distributor in return, is assured of a continuous supply of stone (Wadeson, 1986). The consumer of bluestone is basically the construction industry and the bluestone market suffers the same variability of economic conditions. A study of the market for Pennsylvania bluestone (Mikutowicz and Schenck, 1970a, b) indicated that the demand for the product was related to the degrees of knowledge of applications and qualities of bluestone by the architects, who specify particular building materials. The independence of the operators and the distributors has resulted in a relatively restricted marketing area, a subject being addressed by associations of producers and distributors.

ACKNOWLEDGEMENTS

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NYSGA BLUESTONE ROAD LOG

Leave Oneonta, go west (right) on Rt. I-88 to Sidney, Exit 9 (25 miles).

Go south (left) on Rt. 8, away from Sidney (4.6 miles).

Cross intersection with Rt. 206 in Masonville, continue south on Rt. 8 to Rt. 10 (13.4 miles).

Go north (left) on Rt 10 at Stilesville, just past the river.

Continue 1.2 miles on Rt. 10 and pull off on north (left) side, outcrop on right.

Stop 1.
Thick-bedded (up to 20') sandstone layers interbedded with sandy shale containing sandstone rubble and intraformational breccia. Shale is red and weathers rusty. Good evenly bedded bluestone, some cross-bedding, plant fossils. Well-jointed; the main joint set is roughly N-S and the subsidiary set roughly E-W. Outcrop is 0.5 miles long along the south side of the road. More thinly, parallel-bedded and cross-bedded units can be seen on the east end of outcrop.

Continue on Rt. 10 for 4.9 miles.

Pull off on north (left) side of road into parking area, just west of intersection with Sands Creek Road.

Stop 2.
Rock suitable for bluestone with 1" bedding in 6' thick layers interbedded with shale layers up to a meter in thickness. Channel deposits composed of basal conglomerates, massive and cross-bedded sandstones, grading into parallel bedded layers at the top of each sequence. Some of the shale contains pyrite, weathering to produce the rusty coloration. Toward the east end of outcrop, very thick (30 feet) sandstone, some cross-bedded, overlying a shalely basal conglomerate, and underlying good bluestone layers. Toward the west end, a channel cut into existing cross-beds and overlain by shale overbank deposits.

Continue 0.1 miles on Rt. 10 then turn right (S) on Sands Creek Road.

Go 0.4 miles up Sands Creek Road (Delaware County Rt. 67), turn right into Indian Country Bluestone Company. This is an active, operating quarry. We are here as guests by permission of the company management.

Stop 3. Indian Country Bluestone Company

Prominent cross-bedding is displayed in the quarry. Note: such well-developed cross-beds are of interest only from the sedimentary structure standpoint, they are a disaster for a quarry. We will see a demonstration of bluestone splitting by old methods (by hand) and stone trimming with the guillotine. Please look over the yard to see the variety of products ready for shipping- pattern stone, tread, veneer, and crushed stone produced from the waste bluestone.

Continue south on Sands Creek Road to "T" intersection with Delaware County Rt. 17 (old Rt. 17) in Hancock (9.3 miles).

Turn left on Delaware County Rt. 17, which is also Main Street, following the signs toward Rt. 97.

Go 0.3 miles and turn right onto Front Street (the street sign is on the opposite side of the road across from McDonald's). Front Street is located between a Getty station and a NAPA store.

Follow Front Street to the left and along the railway tracks.

Go straight at stop sign (0.2 miles).

Continue on Front Street another 0.4 miles, passing the Delaware Inn on left. Note the 6', 8', and 10', 4" thick slabs of bluestone, dry-laid more than 100 years ago, that make up the front porch floor of the Inn.

Note bluestone in walls, steps, and sidewalks as we proceed through Hancock.

At the traffic light (0.1 miles), continue east on Delaware County Rt. 17 (Old Rt. 17).

Continue on Old Rt. 17, straight past traffic light, following Rt. 268 under the Rt. 17 overpass (1.6 miles), and stay on Delaware County Rt. 17 when Rt. 268 heads off left towards Walton.

Continue 1.3 miles and pull over before the big outcrop on left (N) side of road.

Stop 4.

Note: This is a dangerous place. Stay off the septum of sandstone on the south side of the road, between old and new

Routes 17. The drop on the other side is as great as the height of this outcrop. Be careful on the road because vehicles move past this blind curve at high speeds. Also, stay back from main outcrop face to avoid danger of rock falls.

Thick sandstone channel with shale overbank deposits. Stratigraphic packages consist of a breccia at the base, cross-bedded layers above, changing to planar layers near the top, typical of fluvial deposition. Some of these packages may represent a single flood deposit. The cross-bedding is at such a large scale that the rock may appear to be massive. Lateral accretion surfaces produced by channel migrations are visible on the upper surface of the outcrop on the south side of the road. Breccia layers contain shale pebbles, plant material, and concretions. Storm beds and features of soft sediment deformation are present.

Continue east 1.8 miles.

Turn left on unnamed road crossing the divided highway (Rt. 17).

Cross bridge (100 yards) and turn right towards Tylers Switch.

Continue east 0.4 miles to Tompkins Bluestone.

Stop 5. Tompkins Bluestone Company

At this stop we will be the guests of Tompkins Bluestone, one of this area's bluestone distributors. Tompkins uses laser-guided computer-controlled 3' and 10' circular saws to cut stone to whatever dimensions the market demands. They are involved in specialty stone cutting and carving as well as the production of stone for the more traditional applications as treads, flooring, and veneer.

Exit from Tompkins and turn right onto Old Rt. 17.

Return to Hancock (6 miles) and take NY 17 west to Deposit (approximately 10 miles).

Exit NY 17 and take NY 8 north toward Sidney.

At Sidney, turn right onto I-88 (east) back to Oneonta.

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STRATIGRAPHY AND DEPOSITIONAL ENVIRONMENTS OF THE LOWER PART OF THE MARCELLUS FORMATION (MIDDLE DEVONIAN) IN EASTERN NEW YORK STATE

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INTRODUCTION

The "layer cake" stratigraphy of the Hamilton Group in western and central New York appears to change into a more complex and confusing sequence towards the east, where strata thicken and are increasingly dominated by coarse-grained siliciclastics. These trends are exemplified by the lower part of the Marcellus Formation. In western and central New York, this part of the Marcellus Formation is characterized by poorly fossiliferous, black shales with limestone interbeds. These limestones vary in type from intervals of barren or fossiliferous carbonate concretions to complex meter-scale packages of skeletal limestone beds that can be easily correlated over most of this region. Lateral facies changes and the dramatically increased thickness make correlation difficult in strata of the lower part of the Marcellus Formation in easternmost exposures. An approximate 13-fold increase in thickness in eastern New York is accompanied by a transition to more medium- to coarse-grained siliciclastics that overlie the initial black shale deposits. However, the distinctive faunas, sedimentary fabrics, and taphonomic indicators present in some of the strata persist through the facies transition. These strata form useful marker beds that provide a more detailed understanding of the evolution of the northern Appalachian Basin during late Eifelian time.

The purpose of this field trip is to examine the stratigraphic, sedimentologic, and taphonomic changes across the limestone-shale/coarse-grained siliciclastic transition in the lower part of the Marcellus Formation. The following report is a result of two separate ongoing studies. The first is a detailed examination of the limestone-rich facies of the Cherry Valley Member and the upper part of the Union Springs Member in western and central New York State (DHG). The second study involves regional correlation of strata in the lower part of the Marcellus Formation in eastern New

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York State (CAVS). This project includes correlation of the limestone-rich facies of the Cherry Valley Member with the equivalent terrigenous sand-rich facies above the Stony Hollow Member (restricted) in the Hudson Valley.

Initial studies of the Marcellus Formation date back to Hall (1839) and Vanuxem (1840). Clarke (1889, 1901a, b, 1903) was the first to focus attention on the formation and its fauna. The classic work of Cooper (1930, 1933, 1934, 1941) provided the framework on which most modern stratigraphic relations in the Hamilton Group are based. Subsequent stratigraphic studies in the lower part of the Marcellus Formation include reports by Chadwick (1933), Cooper (in Goldring, 1943), Goldring (1935, 1943), Rickard (1952, 1984, 1989), Baird and Brett (1986), and Brett and Kloc (in Anderson et al., 1988).

Post-Clarke paleontologic studies of the lower part of the Marcellus Formation chiefly focused on the limestones of the Union Springs and Cherry Valley Members (Flower, 1936, 1943; Miller, 1938; Cottrell, 1972). Paleontologic studies of the black shale facies include those by Goldring (1935, 1943), Brower et al., (1978), and Brower and Nye, (1991).

GEOLOGIC SETTING

Tectonics and Paleogeography

The Hamilton Group (Middle Devonian) of New York State consists mainly of marine basin and basin-margin siliciclastics and minor, widespread carbonates that form an eastward-thickening clastic wedge (Brett, 1986). These siliciclastics represent the first major phase of Acadian orogenic sedimentation in the northern Appalachian Basin (Faill, 1985; Ettensohn, 1985a).

Deposition of the sediments that now form the Hamilton Group followed the lithospheric downwarping associated with tectonic loading (Ettensohn, 1985b; Beaumont, 1988) and a eustatic sea level rise (TR cycle Id of Johnson et al., 1985). These events combined to terminate shallow shelf carbonate deposition of the Onondaga Formation. Deposition of the Marcellus Formation began in the proximal foreland basin and prograded westward onto the more stable craton (Ettensohn, 1985a). The basin was most likely affected by thermohaline density stratification, and this contributed to the decreased oxygen content of the bottom waters (Ettensohn, 1985b; Woodrow, 1985). Intervals of current-generated sedimentary structures, bioturbation, and shell beds that contain benthic fossils all suggest episodic fluctuations in sea floor oxygenation and reworking by bottom currents, possibly storm currents (Baird and Brett, 1986; Brett et al., 1991; Griffing, 1991).

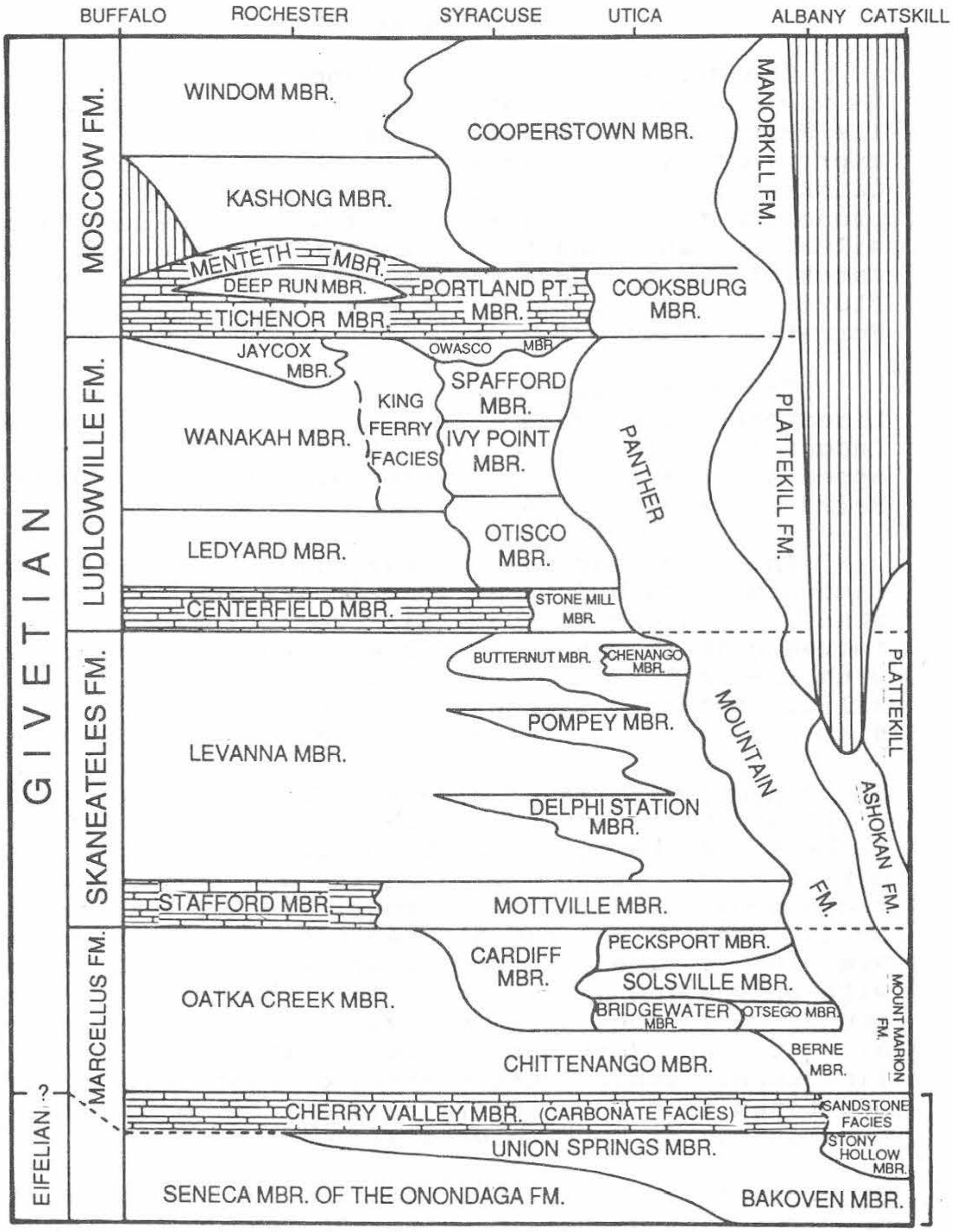


Figure 1.-- Stratigraphic correlation chart for the Hamilton Group in New York State. Bracket to the right of the diagram indicates the units to be discussed in this report. Block pattern highlights geographically widespread limestone members. Based on Rickard (1975), as modified by: Baird (1979), Baird and Brett (1986), Grasso (1986), and Ver Straeten (this report).

Stratigraphic Overview

The Marcellus Formation (upper Eifelian-lower Givetian) is the lowermost formation in the Hamilton Group (Figure 1). The Marcellus Formation ranges in thickness from approximately 7.5 m in western New York to 579.0 m along the Catskill Front (Rickard, 1989). The formation consists predominantly of marine black and gray shales, terrigenous siltstones and sandstones, and minor carbonates. Along the Catskill Front, these marine strata give way to nonmarine, fluvial siliciclastics of the partially equivalent Ashokan Formation (Figure 1).

The Marcellus Formation directly overlies the Onondaga Limestone. The contact represents a major, regional unconformity (Rickard, 1984; Brett and Baird, 1990), yet it appears gradational and conformable locally (Oliver, 1954; Baird and Brett, 1986).

The lower part of the Marcellus Formation also thickens eastward, from a feather edge in western New York to more than 150 m along the Catskill Front (Figure 2). The rapid increase in thickness across eastern New York is associated with the inclusion of relatively coarser grained siliciclastics. This part of the Marcellus Formation is distinctive and can be divided into several members.

<u>Bakoven/Union Springs Members</u>. The Bakoven (eastern New York) and Union Springs (western and central New York) Members are characterized by black to dark gray, pyritiferous, organic-rich shales and limestones. These rocks generally are poorly fossiliferous with low diversity assemblages of pelagic and benthic organisms (Brower and Nye, 1991). The Bakoven Member is predominantly dark shale with only a few limestones. The ratio of limestone to shale abundance in the Union Springs Member increases westward along the New York outcrop belt. The majority of the limestones within these members consist of dark lime mudstones that form small to large (0.1 m to 1.0 m-wide) concretions or centimeter-scale continuous beds.

<u>Stony Hollow Member</u>. The Stony Hollow Member is a calcareous to dolomitic, fine- to medium-grained siliciclastic unit that characteristically weathers buff to dark gray in color. It commonly forms a prominent ridge with the Cherry Valley Member between the less resistant dark shales of the underlying Bakoven and overlying Berne Members. The Stony Hollow Member ranges from finely laminated shales and siltstones with a low degree of burrow mottling in the lower part of the unit to a bioturbated fine- to medium-grained sandstone near the top. Rare pelagic faunal elements in the lower part of the unit are replaced by benthic assemblages with trilobites,

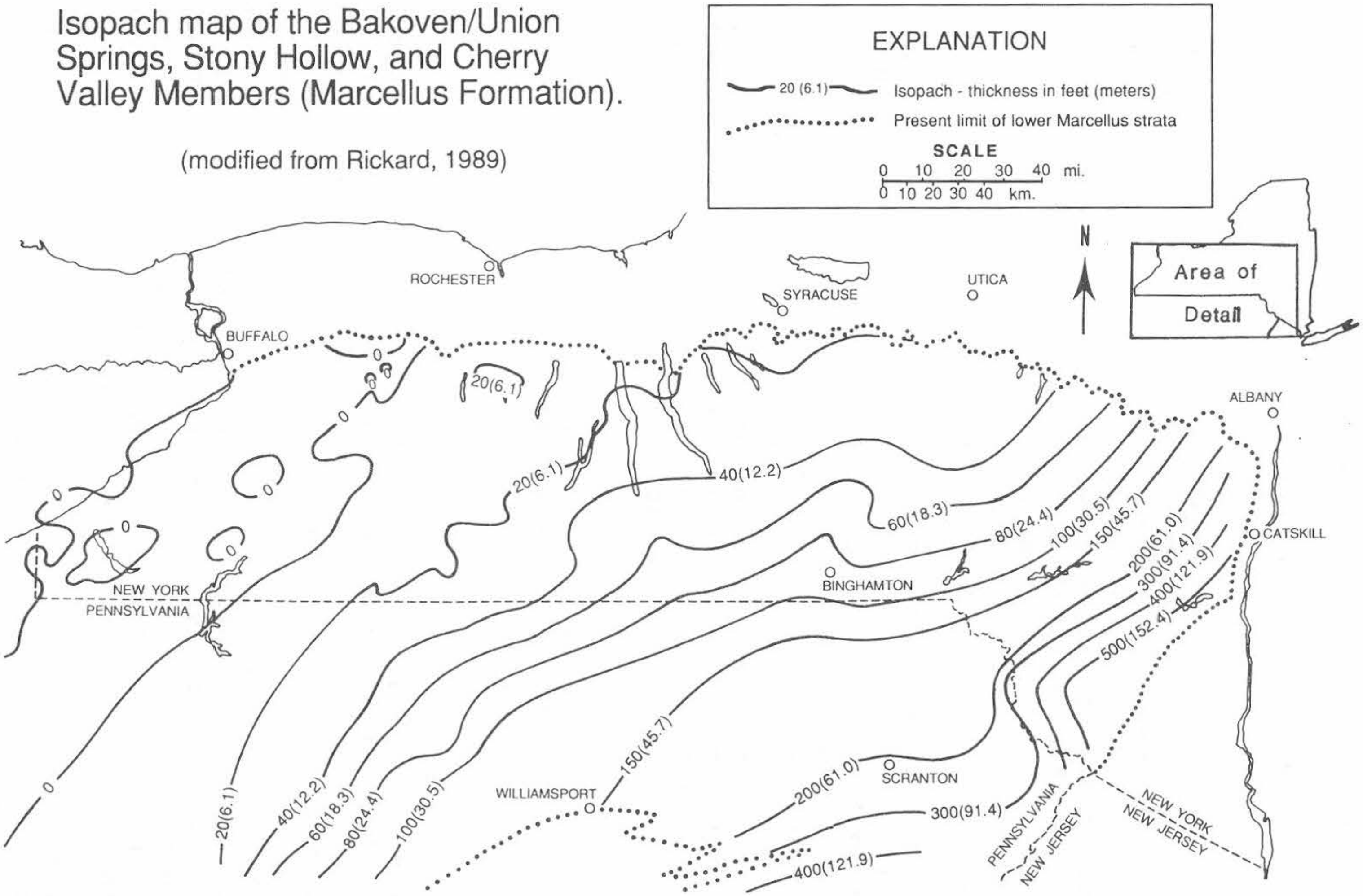


Figure 2.-- Isopach map of the lower part of the Marcellus Formation in New York State and northern Pennsylvania. Modified from Rickard (1989).

brachiopods, crinoids, and small rugose corals near the top. The Stony Hollow is approximately 75 m-thick at Kingston. It thins northward, and at Onesquethaw Creek (southwest of Albany), the Stony Hollow is approximately 6 m-thick. The unit thins west of Onesquethaw Creek and pinches out.

Cherry Valley Member. The Cherry Valley Member forms a package of skeletal limestones and shaly intervals that is recognized from the Genesee Valley (south of Rochester) eastward to Onesquethaw Creek (southwest of Albany). It consists primarily of dark, fetid, argillaceous, massive to nodular bedded styliolinid-cephalopod packstones with a limited benthic skeletal component. The faunas and sedimentary fabrics of the Cherry Valley Member limestones are comparable to the Devonian pelagic/hemipelagic limestones of Europe and north Africa.

Recent studies show that the Cherry Valley Member contains a rapid transition to a more terrigenous sand-rich facies south of Albany along the Catskill Front. The terrigenous sandstones are generally intensely bioturbated and are accompanied by intervening shell hashes, which feature small brachiopods and styliolinids. A southward increase in thickness is associated with the inclusion of thick silty shale intervals in the sandstones. Although uncommon, nautiloid and goniatitic cephalopods also occur in the terrigenous sand-rich facies of the Cherry Valley Member.

THE CHERRY VALLEY MEMBER AND ASSOCIATED LIMESTONES OF WESTERN AND CENTRAL NEW YORK

Introduction

The Cherry Valley Member of the Marcellus Formation was originally defined by Cooper (1930) as one of several relatively thin, widespread limestones of the Hamilton Group in New York State (Figure 1). Detailed study of the limestones of the Hamilton Group reveals that they comprise complex packages of beds, which exhibit a variety of rock types, taphonomic features, and bedding styles (Baird, 1979; Gray, 1984; Grasso, 1986; this report). These limestones represent significant changes in depositional conditions across the northern Appalachian Basin during Middle Devonian time (McCave, 1973; Brett and Baird, 1985).

The westward thinning of the Marcellus Formation across western and central New York leads to the condensation and union of the Cherry Valley Member with limestones in the upper part of the Union Springs Member. This westward amalgamation of limestone beds has hampered the differentiation of characteristic Cherry Valley Member faunal assemblages from those of the Union Springs Member.

In addition, the amalgamation of these limestones has led to erroneous thicknesses previously documented for the Cherry Valley Member. One particular group of Union Springs limestones maintains a close stratigraphic association with the Cherry Valley Member across this outcrop belt. These limestones have previously been included in the Cherry Valley Member at some localities and are distinctly different from other limestones of the Union Springs Member. A more detailed discussion of these beds is provided below.

Chestnut Street Beds

A thin package of centimeter-scale, skeletal-peloidal wackestone and packstone beds underlie the Cherry Valley Member in central New York State. These skeletal limestones feature abundant molts of Proetus haldemani, a trilobite that is rare in other parts of the sequence. The informal name "Chestnut Street Beds" will be used to identify this package of limestones, instead of the name "Proetid Bed" used by Brett and Kloc (in Anderson et al., 1988). The new name is derived from the extensive exposures of these limestones at the Chestnut Street roadcut (STOP 1) located approximately 5.0 km east of the village of Cherry Valley.

The Chestnut Street Beds are the uppermost limestones of the Union Springs Member in central New York (Figure 3). The basal contact of the Chestnut Street Beds is erosional. Dark, argillaceous, lime mudstone concretions are commonly amalgamated to the base of the Chestnut Street Beds. Hypichnial skeletal packstone casts of 1.5 to 2.5 cm-wide Cruziana traces commonly protrude into the underlying shales between these concretions. Where the Chestnut Street Beds directly overlie black shales, the lowermost wackestones or packstones contain small pyritized nodules with 1 to 2 mm-wide, septarian calcite veins. Some veins continue upward beyond the nodule boundaries. In east-central New York, the upper contact of the Chestnut Street Beds is sharp, but conformable, with overlying calcareous silty shales and siltstones. The siltstone beds commonly contain in situ auloporid corals, crinoid ossicles, and the goniatite Agoniatites nodiferous.

Although the Chestnut Street Beds are separated from limestones of the Cherry Valley Member in east-central New York by siltstones, silty gray shales, and black shales, the Chestnut Street Beds are amalgamated to the base of the Cherry Valley Member in west-central New York. The Chestnut Street Beds and the Cherry Valley Member are separated only by a scalloped, pyritized hardground in this area. The Chestnut Street Beds can be distinguished from the directly overlying Cherry Valley limestones by a lighter (white to medium gray) weathering color and a distinct faunal assemblage. Although gastropods, styliolinids, nowakiids, auloporid corals, rhynchonellid brachiopods, and crinoid

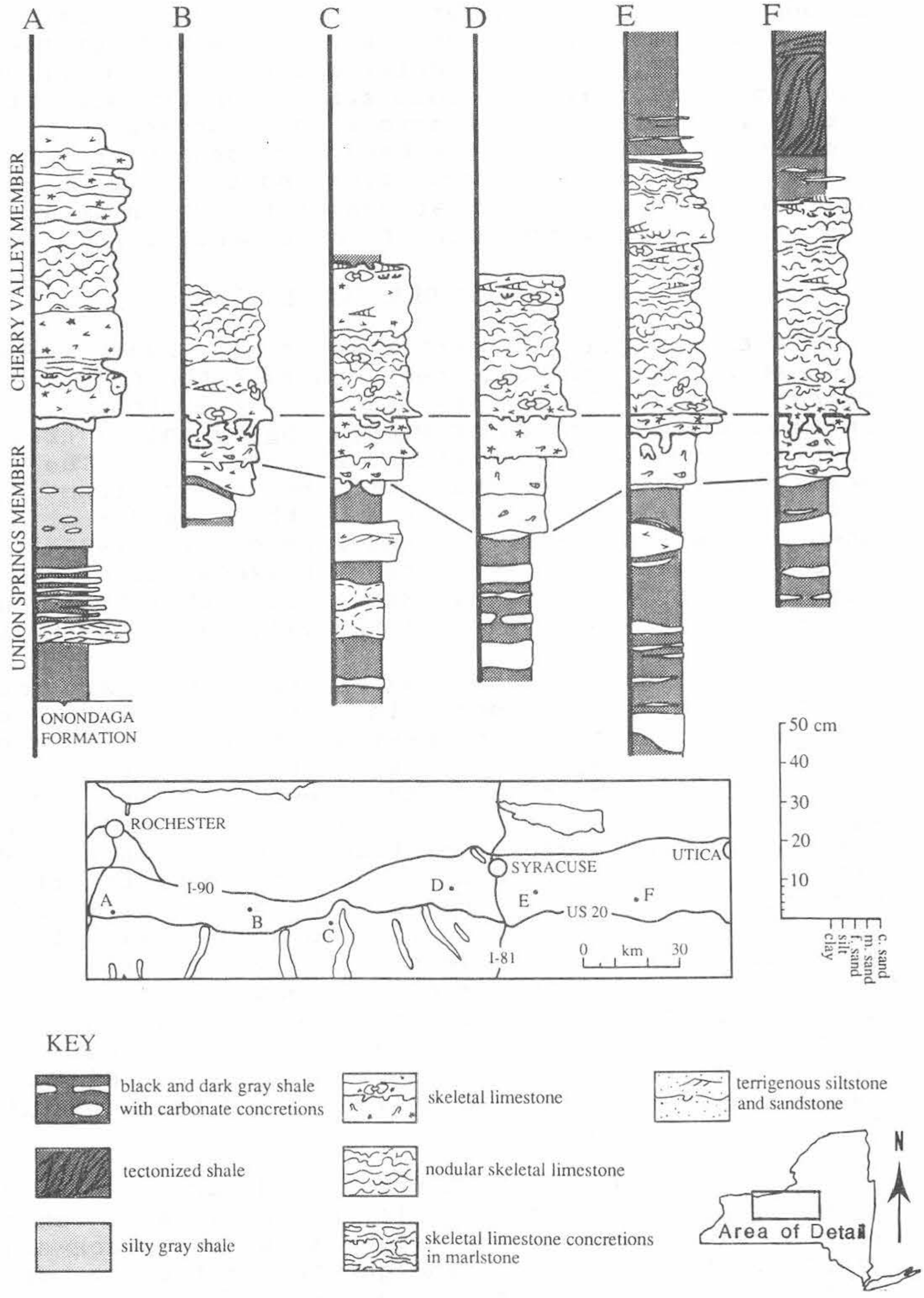
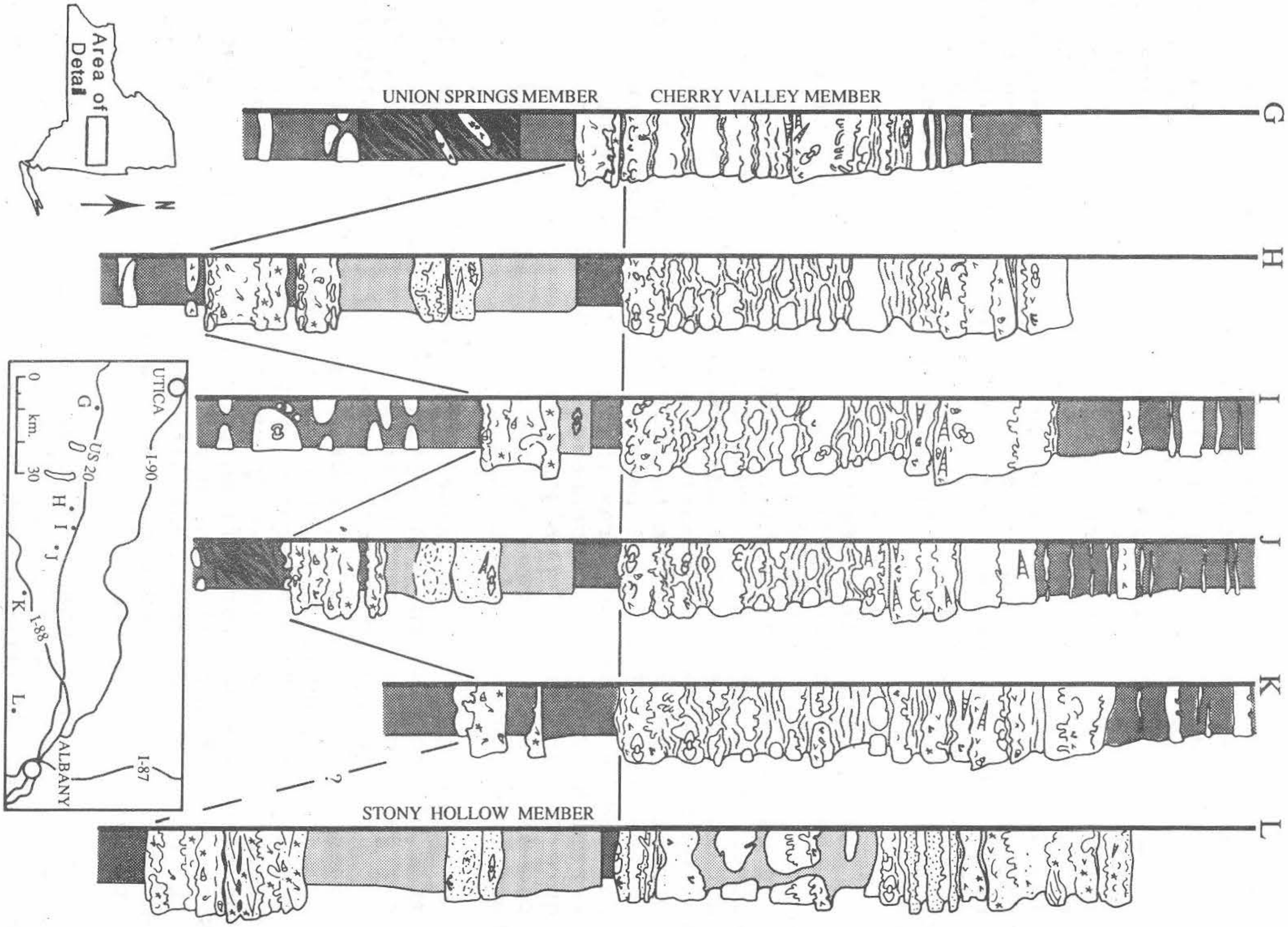


Figure 3.-- Selected stratigraphic sections of the Chestnut Street Beds and the limestone-rich facies of the Cherry Valley Member in New York State. Datum is the base of the Cherry Valley Member. Lower correlation lines mark the base of the Chestnut Street Beds. Locality names for each lettered section are supplied in Appendix I.


UNION SPRINGS MEMBER
CHERRY VALLEY MEMBER
STONY HOLLOW MEMBER
G
H
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US 20
0 km. 30
G
H
I
J
K
L
I-88
I-87
ALBANY
?

debris are common to both limestones, the Chestnut Street Beds lack the characteristic nautiloid and goniatite cephalopod fauna of the Cherry Valley Member. In addition, only the Chestnut Street Beds commonly contain calices of the small crinoid Haplocrinites clio (see Baird and Brett, 1986), atrypid and pentamerid brachiopods, and stereolasmid corals, as well as the proetid trilobites discussed above. The Chestnut Street Beds also contain a distinctive late Eifelian conodont fauna. The conodont Tortodus kockelianus kockelianus (kockelianus zone), previously considered to be a basal Cherry Valley element at Stockbridge Falls (Klapper, 1971), was actually found within the amalgamated Chestnut Street Beds at this locality (G. Klapper, 1991, personal comm.).

The Chestnut Street Beds are commonly united to form one coherent limestone ledge, as at the Chestnut Street roadcut (Section I, Figure 3). There, the beds are bioturbated, and individual bed contacts are diffuse. However, relict fining-upward bedding can be recognized. The bases of these relict beds are defined by the coarse-grained fragmented and disarticulated skeletal fraction. Protrusive feeding and shelter burrows originate at the top of these beds. Individual beds or sets of beds may be separated by calcareous gray shale, as at Cox Ravine (Section H, Figure 3), Rosenberg Road Ravine (Section J, Figure 3), and Mineral Springs (Section K, Figure 3). Hypichnial casts of Cruziana traces are also observed at these limestone-shale contacts.

In addition to the Chestnut Street Beds-Cherry Valley Member contact, sharp, scalloped, commonly pyritized, firmground and hardground surfaces also separate individual Chestnut Street Beds in west-central New York. A relatively greater contrast in color, rock fabric, and skeletal fragmentation/abrasion is observed between individual Chestnut Street Beds separated by these discontinuity surfaces. For example, two Chestnut Street Beds are separated by a scalloped firmground at Seneca Stone Quarry (Section C, Figure 3). The upper of these two beds is a dark, crinoid-rich packstone with heavily abraded skeletal debris. The lower bed consists of a lighter gray bioturbated wackestone.

The number of Chestnut Street Beds decreases westward across central New York. However, changes in the total thickness of the package are not systematic. In fact, the Chestnut Street Beds change in thickness from a maximum of 26 cm to 6 cm across the outcrop (a distance of 20 m) at Rosenberg Road Ravine.

Seven to eight beds of proetid-trilobite-bearing limestone form a 42 cm-thick ledge below the Cherry Valley Member at Long Road Ravine, Albany County (STOP 3). These beds consists primarily of quartz-sand- and crinoid-rich packstones and grainstones. Although bioturbated, these beds exhibit relict small-scale cross-stratification. This package of limestones appears to be completely or partly equivalent to the Chestnut Street Beds farther west (as is discussed in detail later in this report).

Limestones of the Cherry Valley Member

The dark-gray- and brown-weathering, hackly fracturing, argillaceous skeletal limestones of the Cherry Valley Member form a distinctive, meter-scale, resistant package within an interval of relatively non-resistant black shales. Natural exposures, such as the common stepped waterfalls in stream beds, are spotty across New York State. Carbonate dissolution, tufa crusts, and heavy moss overgrowths combine to obscure subtle textural details in these exposures. However, an extensive collection of polished rock samples from these outcrops, along with a limited number of good quarry and roadcut exposures, illustrate many of the details in these strata.

The limestone-rich facies of the Cherry Valley Member can be recognized in exposures that extend from the General Crushed Stone ("5 Points") Quarry, near Lima, Livingston County, to Onesquethaw Creek, Albany County. The limestone package ranges in thickness from a minimum of 0.37 m at Flint Creek, Ontario County, to a maximum of 2.4 m at Onesquethaw Creek. No exposures of the Cherry Valley Member are at present known west of the Genesee Valley south of Rochester. The disappearance of the Cherry Valley and Union Springs Members in western New York is most likely due to erosional truncation of the members prior to deposition of the overlying Oatka Creek Member (see Baird and Brett, 1986). However, early reports of Cherry Valley-like limestones that directly overlie the Onondaga Limestone in salt shafts at Retsof, Livingston County (Luther, 1894), and of Cherry Valley Member faunas within the uppermost Onondaga Limestone at Erie County (Clarke, 1901) suggest that part of the Cherry Valley may reappear farther west below the regional sub-Oatka Creek disconformity (Rickard, 1984).

The carbonate content of the limestone-rich facies of the Cherry Valley Member consists primarily of skeletons of the minute dacryoconarid Styliolina fissurella and neomorphic calcite microspar. The microspar texture obscures the differentiation of primary cements, fine skeletal debris, and peloids. Brower and Nye (1991) estimated the composition of the Cherry Valley Member, near Syracuse, to include 86% calcite, 12% clay, and 1% organic carbon, plus minor pyrite. The clay and organic content of the Cherry Valley Member increases to the east and varies vertically within each section. The styliolinid packstones and (minor) grainstones of the Cherry Valley Member also contain auloporid corals, minute brachiopods, bivalves, gastropods, nowakiids, coleolids, placoderm fish plates, and a famous nautiloid and goniatite cephalopod fauna (see Flower, 1936; Miller, 1938).

The basal contact is sharp across central and western New York, especially where the Cherry Valley Member overlies the amalgamated Chestnut Street Beds or other concretionary limestones of the Union Springs Member (as at the "5 Points" Quarry). Rickard (1952) defined the top of the Cherry Valley Limestone at the contact of the uppermost massive limestone with overlying black shales. This contact is gradational and conformable in east-central New York, where the Cherry Valley underlies a thin interval of relatively resistant, calcareous, fossiliferous black shales and thin dacryoconarid packstones. In contrast, the upper contact in west-central New York is generally a sharp erosional discontinuity with the overlying poorly fossiliferous black shales of the Oatka Creek Member (discussed above).

Rickard (1952) recognized a three-part subdivision within the Cherry Valley Member of east-central New York. This subdivision consists of (1) a lower massive limestone, (2) a middle nodular limestone/shaly interval, and (3) an upper massive limestone. Although this basic three-part subdivision can be recognized across central New York, closer inspection of the Cherry Valley reveals a gradation in textures between the massive and nodular intervals. For example, the continuous packstone bed that forms the base of the Cherry Valley Member at the Chestnut Street roadcut (STOP 1) is as extensively bioturbated and lumpy as the overlying two layers of nested, diffusely-bounded, concretionary packstone bodies (Section I, Figure 3). The highly calcareous, skeletal shales or marlstones that separate these "nodules" both exhibit the continuation of diffuse migrating burrow traces that originate in the packstones and also contain similar skeletal debris. The nested packstone nodules weather to form part of what Rickard (1952) considered the lower massive subdivision, due to the relatively greater calcium carbonate content of the lower marlstones. Marlstones above this level weather more like shale and exhibit differential compaction around the packstone "nodules."

The fabric of the middle nodular limestone/shaly interval varies systematically across the outcrop belt. This interval consists of tightly interlocked 2 to 6 cm-wide nodules with thin discontinuous shaly partings (classic nodular limestone fabric of Garrison and Fischer, 1969; Kukal, 1975; Raiswell, 1987) in west-central New York. This texture grades into numerous 10 to 20 cm-wide, diffusely bounded concretionary bodies in marlstone in east-central New York, and eventually into a few layers of 10 to 30 cm-wide, sharply bounded concretions in slightly calcareous silty shales in eastern New York (Figure 3).

The upper massive limestone beds are bioturbated, but they retain some primary sedimentary structures. Sharp, erosional contacts at the base of beds and relict, planar omission surfaces are locally preserved in these beds. The argillaceous omission surfaces commonly display protrusive, lined and unlined burrows, in contrast to the retrusive burrows that dominate most Cherry Valley beds.

Complete skeletons and very coarse-grained skeletal debris most commonly occur in the lower and upper massive limestone beds. These skeletal accumulations, recognized in the condensed Cherry Valley Member in west-central New York, constitute thin, highly concentrated intervals. For example, Clarke (1901b) and Flower (1936) noted that cephalopods of the Cherry Valley Member are most common in one or two intervals in central New York. In contrast, the numerous "shell beds" of the Cherry Valley Member in east-central and eastern New York are less concentrated. This makes the correlation of individual skeletal accumulations between west-central and east-central New York sections difficult. Although goniatites are common throughout the limestones of the Cherry Valley Member, orthocone nautiloids appear to be more common in the upper massive limestone beds. A limited number of bedding plane exposures in west-central New York exhibit a roughly unidirectional, southeasterly orientation from the orthocones (see Figure 4). This orientation is nearly perpendicular to the isopachs for the lower part of the Marcellus Formation (compare with Figure 2). Only 15% to 30% of the complete goniatite and nautiloid conchs in these skeletal concentrations have a non-horizontal orientation. Reoriented geopetal fills of some vertically or subvertically oriented cephalopods indicate a post-cementation sedimentary reworking origin for such orientations. Horizontal geopetal fills in other non-horizontal conchs may represent either original vertical emplacement or early reorientation by burrow activity.

Complete conchs of _Agoniatites_ _vanuxemi_ commonly display fine, well preserved ornamentation. However, partial preservation of cephalopods has long been noted as a characteristic of the Cherry Valley Member (Flower, 1936). Partial preservation of Cherry Valley cephalopods is most commonly due to synsedimentary fragmentation. Loss of the upper parts of walls and septa in horizontally oriented conchs is also observed at packstone-marlstone contacts along microstylolitic surfaces, and at sedimentary discontinuities (e.g. the Cherry Valley-Oatka Creek Member contact in west-central New York).

Cherry Valley Member cephalopod concentrations are closely associated with thickets of in situ auloporids. Auloporids are uncommonly found as encrustations on goniatite conchs. Coleolid tubes commonly occur with cephalopod concentrations as horizontally oriented broken fragments, but they are also locally clumped into patches of vertically or subvertically oriented tubes. Cephalopod conchs also uncommonly display shell borings of probable sponge origin.

The original microfabric of calcitic dacryoconarids and auloporids is preserved in these strata, and they are the only limestone skeletal components that occur in the marlstones. Most other large skeletons display a post-compactional replacement of shell carbonate by a coarse, brown epitaxial calcite spar. Thin pyritic coatings line the interior of many cephalopod conchs and auloporid corallites. Surficial weathering of the pyrite and calcite spar gives many skeletal concentrations an orange, limonitic, moldic appearance on outcrop.

Cherry Valley Member at "5 Points" Quarry

The westernmost exposure of the Cherry Valley Member at the "5 Points" Quarry, near Lima (Section A, Figure 3), retains the three-part subdivision but differs significantly from more easterly exposures in composition and texture. It consists of light- to medium-gray-weathering crinoid-styliolinid grainstones and minor packstones. Large disarticulated crinoid ossicles are common throughout, but are most abundant in the upper interval, which is thin-bedded instead of massive. Fenestrate bryozoan fragments are common in the grainstones. In situ auloporids are abundant. In contrast, diagnostic cephalopods of the Cherry Valley Member have been found (G. Kloc, 1990, personal comm.) but are relatively uncommon by comparison with their abundance farther east.

Cherry Valley Member in Transition

The limestones of the Cherry Valley Member begin to show drastic changes in sedimentary fabrics with the influx of coarse-grained, siliciclastic sediments at Long Road Ravine (STOP 3), Albany County. The massive/bioturbated nature of the lower and upper subdivisions gives way to thin, heterolithic, terrigenous silt/sandstone and argillaceous limestone beds. Heterolithic fine quartz sandstone-skeletal packstone beds in the upper subdivision take on an alternating laminated and burrow reworked bedding style. Burrow reworking is greatly decreased relative to sections farther west. Primary packstone and grainstone bedding is also partly preserved in the middle nodular/shaly interval. Preservation of carbonate skeletons is moldic in the silty/sandy beds. Skeletal fragments in the uppermost

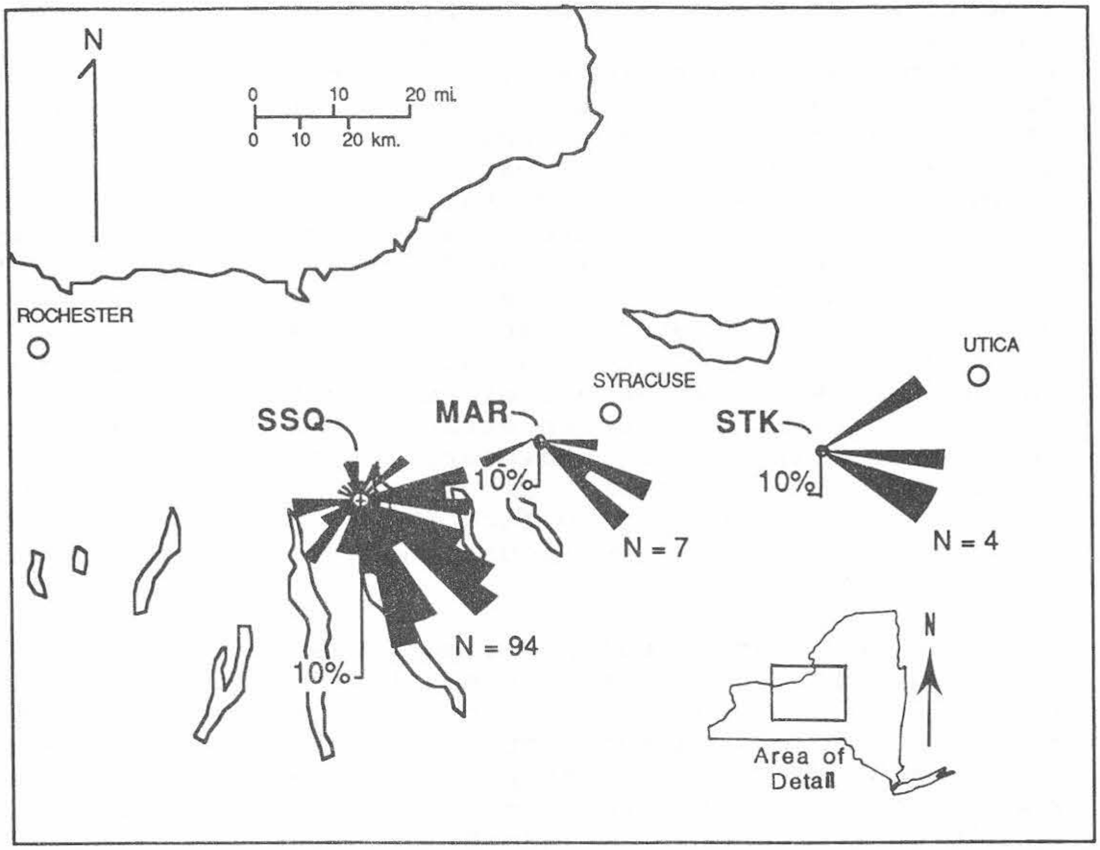

Figure 4.-- Paleocurrent roses generated from orthocone cephalopod orientations from the uppermost limestones of the Cherry Valley Member in west-central New York. SSQ = Seneca Stone Quarry, MAR = Pleasant Valley Road in Marcellus, STK = Stockbridge Falls.

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beds of the Cherry Valley Member are highly comminuted and abraded.

<u>Comparison with Other Devonian Cephalopod Limestones</u>

Relatively thin, red and black cephalopod-rich limestones represent a characteristic facies in Paleozoic and Mesozoic "miogeoclinal" strata of the world (Wendt and Aigner, 1982), particularly in the Middle Devonian to Lower Carboniferous of Europe and north Africa (Tucker, 1974; Bandel, 1974; Wendt and Aigner, 1985). These extensively studied Devonian cephalopod limestones share important similarities with the limestones of the Cherry Valley Member.

Wendt and Aigner (1985) defined several recurring subfacies in the Devonian-Carboniferous cephalopod limestones of Europe and north Africa (Morocco). These include (1) thick-bedded, crinoid-rich dacryoconarid packstones and grainstones; (2) dacryoconarid packstones and grainstones with abundant orthocone nautiloids and iron-stained discontinuity surfaces; (3) thin-bedded, orthocone-poor dacryoconarid nodular packstones; and (4) thin-bedded, nodular packstones in marly shale. Lateral changes in these subfacies (see Figure 5) also show strong similarities to the changes observed in the Cherry Valley Member. In addition, unidirectional orientations observed in the orthocones of the Moroccan strata are usually sub-perpendicular to isopachs (Wendt et al., 1984), as are those in the Cherry Valley Member of west-central New York. Cephalopods in the Moroccan strata are also found in highly concentrated intervals. However, cephalopod concentrations in the Moroccan limestones can range from 40 to 2000 per square meter of bedding surface (Wendt and Aigner, 1985), as compared with 5 to 15 per square meter in the Cherry Valley Member.

The Devonian cephalopod limestones of Europe and north Africa are interpreted to represent very slow pelagic and hemipelagic sedimentation on storm-swept, sediment-starved marine platforms and adjacent slopes (Tucker, 1974; Wendt and Aigner, 1985). Nodular packstone/marly shale sequences appear to represent more offshore platform and slope conditions (Figure 5). In addition, Wendt and Aigner (1985) suggested that the absolute depths of the platforms where cephalopods limestones accumulated ranged from near sea level to about 100 m.

A similar interpretation seems suitable for the Cherry Valley Member. The nodular/concretionary subfacies is associated with deeper water conditions in the scheme of Wendt and Aigner (1985). However, similar textures have been reported from rocks of shallow-water origin (Kukal, 1975). Textures of this subfacies probably represent early diagenesis resulting from an increase in the proximity to fine-grained, terrigenous sediment sources and from changes in the accumulation rates of terrigenous mud relative to biogenic carbonate (Raiswell, 1987). The lack of slump and debris flows (common in other Devonian cephalopod limestones) in the nodular limestones of the Cherry Valley indicates deposition on a gentler slope than its European and African counterparts. Less concentrated cephalopod accumulations in the Cherry Valley Member may simply reflect shorter periods of sediment starvation and/or less frequent reworking by strong bottom currents.

The faunas of the Chestnut Street Beds and the Cherry Valley Member are commonly associated with a change from exaerobic to dyaserobic and marginally aerobic sea-floor

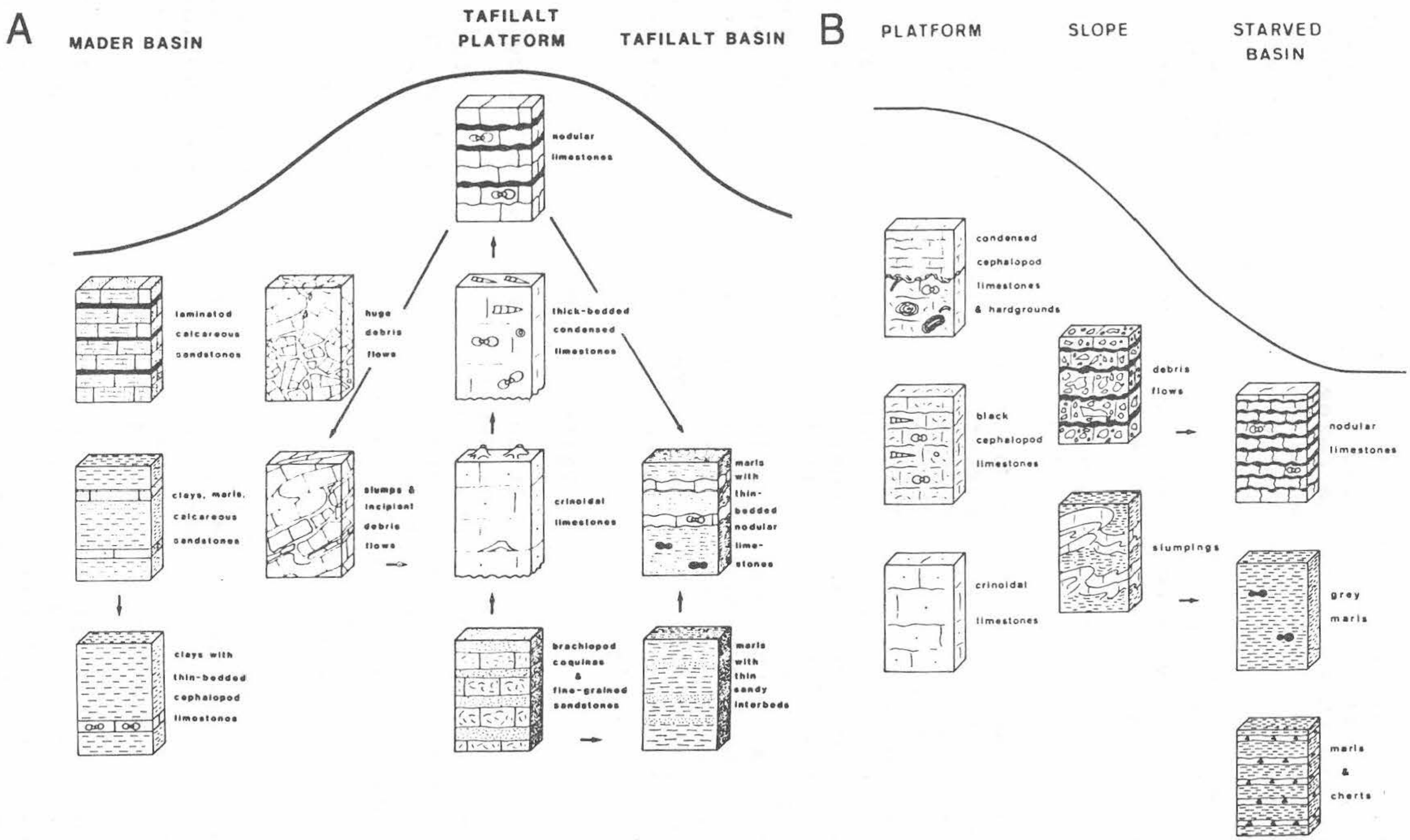

Figure 5.-- Generalized subfacies distribution for Upper Devonian cephalopod limestones in (A) the Anti-Atlas of Morrocco and in (B) the Montange Noire of France. From Wendt and Aigner (1985).

conditions (Brett and Kloc, in Anderson et al., 1988; Brower and Nye, 1991). Few unambiguous taphonomic indicators present are that aid in the determination of water depth for the deposition of these two limestone packages. However, relict graded beds, submarine erosional discontinuity surfaces, and oriented orthocone cephalopods within these limestone packages in western and west-central New York suggest that the sea floor in that area was shallow enough for episodic reworking by strong storms (Brett and Kloc, in Anderson et al., 1988; this report).

STRATIGRAPHIC RELATIONS IN THE LOWER PART OF THE MARCELLUS FORMATION (CAVS)

The lower part of the Marcellus Formation in eastern New York State is a southeasterly, rapidly thickening wedge of strata that ranges in thickness from approximately 12 m at Cherry Valley to more than 150 m at Kingston. Correlation of key units through this interval along the outcrop belt provides a more detailed understanding of relationships between the Cherry Valley, Stony Hollow, and Bakoven/Union Springs Members (see Figure 6). Recognition of strata in the Hudson Valley equivalent to the limestone-rich facies of the Cherry Valley Member permits correlation of the member into coarse, siliciclastic-dominated facies in the easternmost outcrops. A framework of the internal stratigraphy of the Stony Hollow Member and the upper parts of the Bakoven/Union Springs Members has also emerged and will be summarized below.

Cherry Valley Member

Figure 7 shows correlations for the Cherry Valley and Stony Hollow Members and the upper parts of the Bakoven/Union Springs Members along 143 km of the outcrop belt in eastern New York State. A key point to this new stratigraphic interpretation is the separation of the terrigenous sand-rich facies of the Cherry Valley Member from the underlying Stony Hollow Member (restricted) between Localities 4 and 13 (west of Clarksville to Kingston, respectively).

Cooper (1941) first described the Stony Hollow Member and regarded it as the sandy equivalent of the Cherry Valley Member. He noted that the upper part of the Stony Hollow Member at Onesquethaw Creek had a few layers of limestone that became the Cherry Valley Member to the west.

The stratigraphic revision proposed in this report is based upon the recognition of two relatively thin units with proetid trilobites and an intervening massive sandstone near the top of the Stony Hollow Member (discussed below). These data and use of the black shales of the Berne/Chittenango Members that overlie the Cherry Valley Member permit

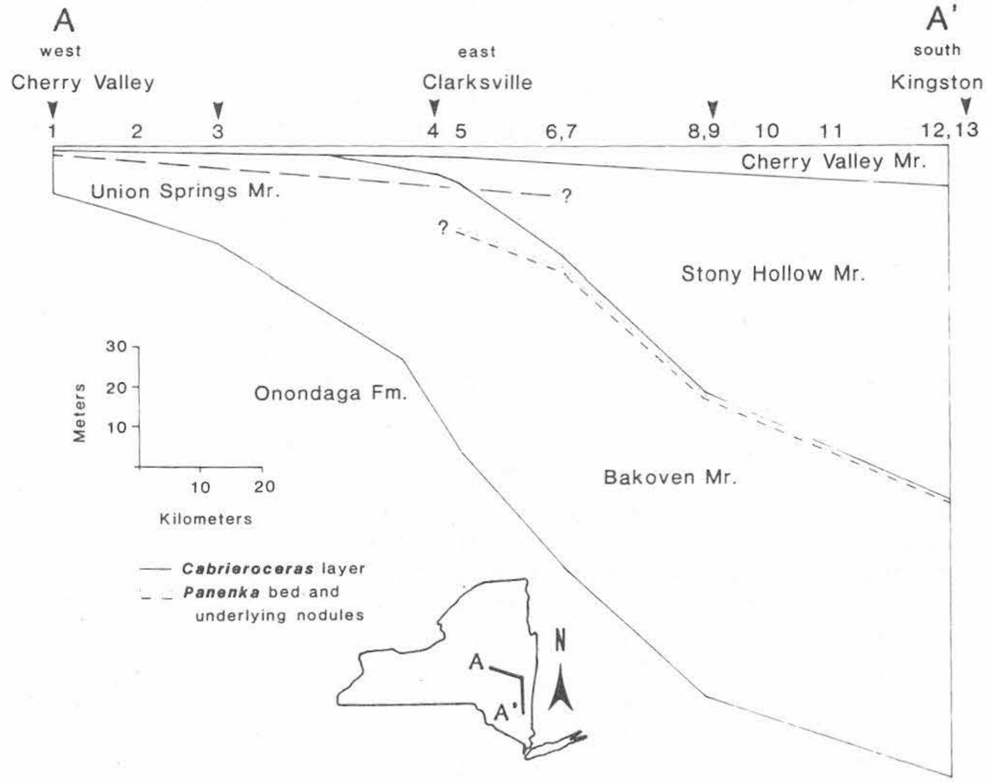

FIGURE 6.-- Thicknesses and relationships of the lower part of the Marcellus Formation in eastern New York State. Localities 1-13 are shown on Figure 8 and listed in Appendix II. Arrows indicate field trip stops (left to right, STOPS 1-5, respectively). Thicknesses of Bakoven/Union Springs Members from Rickard (1989).

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physical correlation of the classic limestone-rich facies of the Cherry Valley Member with the equivalent terrigenous sand-rich facies in the Hudson Valley. It is proposed herein that these equivalent sand-rich strata be separated from the top of the Stony Hollow Member (restricted) and be included in the Cherry Valley Member. Therefore, the Cherry Valley "Limestone" (designation abandoned) is only a facies of a more areally extensive lithosome termed the "Cherry Valley Member" and the term "Cherry Valley Limestone" is no longer valid.

The argillaceous sandstones of the Cherry Valley Member in the Hudson Valley are fine-grained and are generally highly bioturbated. Some sandstone layers may exhibit planar bedding to cross-stratification, a low degree of burrow mottling (including apparent escape burrows), and sole marks. Body fossils are not generally common in the sand-rich facies. However, calcareous bedded to nodular layers that feature small brachiopod and styliolinid shell hashes with abundant quartz sand do occur. Nautiloid and goniatite cephalopods are uncommon to rare in the terrigenous sandstone facies.

Between Cherry Valley and Clarksville, the Cherry Valley Member has three subdivisions and nearly doubles in thickness from 1.3 to 2.4 m. The member is not well exposed through much of the Hudson Valley, but a completely exposed section measures 10 m in thickness near Kingston. In the latter area, it is composed of five highly bioturbated arenaceous subdivisions separated by sandy shale intervals, all on the order of 1 m-thick.

Stony Hollow Member

"Proetid Units". Two richly fossiliferous units similar to the previously discussed Chestnut Street Beds occur in the upper part of the Stony Hollow Member. These two units (herein termed the "Upper" and "Lower Proetid Units") are characterized by a relatively rich benthic fauna that features proetid trilobites. Other faunal elements include abundant crinoid debris, auloporid and small rugose corals, and brachiopods. These are accompanied by uncommon orthoconic nautiloid and goniatitic cephalopods and fish bones. The Upper and Lower Proetid Units of the Stony Hollow Member generally are separated by a 3.2-6.0 m-thick interval that features a prominent massive to thin-bedded sandstone unit. It is not currently known whether the Chestnut Street Beds near the top of the Union Springs Member represent the Upper or Lower Proetid Unit or if the former represent both Proetid Units of the Stony Hollow Member.

The Upper Proetid Unit of the Stony Hollow Member in the northern part of the outcrop belt is a coarse crinoidal limestone faunally similar to the Chestnut Street Beds near the top of the Union Springs Member to the west. South of Clarksville (Locality 6 and vicinity), the Upper Proetid Unit appears as a brachiopod-rich bed, dominated by Pacificocoelia? and athyridacean brachiopods. In the Kingston area (Locality 12), the Upper Proetid Unit is represented by a series of thin, fossiliferous, proetid-bearing beds that occur in sandy shales.

The Lower Proetid Unit of the Stony Hollow Member underlies a thin-bedded to massive sandstone unit near the top of the Stony Hollow Member. The faunal assemblage of the lower unit is nearly identical to that of the upper unit and has proetid trilobites, crinoid debris, brachiopods, and small rugose corals. At some localities, the Lower Proetid Unit overlies a brachiopod-rich bed (Localities 6, 7, 12, and 13/STOP 5b). At Locality 9 (STOP 4), the lower proetid unit is composed of two or more beds that total 0.6 m in thickness.

<u>Massive Sandstone Unit</u>. A thick, massive to thin-bedded sandstone occurs between the Upper and Lower Proetid Units near the top of the Stony Hollow Member. The sandstone unit is generally on the order of 4 m in thickness and commonly forms the caprock of prominent waterfalls and roadcuts. This argillaceous, quartz-rich sandstone is generally highly bioturbated, although it appears less burrowed and more thin-bedded in the northern and southern parts of the study area. Body fossils are less common in the massive sandstone unit than in the underlying and overlying Proetid Units.

<u>Other Regionally Correlatable Units Within the Stony Hollow Member</u>. At least three other prominent units within the Stony Hollow Member are locally correlatable along the outcrop belt. The first is a resistant, argillaceous, fine-grained sandstone that features scattered auloporid corals and several thin beds with small streptelasmid rugose corals. This unit occurs 9-12 m below the Lower Proetid Unit and ranges from 3.5 to 6+ m in thickness. It is moderately to highly bioturbated and features scattered small fossil debris and uncommon fish plates. This unit also forms the caprock of some exposures, as in Plattekill Creek in the village of Mount Marion, 10.0 km north of Kingston.

Another correlatable sandstone unit occurs approximately 5-9 m below the unit described above (13-23 m below the Lower Proetid Unit) between Kingston and north of Catskill. In the Kingston area (Localities 12 and 13/STOP 5b), it is represented by 0.9 m of buff-weathering, fine-grained sandstone. This sandstone may be subdivided into two subunits: a lower, more argillaceous part that contains abundant skeletal components and an upper, more resistant part that is highly bioturbated and contains scattered auloporid and small rugose corals. North of Catskill (Localities 8 and 9/STOP 4), the sandstone (1.6 m-thick) forms the caprock of several waterfalls.

A 7-10 m-thick interval of thin, buff-weathering beds that occurs in the lower part of the Stony Hollow Member is also correlatable between Kingston and the Catskill area. These thin dolomitic beds, which range from 2 to 22 cm in thickness, form resistant, buff-weathering bands on highly weathered outcrops. They appear to be moderately to highly burrow-mottled in contrast to the thinly laminated character of the surrounding strata. Small body fossils (nowakiids and small brachiopods) are uncommon to rare in the thin dolomitic beds.

As the Stony Hollow Member thins north of Localities 8 and 9 (STOP 4), these three units become indistinct. At Localities 6 and 7, an interval with occasional small chonetid brachiopods characterizes the middle part of the Member.

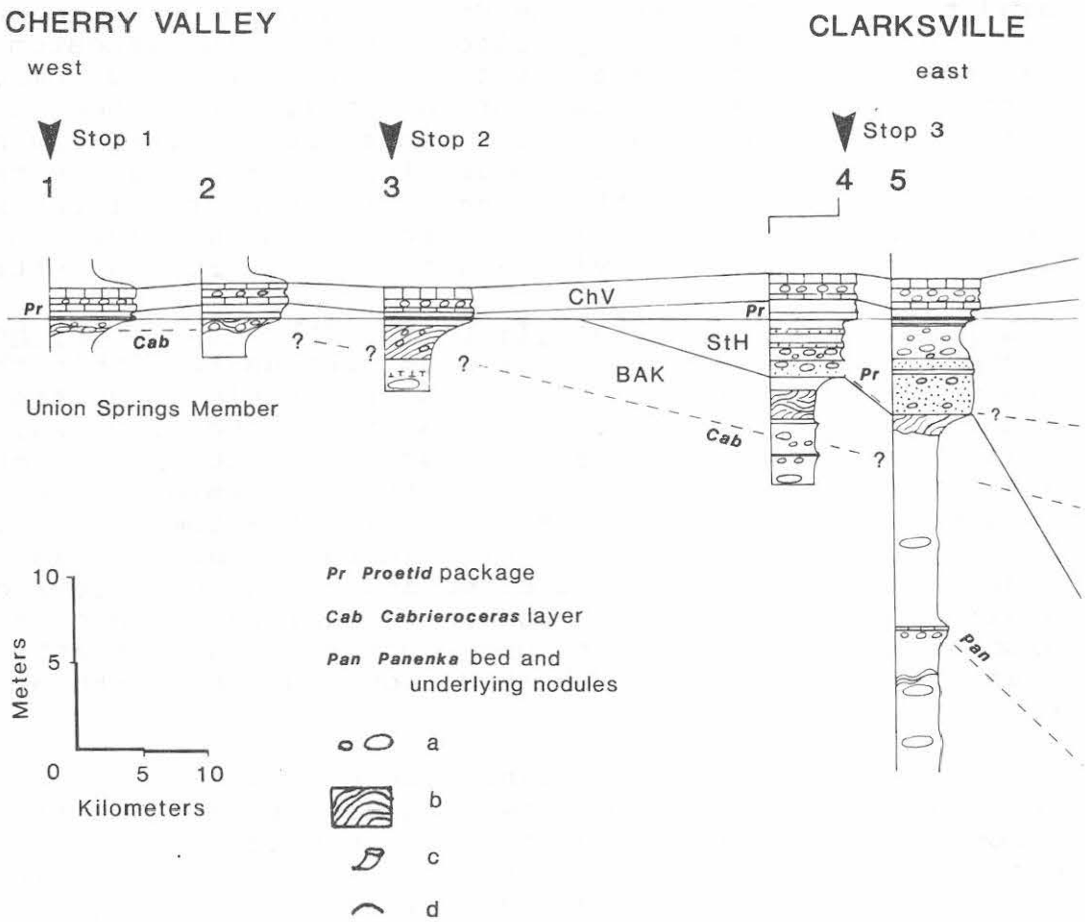
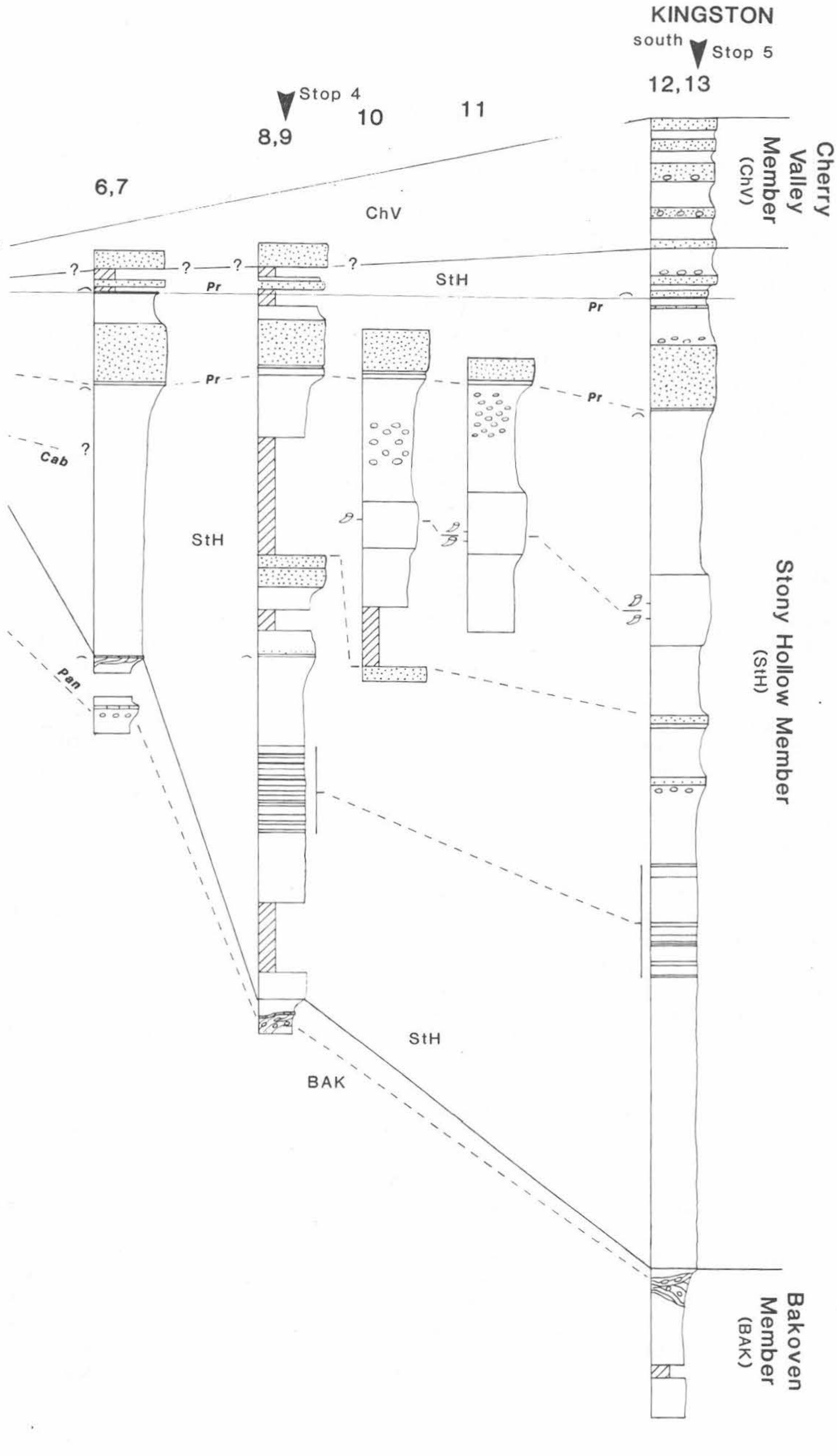


FIGURE 7.-- Correlations in the lower part of the Marcellus Formation across the field area. Datum equals the Upper Proetid Unit/Chestnut Steet Bed. Localities are shown on Figure 8 and are listed in Appendix II.
a, calcareous nodule horizons; b, tectonized shale intervals; c, rugose coral beds; d, brachiopod-rich beds.



KINGSTON
south
Stop 5
12,13
Stop 4
8,9
10
11
6,7
ChV
StH
Pr
Cab
Pan
BAK
Cherry Valley Member (ChV)
Stony Hollow Member (StH)
Bakoven Member (BAK)

Bakoven/Union Springs Members

Tectonized Shale Interval. The uppermost strata of the Bakoven and Union Springs Members in eastern New York are marked by a widespread, tectonized shale interval. The interval occurs at, or within a meter below, the contact of the black shales with the overlying Stony Hollow Member, or in the absence of the latter, at or near the base of the Chestnut Street Beds near the top of the Union Springs Member. This tectonized shale interval has been interpreted as a decollement produced during late stages of the Acadian or Alleghenian Orogeny (see Bosworth, 1984 a, b). Faults in the Stony Hollow and/or Cherry Valley Members at Localities 1 (STOP 1) and 5 may be associated with the same thrusting event. Similar smaller scale structures occur throughout other parts of the Bakoven and Union Springs Members, where they generally underlie concretionary to thin-bedded limestone layers.

Cabrieroceras Bed. A series of calcareous nodule layers occurs in the upper part of the Union Springs Member at Cherry Valley (Locality 1/STOP 1). A layer of larger nodules approximately 0.7 m below the Upper Proetid Bed features the goniatite Cabrieroceras. This unit was previously noted by Flower (1936, p. 236; 1943, p. 17-18, "Werneroceras Bed") and Miller (1938, p. 60, "Anarcestes Limestone") and was correlated through central New York State by Flower (1936). Brett and Kloc (in Anderson et al., 1988, p. 123) informally renamed the unit the Cabrieroceras Bed and noted that Rickard (1952, 1981) and subsequent authors (see House, 1962, 1981; Klapper, 1971, 1981) confused Flower's nodular Werneroceras Bed with the overlying Chestnut Street Beds. Rickard (1952) did note the true Cabrieroceras Bed approximately 3.4 m below the base of the Stony Hollow Member at Locality 4 (STOP 3) of this report. The Cabrieroceras Bed appears to pass laterally into the Stony Hollow Member south of Clarksville. Cabrieroceras has been found in the upper part of the Stony Hollow Member in northern Greene County (Localities 6 and 7).

Panenka Bed. A thin, condensed, fossil-rich limestone (ca. 0.1-0.3 m-thick) occurs in the upper part of the Bakoven Member in the northern part of the Hudson Valley outcrop belt. This unit, herein termed the "Panenka Bed", is characterized by a fauna of small brachiopods and bivalves, accompanied by the large bivalve Panenka. Gastropods, goniatites, and orthoconic cephalopods are also common in the unit. At Onesquethaw Creek (Locality 5), orthoconic cephalopods on the top surface of the Panenka bed exhibit a distinct north-south orientation. The Panenka Bed underlies the base of the Stony Hollow by approximately 12 m at Onesquethaw Creek. As the Stony Hollow Member thickens south of Clarksville, the interval between the base of it and the Panenka Bed thins, and at Locality 8, the two units are separated by 1.0 m of black shale. The Panenka Bed disappears north of Kingston.

The Panenka Bed overlies another calcareous nodular layer with goniatites (including Cabrieroceras). This nodular unit generally underlies the Panenka Bed by approximately 0.5 m in the northern part of the Hudson Valley and appears to correlate with a similar nodular unit within the tectonized shale interval at the top of the Bakoven at Kingston (Locality 13/STOP 5b).

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APPENDIX I

The following list is a locality directory for the lettered stratigraphic sections of the Chestnut Street Beds and the Cherry Valley Member as featured in Figure 3.

Section A - Exposures along the southern lip of the General Crushed Stone ("5 Points") Quarry, located 3.7 km northwest of the village of Lima, Livingston County.
Section B - Low falls on Flint Creek, located 0.5 km northeast of the Wheat Road bridge that crosses Flint Creek and 2.9 km southwest of the village of Phelps, Ontario County.
Section C - Exposures along the southern lip of Seneca Stone Quarry, located 5.1 km south-southeast of the village of Seneca Falls and 1.8 km west of the village of Canoga Springs, Seneca County.
Section D - Low falls directly north of Pleasant Valley Road, located 1.4 km east of the intersection with N.Y. Rte. 175, village of Marcellus, Onondaga County.
Section E - Low falls 0.3 km south of N.Y. Rte. 173, located 3.1 km west of the village of Manlius and 4.9 km east of the village of Jamesville, Onondaga County.
Section F - Roadcut on the north side of Stockbridge Falls Road, located 3.4 km southwest of the village of Munnsville, Madison County.
Section G - Low falls in creek along Gulf Road, located 3.1 km east of the village of West Winfield, Herkimer County.
Section H - High falls in Cox Ravine (type-section of the Cherry Valley Member), located 1.1 km northwest of the village of Cherry Valley, Otsego County.
Section I - Chestnut Street roadcut (STOP 1), located directly southwest of the intersection of Chestnut Street (Otsego Co. Rte. 54) and U.S. Rte. 20, 3.8 km east of the village of Cherry Valley, Otsego County.
Section J - Low falls in the ravine 0.1 km north of Rosenberg Road, located 0.95 km west of the junction with Schoharie Co. Rte. 40 and 2.7 km northwest of the village of Seward, Schoharie County.
Section K - Stream cut exposures, located 0.25 km south of Schoharie Co. Rte. 1 in the village of Mineral Springs (STOP 2), Schoharie County.
Section L - Stepped falls in the ravine directly north of Long Road (STOP 3), located 3.8 km from the junction with N.Y. Rte. 443 and 1.8 km southeast of the village of Thompsons Lake, Albany County.

APPENDIX II

The following list is a locality directory for the numbered stratigraphic sections of the lower part of the Marcellus Formation that are shown in Figures 7 and 8.

1. Outcrop near intersection of Chestnut Street and U.S. Rte. 20, 3.8 km northeast of Cherry Valley (Otsego Co.). Equals Section I of Appendix I.
2. Waterfall in stream along Rosenberg Road, 2.7 km northwest of Seward (Schoharie Co.). Equals Section J of Appendix I.
3. Outcrop in outlet stream of Cobleskill Reservoir, 200 m south of Mineral Springs (Schoharie Co.). Equals Section K of Appendix I.
4. Ravine along Long Road, 2.3 km southeast of Thompsons Lake (Albany Co.). Equals Section L of Appendix I.
5. South branch of Onesquethaw Creek, 3.7 km northwest of Clarksville (Albany Co.).
6. Ravine south of farm pond along Greene Co. Rte. 51, 3.9 km north-northwest of Roberts Hill (Greene Co.).
7. Ravine south of Hass Hill Road, 1.5 km southwest of Roberts Hill (Greene Co.).
8. Ravine, south side of power lines, 0.4 km west of south end of Hollister Lake (Greene Co.).
9. Buttermilk Falls, 1.0 km west of Green's Lake (Greene Co.).
10. Underhill Road, 0.6 km north of New York Rte. 23a, 2.3 km east-northeast of Kiskatom; and north side of ravine, 1.3 km north-northeast of Underhill Road outcrop (Greene Co.).
11. Roadcut along New York Rte. 32, 1.0 km west of Katsbaan (Ulster Co.).
12. Roadcuts along New York Rte. 209, 1.2 km north of toll booth, Exit 19 of New York State Thruway (Ulster Co.).
13. Roadcuts and railroad cut along New York Rte. 28, 1.2-3.2 km west-northwest of toll booth, Exit 19 of New York State Thruway (Ulster Co.).

ROAD LOG

The road log begins at Exit 17 of Interstate Route 88, east of Oneonta, New York.

CUMULATIVE MILEAGE	MILES FROM LAST POINT	ROUTE DESCRIPTION
0.0	0.0	Leave I-88 eastbound at Exit 17.
0.3	0.3	Turn left (north) onto N.Y. Rte. 28.
1.8	1.5	Cross intersection with N.Y. Rte. 7. Continue north on Rte. 28. Note exposures of Gilboa Fm. in roadcuts north of this intersection.
9.2	7.4	Enter village of Milford.
9.6	0.4	Turn right (east) onto N.Y. Rte. 166 northbound at the junction.
10.9	1.3	Note exposures of the Panther Mountain Fm. in small roadcut on the north side of the road.
24.5	13.6	Enter village of Roseboom.
26.7	2.2	Cross junction with Otsego Co. Rte. 33. Continue north on Rte. 166. Note outcrop of lower Hamilton Group shales on west side of the road, north of the junction.
28.5	1.8	Enter village of Cherry Valley.
28.8	0.3	Turn right (east) at intersection with Otsego Co. Rte. 54. Continue on combined Rte. 166 northbound/54 eastbound.
29.4	0.6	Veer right onto Rte. 54 (leaving Rte. 166) at fork.
31.9	2.5	Note roadcut of lower strata of the Marcellus Fm. on the right (south) side of the road.
32.1	0.2	Park at the east end of the roadcut. Note the panorama view of the Mohawk Valley and the southern Adirondacks to the north.

STOP 1. Chestnut Street Roadcut

This outcrop features the uppermost part of the Union Springs Member, an excellent exposure of the complete Cherry Valley Member, and the lowermost part of the Chittenango Member. Two thin zones of intensely deformed shale within the

Union Springs Member and a thrust fault that ramps up through the Cherry Valley Member can be observed at this locality. Bosworth (1984a, b) associated these deformation features with northwest-directed overthrusting that occurred during either Acadian or Alleghenian orogenic events.

The exposed 4.0 m of Union Springs Member shales contain several intervals of carbonate concretions at this locality. One on these concretion intervals commonly features the goniatite Cabrieroceras pleibeforme and lies approximately 1.0 m below the base of the Cherry Valley Member. The Chestnut Street Beds form one 20 cm-thick ledge of skeletal wackestones and packstones around 17 to 20 cm below the base of the Cherry Valley Member. Although the individual Chestnut Street Beds are amalgamated and burrow-mottled, distinction of the uppermost darker gray, crinoid-rich packstone bed can be made on outcrop. A thin, silty, calcareous, gray to black shale interval between the Chestnut Street Beds and the Cherry Valley Member features the goniatite Agoniatites nodiferous. The clean weathered faces of this roadcut make it one of the best places to view internal stratigraphy of the Cherry Valley Member. The 1.2 m-thick section of Cherry Valley contains a 15 to 20 cm-thick lower massive packstone bed, a 60 to 65 cm-thick middle nodular packstone/marlstone interval, and a 37 to 40 cm-thick package of upper packstone beds. Concentrations of whole and partially preserved goniatites can be observed in the basal bed and upper packstone intervals. Orthocone nautiloids can also be observed in the base of the upper packstone interval, especially 30 cm below the upper contact of the Cherry Valley Member. Planolites traces are common on the bedding plane surfaces of fallen blocks of the Cherry Valley Member. Calcareous black shales with small brachiopods, dacryoconarids, and small orthocone nautiloids mark the transition from the Cherry Valley Member into the overlying sooty black shales of the Chittenango Member.

32.1	0.0	Proceed east on Otsego Co. Rte 54.

FIGURE 8.--Generalized geologic and locality map. Geologic map modified from Fisher et al. (1970). Explanation: (Don) Onondaga Limestone and lower units; (Dhm) lower part of the Marcellus Formation (Bakoven/Union Springs, Stony Hollow, and Cherry Valley Members), Mount Marion Formation, and Panther Mountain Formation to north (Hamilton Group); (Dhmr) Marcellus Formation (Hamilton Group); (Dh) Panther Mountain Formation and higher units; (Dhc) continental facies of the Hamilton group and higher units. Localities are listed in Appendix II.


N
CHERRY VALLEY
Otsego Lake
Stop 1
1
Don
Dh
Dh
Dh
2
Dhmr
Stop 2
3
SCHOHARIE
Stop 3
4
ALBANY
Don
OTSEGO CO.
Dh
Dhmr
5
Dhm
CLARKSVILLE
SCHOHARIE CO.
Alcove Res.
ALBANY CO.
Dhc
COEYMANS HOLLOW
6,7
Dhm
HUDSON RIVER
GREENE CO.
Don
Dhc
Stop 4
8,9
CATSKILL
Study Area
10
ULSTER CO.
11
SAUGERTIES
Miles
0
5
15
0
10
20
Kilometers
Don
Ashokan Res.
Dhm
KINGSTON
12,13
Stop 5

32.2	0.1	Turn left (north) off Rte. 54 and then right (east) onto U.S. Rte. 20.
34.4	2.2	Note outcrops of Kalkberg Fm. along U.S. Rte. 20 west of Sharon Springs.
35.7	1.3	Enter village of Sharon Springs.
36.1	0.4	Cross intersection with N.Y. Rte. 10. Continue on Rte. 20.
39.2	3.1	Note outcrops of Onondaga Limestone.
40.0	0.8	Enter village of Sharon.
41.2	1.2	Turn right (south) onto N.Y. Rte. 145.
45.0	3.8	Note exposures of the Union Springs Mbr. of the Marcellus Fm. and the underlying Onondaga Limestone in the drainage ditch on the right (west) side of the road for the next 0.4 miles.
48.6	3.6	Note roadcut in Onondaga Fm. on the right (west) side of the road.
48.7	0.1	Enter village of Lawyersville.
49.0	0.3	Turn left (east) at stop sign, remaining on Rte. 145.
50.0	1.0	Enter village of Cobleskill.
51.1	1.1	Continue straight through the intersection with N.Y. Rte. 7 (at stop light) onto South Grand St.
51.6	0.5	Proceed under overpass of I-88.
51.8	0.2	Turn left (east) onto Mineral Springs Rd. (Schoharie Co. Rte. 1).
53.3	1.5	Turn right (south) onto Green Hill Rd.
53.4	0.1	Turn left into vacant lot and park. Walk down into the ravine at the back of the lot and proceed upstream to outcrops.

STOP 2. Mineral Springs Section.

This stop exposes the upper 4.5 meters of the Union Springs Member, a complete section of the Cherry Valley Member (1.4 m-thick), and the lowest part of the Chittenango/Berne Member. Calcareous dark gray shales a meter above the base of the section contain small brachiopods and bivalves; these include the brachiopod Leiorhynchus and the large bivalve Panenka, along with nautiloid and goniatite cephalopods. A 2 m-thick interval of tectonized shale occurs in the upper part of the Union Springs Member above these fossiliferous shales. The Chestnut Street Beds form a 20 cm-thick package of two skeletal wacke/packstone beds separated by gray calcareous shales. The basal Chestnut Street bed directly overlies a 5 cm-thick dark, argillaceous concretionary lime mudstone layer with pyritized burrows. The internal stratigraphy of the Cherry Valley Member at this locality is similar in most ways to that of STOP 1. However, individual packstone beds of the upper massive interval increase in thickness and are separated by shaly partings.

53.4	0.0	Return to junction with Mineral Springs Rd.
53.5	0.1	Turn right (east) onto Mineral Springs Rd.
54.5	1.0	Note hillside exposure of the Chittenango Mbr. of the Marcellus Fm. on the south side of the road.
55.4	0.9	Turn left (north) onto N.Y. Rte. 145.
55.5	0.1	Turn right onto I-88 eastbound at Exit 22.
55.8	0.3	Note Carlisle Center and Esopus Fms. in roadcuts.
56.1	0.3	Note Oriskany and Becraft Fms. in roadcuts south of I-88.
56.8	0.7	Note Becraft and Kalkberg Fms. in roadcuts south of I-88.
57.2	0.4	Note Coeymans Fm. in low roadcuts on the south side of the highway.
59.2	2.0	Note exposures of the Manlius, Rondout, and Brayman Fms. in large roadcut to the south of I-88.
61.1	1.9	Leave I-88 eastbound at Exit 23.
61.5	0.4	Turn right (south) onto N.Y. Rte. 30A.
62.3	0.8	Rte. 30A ends. Veer right and continue southbound on N.Y. Rte 30.
63.8	1.5	Turn left (east) onto N.Y. Rte. 443.
64.0	0.2	Enter village of Shutters Corners.
67.1	3.1	Note exposures of Onondaga Limestone in roadcuts and in adjacent Fox Creek for the next 1.5 miles.
69.1	2.0	Cross Albany County line.
69.5	0.4	Enter village of West Berne.

71.7	2.2	Note Onondaga Limestone exposures behind Highway Dept. Building (south of Rte. 443).
71.8	0.1	Enter village of Berne.
72.5	0.7	Continue on Rte. 443, through the intersection with N.Y. Rte. 156 (around 2 sharp bends).
75.3	2.8	Cross junction with Cole Hill Rd. Note that large roadcuts on Cole Hill Rd. feature the type-Berne and the complete Otsego Mbrs. of the Mount Marion Fm. (= proximal marine facies of the upper part of the Marcellus Fm.).
76.0	0.7	Cross junction with N.Y. Rte. 157A. Continue on Rte. 443.
76.1	0.1	Enter village of East Berne.
76.5	0.4	Turn left (east) onto Long Rd.
78.0	1.5	Cross intersection with Saw Mill Rd. Continue on Long Rd.
78.7	0.7	Note small quarry in the Berne Mbr. (south of the road).
79.05	0.35	Park on the side of Long Rd. Walk down into the deep, heavily wooded ravine directly north of the road.

STOP 3. SECTION AT LONG ROAD RAVINE (THOMPSON'S LAKE SECTION OF RICKARD, 1952)

This section illustrates the eastward appearance of coarse-grained siliciclastics in the lower part of the Marcellus Formation. The stop exposes a relatively thick section of the Bakoven Member, a 4.5 m-thick section of the Stony Hollow Member, and a 1.4 m-thick section of the Cherry Valley Member.

The Bakoven Member (= Union Springs Member) is exposed downstream from and within the lowermost waterfall. The shales contain carbonate concretion horizons and thin-bedded limestones. Rickard (1952) recognized the true concretionary Cabrieroceras bed near the base of the lowermost waterfall. A 1.75 m-thick interval of tectonized shale occurs in the face of this waterfall.

A 2.0 cm-thick bed of the Lower Proetid Unit occurs at the base of the Stony Hollow Member, just below the crest of the lowermost waterfall. The overlying 1.8 meters of buff-weathering, pyrite-rich, calcareous siltstones and shales are equivalent to the massive sandstone unit near the top of the Stony Hollow in the Hudson Valley (STOPS 4 and 5b). The overlying Upper Proetid Unit consists of a package of coarsely crinoidal packstone and grainstone beds that display relict cross-stratification. Quartz silt and sand are common in these limestone beds at this locality, in contrast to previous stops.

The Cherry Valley Member is exposed within and caps the uppermost waterfall. The lower and upper Cherry Valley intervals are heterolithic at this locality and contain thin interbeds of terrigenous siltstone/sandstone with skeletal packstone and grainstone beds. The middle nodular interval consists of a few large, sharply bounded concretions in dark gray shale. Coarse crinoidal beds with finely comminuted brachiopod valves are common in the uppermost Cherry Valley.

79.2	0.15	Turn right (south) onto Elm Dr.
81.2	2.0	Turn right (west) onto Stage Rd. (Albany Co. Rte. 303) at stop sign.
81.6	0.4	Veer left at fork near the junction of combined N.Y. Rte. 443/85 and Stage Rd. Turn left (east) onto Rte. 443/85. Note sandstone at the top of the Berne Mbr. (Mount Marion Fm.) on right past the intersection.
82.3	0.7	Note Helderberg Lake on left (north) side.
82.5	0.2	Cross junction with Duck Hill Rd. Note exposures of the Berne Mbr. behind the house at the corner.
82.9	0.4	Note the large parking area on the north side of the road, which provides parking for access to exposures of the lower part of the Marcellus Fm. in and below the falls on Onesquethaw Creek, directly to the north.
83.9	1.0	Continue on Rte. 443 as Rte. 85 turns to the left (east).
85.2	1.3	Note exposure of Onondaga Limestone in roadcut on north side.
85.3	0.1	Enter village of Clarksville.
85.4	0.1	Turn right (south) onto Clarksville South Rd. (Albany Co. Rte. 312). Continue on this route past the intersection with Bennett Hill Rd.
87.5	2.1	Note quarry in the Otsego Mbr.

91.3	3.8	Turn left (east) onto N.Y. Rte. 143 in the village of Dormansville. Note outcrops of the upper part of the Mount Marion Fm. on the south side of the intersection.
91.5	0.2	Turn right (south) onto N.Y. Rte. 32 at the intersection with Rte. 143.
91.8	0.3	Note quarry in the upper part of the Mount Marion Fm. on the right (west) side of the road.
97.2	5.4	Enter village of Greenville. Continue on Rte. 32.
97.8	0.6	Cross junction with N.Y. Rte. 81. Continue on Rte. 32.
99.4	1.6	Note outcrops of the Plattekill Fm. along the road for the next 0.4 mile.
101.3	1.9	Note scenic views of the northern Catskills for the next 0.5 mile.
101.9	0.6	Enter village of Freehold. Continue on Rte. 32.
105.6	3.7	Cross Catskill Creek.
106.7	1.1	Rte. 32 joins with N.Y. Rte. 23 and heads east. Continue on combined Rte. 32/23. Note exposures of Plattekill Fm. red beds and fluvial sandstones in roadcuts.
107.7	1.0	Rte. 32 veers right (south). Continue east on Rte. 23.
108.4	0.7	Note additional outcrops of the Plattekill Fm. for the next 2.4 miles.
112.0	3.6	Note outcrops of the Mount Marion Fm. over the next 0.4 mile.
113.1	1.1	Turn left (north) onto Cauterskill Rd.
113.4	0.3	Turn right (east) onto Old Rte. 23 (N.Y. Rte. 23B). Cross old stone bridge over Catskill Creek.
113.6	0.2	Turn left (north) onto Greene Co. Rte. 49 in the village of Leeds.
114.1	0.5	Cross intersection with Sandy Plains Rd.
114.9	0.8	Note scenic view of Potic Mtn. to the left (west).

115.5	0.6	Note outcrops of the Onondaga Limestone in roadcuts for the next 0.2 mile.
115.8	0.3	Turn left (west) onto Buttermilk Falls Rd.
116.4	0.6	Turn vehicles around and park in the parking area on the south side of Buttermilk Falls Rd. Follow dirt trail southward to waterfalls.

STOP 4. BUTTERMILK FALLS

This locality exposes approximately 41 m of the Stony Hollow Member and 3.7 m of the terrigenous sand-rich facies of the Cherry Valley Member. Strata at the base of the lowest falls lie approximately 12 m above the Bakoven/Stony Hollow Member contact, which is exposed in a ravine 2.1 km to the north (Locality 8).

A 7 m-thick interval in the lowest falls features a series of thin dolomitic beds (ca. 2-22 cm-thick). This interval, which is best seen in the high bank north of the lowest falls, correlates with a similar package of thin dolomitic beds in the lower part of the Stony Hollow Member at Kingston (Locality 13/Stop 5b). Fine-grained resistant sandstones forms the caprock of the lowest falls.

A second falls 4 m upsection is capped by 1.6 m of highly bioturbated, argillaceous sandstone. The lower part of the unit appears medium bedded, and the upper part is massive. Prominent subhorizontal to vertical Zoophycus traces occur in the upper 0.9 m. This unit is also traceable to Kingston (Locality 13, STOP 5b), where it is represented by 0.9 m of fossiliferous, bioturbated fine sandstone.

The main upper waterfall at Buttermilk Falls is capped by the massive sandstone (3.3 m-thick) near the top of the Stony Hollow Member. At this locality, the massive sandstone unit is highly bioturbated and displays a prominent slaty cleavage. A 0.3 m-thick recessed interval of sandy shale separates the massive sandstone from the underlying Lower Proetid Unit. This highly fossiliferous unit is composed of 2-3 separate beds that total 0.6 m in thickness at Buttermilk Falls.

The massive sandstone that caps the main falls at Buttermilk Falls is overlain by siltstones and silty shales in the upper part of the Stony Hollow Member. The Upper Proetid Unit is covered at STOP 4. A 0.6 m-thick sandstone that forms a small cascade upstream of the main falls may correlate with a similar thin unit that overlies the Upper Proetid Unit or the Chestnut Street Beds at most localities.

The Cherry Valley Member forms a 2 m-high falls at the top of the outcrop. The argillaceous sandstone of the Member is highly cleaved and has several recessed carbonate-rich intervals with numerous small brachiopods. The sandstones appear to be highly bioturbated. A small outcrop of sandy shales that is exposed approximately 5 m upstream is apparently part of the Cherry Valley Member. An abandoned quarry in the south bank of the creek exposes the lowermost black shales of the overlying Berne Member.

116.4	0.0	Retrace route (east) on Buttermilk Falls Rd.
117.0	0.6	Turn right (south) onto Greene Co. Rte. 49 and return to the village of Leeds.

TO RETURN TO ONEONTA:

119.2	2.2	Turn right (west) onto N.Y. Rte. 23b.
119.4	0.2	Turn left (south) onto Cauterskill Rd.
119.7	0.3	Turn right (west) onto N.Y. Rte 23 and follow highway to Oneonta.

TO OPTIONAL STOP 5: ROUTE 28 EXPOSURES, KINGSTON

119.2	2.2	Turn left (east) onto N.Y. Rte. 23b.
120.3	1.1	Turn left (north) at entrance to N.Y. State Thruway.
120.5	0.2	Stop and get ticket at tollbooth. Enter southbound ramp toward New York City. Note outcrops of the Lower Devonian New Scotland and Becraft Fms. for the next 1.2 miles.
121.8	1.3	Cross Catskill Creek.
122.1	0.3	Note outcrops of the Tristates Group (Esopus, Carlisle Center, and Schoharie Fms.) and the Onondaga Fm. for the next 8.9 miles.
123.9	1.8	Cross Kaaterskill Creek.
131.9	8.0	Note exposures of the Glenerie Fm. and the underlying upper part of the Helderberg Group (Becraft, Alsen, and Port Ewen Fms.) for the next 0.6 miles.

132.9	1.0	Note exposures of the Schoharie and Onondaga Fms. for the next 1.1 miles.
133.2	0.3	Exit 20 (Saugerties).
134.4	1.2	Mount Marion visible to the right of the Thruway. Note large quarry at the north end of the ridge which exposes the Berne and Otsego Mbrs. of the Mount Marion Fm.
137.5	3.1	Note exposure of Stony Hollow Mbr. on left (east) side of highway.
140.5	3.0	Note dark to rusty shales of the Berne Mbr. exposed on right (west) side of highway.
142.2	1.7	Note outcrops of Stony Hollow Mbr. for the next 0.7 miles.
142.9	0.7	Exit N.Y. State Thruway at Kingston.
143.6	0.7	Pay toll at tollbooth.
143.7	0.1	Enter traffic circle.
143.8	0.1	Turn right (northwest) to exit traffic circle onto N.Y. Rte. 28 West.
144.6	0.8	Turn right (northeast) onto Forest Hill Drive, then immediately turn right (southeast) onto City View Terrace.
144.65	0.05	Park on right (southwest) side of road.

STOP 5a. CITY VIEW TERRACE ROADCUT

This outcrop exposes the upper 10 m of the Bakoven Member and the lower 23 m of the Stony Hollow Member. The transition between the members is gradational at this locality, and it is underlain by a 1-3 m-thick zone of tectonized shale. Calcareous nodules within this interval have yielded Cabrieroceras and another unidentified goniatite (G. Kloc, 1991, personal commun.). The lower strata of the Stony Hollow Member at this locality are thinly laminated and exhibit little to no burrow-mottling (chiefly small Planolites traces). Body fossils are rare and are chiefly small tentaculids (Nowakia?) and rare small brachiopods and bivalves. The upper part of this outcrop is exposed along the adjacent road to the northwest, where thin (ca. 3-10 cm) dolomitic beds appear in the section.

144.65	0.0	Turn around and return to N.Y. Rte. 28.

144.7	0.05	Turn right (northwest) onto Rte. 28.
144.8	0.1	Stony Hollow Mbr. exposed in roadcut.
145.1	0.3	Park at the end of the roadcut. Walk back along the highway to the beginning of the outcrop.

STOP 5B. LARGE ROUTE 28 ROADCUT

This large roadcut exposes approximately 60 m of the Stony Hollow Member. The base of the outcrop lies approximately 11 m above the contact with the underlying Bakoven Member (see STOP 5a). The thinly laminated character of the unit is visible on many of the rock surfaces, and the low degree of burrow-mottling (chiefly small Chondrites burrows) is also notable. These features characterize much of the Stony Hollow Member at Kingston. A series of thin (ca. 3-17 cm-thick), buff-weathering dolomitic beds occurs 11-21 m above the base of the outcrop. Two thicker, buff-weathering units are visible farther upsection (27 and 33 m above the base of the outcrop, respectively). The lower bed (ca. 0.8 m-thick) is underlain by a layer of rusty-weathering burrows and concretions. The upper unit (ca. 0.9 m-thick) features an upper bioturbated sandstone (with Planolites and Teichichnus traces and small rugose corals) and a lower fossiliferous unit featuring hyoliths, coleolid tubes, gastropods, and cephalopods. Small bivalves may also be found in this interval. This unit is correlatable with the lower massive sandstone at Buttermilk Falls (Locality 9, Stop 4).

The next key unit is a thick, massive, light buff-weathering unit (ca. 5-6 m-thick) that is traceable northward along the outcrop belt as far as Catskill. Auloporid corals are scattered throughout the unit, and two horizons feature small streptelasmid rugose corals. The overlying interval (ca. 12 m-thick) is characterized by darker, blocky-weathering, calcareous shale-rich strata.

The top of the outcrop is marked by the resistant, thin-bedded to massive sandstone near the top of the Stony Hollow Member. The base of this interval is marked by the Lower Proetid Unit and an underlying brachiopod-rich bed. The upper surface of the sandstone marks the base of the Upper Proetid Unit at outcrops in the Kingston area.

The overlying transition into the Cherry Valley Member is not well exposed along Route 28, but a complete section through this interval may be seen 1.7 km to the east along Route 209. At the latter outcrop, the top of the thin-bedded to massive sandstone unit is overlain by a 3.5 m-thick interval of silty shales which feature a series of thin, richly fossiliferous beds with proetid trilobites. These beds may have small calcareous nodules or thin limestone lenses. This interval represents the Upper Proetid Unit in the Kingston area. An overlying thin, calcareous sandstone bed contains numerous brachiopods, which include Productella, Kayserella, and Leptaena. The brachiopod bed is capped by a prominent, 60 cm-thick, bioturbated sandstone that lies approximately 5 m below the top of the Route 209 outcrop. The two lower beds of the Cherry Valley Member are present in the upper 2.7 m of the Route 209 cut.

145.1	0.0	Continue northwest on Rte. 28.
145.2	0.1	Park on the right (northeast) side of Rte. 28 opposite a rock exposure and walk across the highway.

STOP 5C. CHERRY VALLEY MEMBER ALONG ROUTE 28 AND ADJACENT RAILROAD CUT

This roadcut and the adjacent railway cut expose a 10 m-thick section of the terrigenous sand-rich facies of the Cherry Valley Member. At this locality, the member is composed of five highly bioturbated arenaceous units separated by sandy shale intervals. These units range from 0.7-1.4 m-thick. The sandstones display prominent cleavage, especially in the railroad cut. (Note that the upper sandstone is missing in the highway cut and the lower sandstone is exposed downsection from the railway cut.) Body fossils are relatively uncommon in the sandstones, though carbonate-rich nodules feature shelly to styliolinid-rich hashes with abundant quartz sand. Leiorhyncus, other small brachiopods, and auloporid corals may be present near the tops or bases of the sandstone units. Uncommon to rare nautiloid and goniatite cephalopods may also be found. Teichichnus is the most prominent trace in the sandstones. The intervening sandy shales feature numerous limonitic burrows, accompanied by small brachiopods and bivalves.

145.2	0.0	END OF TRIP. PROCEED ON N.Y. RTE. 28 WEST TO RETURN TO ONEONTA.

A SURVEY OF PRECAMBRIAN TO PLEISTOCENE GENERAL GEOLOGY IN CENTRAL NEW YORK

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The intent of this field trip is to introduce several aspects of geology covered in the New York State earth science syllabus. We shall examine rocks that range in age from Precambrian to Pleistocene. Many rock types from a wide variety of Paleozoic depositional environments will be seen. Opportunities to collect a wide range of fossils and minerals will be provided. The field trip will develop a tectonic framework of deposition as we move higher into the stratigraphic record. Sedimentary structures and erosional features are well represented at these stops. There are several good photo opportunities of faults and potholes in addition to the rocks. We will also be seeing a wide variety of landscape features produced during the last glacial retreat, including lakes, U shaped valleys, kettle ponds, meltwater channels, sinkholes, and drumlins. An excellent general field trip of the Utica area to the west by Tewksbury and Allers (1984) is highly recommended.

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ROAD LOG

0.0	0.0	Turn left onto West Street. The campus of Hartwick College is on the right.
0.6	0.6	Turn right at the stoplight onto Chestnut Street (Rts. 7 and 23).
2.5	3.1	Turn right onto Rte. 23 N.
0.3	3.4	Junction of Routes 23 and 205. Turn right onto Rte. 205 N. You will be driving up the Otego Creek Valley, a very good example of glacial "U" shaped valley erosion.
21.9	25.3	Intersection of Routes 205 and 80. Turn right onto 80 E.
1.8	27.1	Intersection of routes 80 and 28. Turn left onto Rte. 28N. Canadarago Lake will be to the right. This lake, *a fingerlet*, was formed at the end of the last ice age by the blockage of glacial meltwaters downstream which formed a natural dam. *Fingerlet*: Muskatt's term for a smaller version of a Finger Lake.
11.2	38.3	Intersection of routes 28 and 20. Turn right onto Rte. 20 E.
0.2	38.5	Intersection of Rtes. 20 and 167 at the traffic light in Richfield Springs. Turn left onto 167 N.
9.1	47.6	Intersection of Rtes. 168 and 167. Continue N. on 167.
1.5	49.1	Outcrop of Utica Shale (Ordovician). To the east, a good meltwater channel can be viewed.
1.9	51.0	Junction with Newville Rd. at Wrights Corners. Turn left and continue to follow route 167 N.
1.4	52.4	Cross under the NYS Thruway. An outcrop of Trenton Limestone (Ordovician) containing trilobites is on the east side of the road.
0.8	53.2	Junction with Route 5S. Turn left and then make an immediate right to continue on route 167 N. The rocks outcropping to the east will be stop 2 in this road log.
2.0	55.2	City of Little Falls. Turn right onto Albany St.

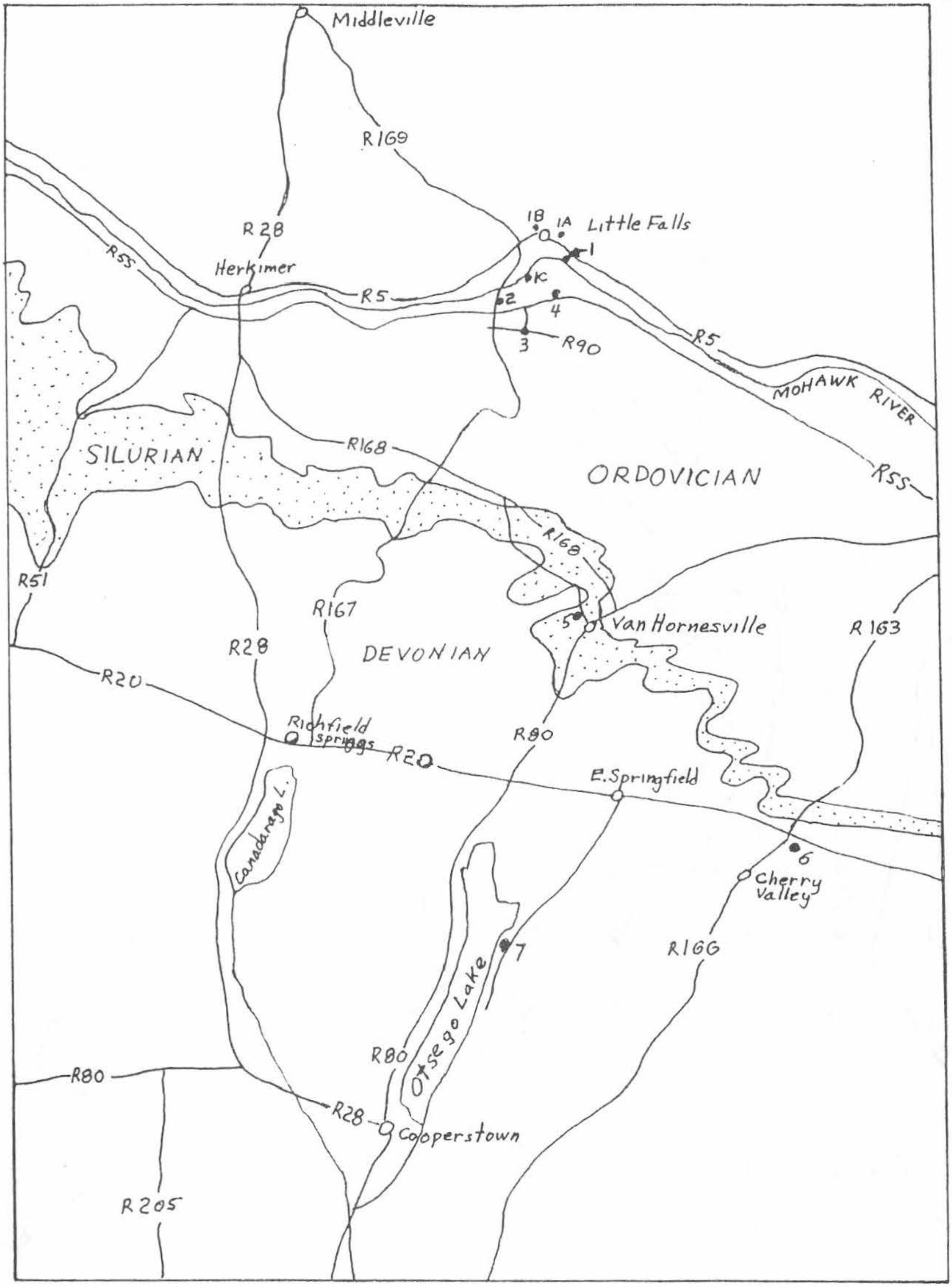

Figure 1. General Map of Field Area

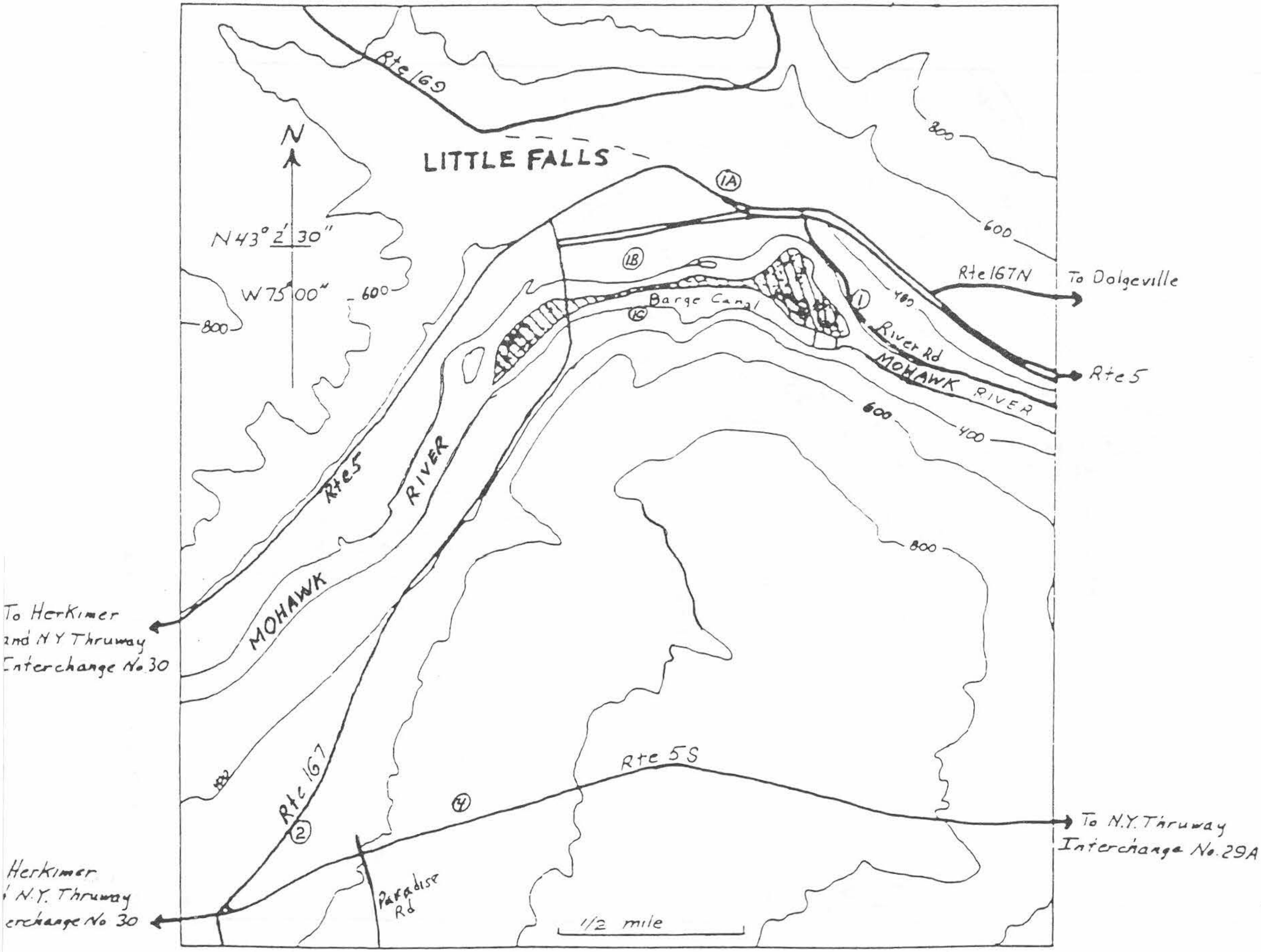

Figure 2. General Map showing Moss Island, shaded, and field trip stops

0.3 55.5 Turn right onto South Ann. St., then left onto 167 N./5 E. combination.

0.2 55.7 Turn right at the traffic light onto River Road. Precambrian basement material can be seen in an outcrop on route 5 to the east and up the hill.

0.5 56.2 Little Falls Sewage Treatment Plant.

STOP #1 - A very good outcrop of Precambrian material can be seen on the NE. side of the road just before reaching the plant. This rock is a quartzose syenite gneiss of Precambrian age. Multiple veins of quartz that have filled fractures in this material attest to a history of deformation in these rocks. Please do not use rock hammers at this outcrop, as the use of them would mar the large pothole, a very unusual erosional feature that is visible here. For some periods of the Pleistocene ice blocked the flow of water through the St. Lawrence River creating Lake Iroquois. This greatly increased the flow of water through the Mohawk R., for some 1000 yrs., from about 12,000 to 11,000 yrs ago, which carved this pothole and those across the river. Across the river is Moss Island. In May, 1976, Moss Is. was declared a National Landmark. It is one of only 400 such registered natural treasures in the U.S. The shoreline of Moss Is. displays several of these potholes, some more than 30 feet deep and about 20 feet in diameter, and makes an excellent photo opportunity. Lock 17, which is the highest lift lock on the barge canal system (40.5 ft.), can also be seen.

0.5 56.7 Intersection of River Rd. with 167/5 combination. Go straight at this light onto East Main St. (Rte. 169 N.) One block later, you will see the Board of Education building on the NE corner. Park here.

STOP #1A. In the town park ahead is a good exposure of bedrock that displays roche moutonnee glacial alteration. Turn left out of the Board of Education parking lot onto E. John St., cross E. Main St., and continue.

0.3 57.0 555 E. John St. Boulder on display in front of the Rockton Plaza was removed from a pothole during excavation.

STOP #1B. Photo opportunity of a well scoured boulder, an abrasion tool used by the swirling waters for the creation of potholes. This road bends to the right. At the next traffic light, turn left.

0.3 57.3 In front of the McDonalds store, turn left onto the turnoff for Rte. 5 and then left onto route 5. Turn right immediately to be on Route 167 S. You will be on a bridge crossing the Mohawk River. Moss Island can be viewed on the left (East).

1.0 58.3 Turn left onto Casler Road. Make an immediate left onto Flint Rd. and continue.

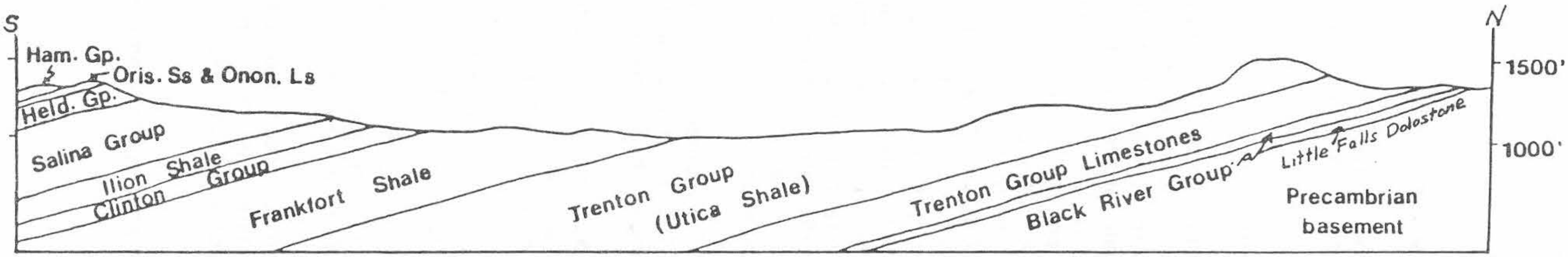

Figure 3. Schematic geologic cross section of central New York State. Average general dip is less than 1° to the South

0.3 58.6 STOP #1C Parking Lot for church on right, W. of the church. Walk to fence in the back center of lot. Across abandoned railroad right of way observe nonconformal contact of Little Falls Dolostone (Cambrian) overlying Precambrian syenite gneiss. Dolostone is overlain by classic manmade limestone wall. Return to Rte. 167.

0.3 58.9 Turn left and continue on Route 167 S.

1.8 60.7 STOP #2 Outcrop of Late Cambrian Little Falls Dolostone. This unit directly underlies the Black River Group (Ordovician) in this area. It nonconformably overlies a Precambrian syenite gneiss. The Little Falls Dolostone is a sandy, medium-grained dolostone with some sandstones occurring near the base of the unit. Except for the abundant colonial algae <u>Cryptozoan,</u> not seen at this site, the unit seems to be barren in this area. Quartz crystals from this site are locally known as Little Falls, Herkimer, or Middleville Diamonds. Anthraxolite is a black, lustrous, carbonaceous mineral often found in association with the quartz. Calcite, dolomite, pyrite, marcasite, galena, sphalerite, chalcopyrite, hematite, and glauconite have also been reported.

0.1 60.8 Intersection of Rtes. 167 and 5S. Turn left, and follow route 5S to the east.

0.4 61.2 Turn right onto Paradise Rd.

0.8 62.0 STOP #3 Paradise Road bridge over Thruway. Park at S end. Ordovician Dolgeville Facies, a mixture of calcisiltites and black shales, overlain by the Ordovician Utica Shale can be seen in rock cuts on both sides of the Thruway. It is a deeper water facies of the Trenton limestones. The contortions seen in the Dolgeville are considered to be due to a submarine slide when the unit was in a soft sediment phase. Graptolites and trilobites from the Utica Shale exposed in the ditch at the SW corner of the bridge have been found. <u>WARNING</u>: Do not climb down the cliff faces. It is illegal and dangerous to go rock collecting there. Looking at the NE. wall a very visible fault can be seen. It the first of many major faults that cross the Thruway further to the East. (This stop is a good camera opportunity).

0.8 62.8 Return to Route 5S. Turn right, and continue east.

0.4 63.2 STOP #4 **(LUNCH)**
North of the road is an abandoned quarry. This is now part of a private home, so please do not do damage to these rocks. 2-1/2 feet of Gull River Limestone underlie 18 feet of Kings Falls Limestone. These rocks are very fossiliferous, shelly and nonshelly calcarenites from the Ordovician Trenton Group. See stop #2, Cameron (1972) for more detail.

Turn around and follow Route 5S west.

0.8	64.0	Intersection of Rtes. 5S and 167. Turn left and follow 167 S.
2.6	66.6	Intersection of Rtes. 167 and 168. Turn left and follow 168 E.
1.5	68.1	Intersection of Rte. 168 and Upper Deck Rd. Turn right and continue on Deck Rd.
1.3	69.4	Intersection of Upper Deck Road and Travis Rd. Turn right on Travis Road and continue south.
4.4	74.3	STOP #5 Outcrop of Silurian Otsquago Sandstone, a formation of the Clinton Group. Coarse red sands, cross bedding well displayed. Interbedded green shales represent the Sauquoit Formation seen further to the west. A deltaic distributary channel environment of deposition is postulated for these rocks.
0.2	74.5	Intersection of Travis Rd. and Rte. 80 Turn right and follow Rte. 80 S.
5.8	80.3	Intersection of Rtes. 80 and 20. Turn left and follow Rte. 20 E.
8.2	88.5	STOP #6 Intersection of routes 20 and 166. Drive under the Route 166 bridge and park along the outcrop on Route 20. Kalkberg Limestone, Oriskany Sandstone, Esopus shale, Carlisle Center and Onondaga Formations. A 1-3 cm (Tioga) bentonite (volcanic ash) layer is found in the Kalkberg. It has been dated as being about 395 million years old. Several rock types are visible at this outcrop that represent a wide variety of Devonian depositional environments. See Fisher (1979), stop #6 for more details. Turn around, and travel W. on Rte. 20.
4.5	93.0	Junction of Route 20 and Dugway Road (County #31). Turn left and follow Dugway Road South.
4.0	97.0	Glimmerglass State Park on right.
2.4	99.4	STOP #7 Road Cut in Dugway Rd. Otsego shales (Middle Devonian). Epifaunal filter feeders were very common in this deltaic environment. The rise of highlands in the east (due to tectonic uplift) have altered the environment of deposition to something quite different from those seen at Cherry Valley. For more information, see Grasso (1977).
4.2	103.6	Junction of Dugway Rd. and River Street (Just after crossing a bridge over the Susquehana). Turn right onto River Street. A small park at the

junction with Lake Street marks the start of the Susquehana River. Here the waters of Otsego Lake flow over a Quaternary morraine that dams the lake. Turn left onto Lake Street.

0.7 104.3 Intersection of Lake Street with Rte. 80. Turn left. Continue straight on Rte. 28 S.

9.8 114.1 Intersection of Routes 28 and 7. Turn Right, taking Rte. 7 E.

6.8 120.9 Stoplight. Intersection of Main and Chestnut Streets in Oneonta. Turn right.

0.3 121.2 Stoplight. Intersection of Chestnut St. and West Ave. Turn right.

0.6 122.0 S.U.N.Y. Oneonta Campus Entrance on right. **END.**

FIELD ILLUSTRATIONS OF ROCK TYPES AND GEOLOGIC FEATURES IN THE UPPER SUSQUEHANNA VALLEY AND ADJACENT MOHAWK REGION

by

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Introduction

The outcrops and surficial features observed on this trip have been selected (1) to show introductory students the changes in lower Paleozoic stratigraphy through time (2) to illustrate several sedimentary and topographic features (3) to help students gain a better understanding of the geologic framework of this area.

The trip starts at the large Upper Devonian flood plain channel behind the F. W. Miller Science Building on the Hartwick College campus in Oneonta and ends in Precambrian garnet gneiss six miles east of Canajoharie. Progressively older beds are exposed to the north because of three factors: the gentle southerly dip of the beds, the erosion by the Mohawk River and the uplift, tilting and erosion of large fault blocks associated with normal faults in the Mohawk River Valley.

The clastic sediments are the result of the Middle Ordovician Taconian Orogeny (470-435 m.y. ago) and the Middle and Late Devonian Acadian Orogeny (385-355 m.y. ago) (Fisher, 1965). Both of these times of crustal unrest and uplift east of the Hudson River and in New England provided an influx of clays, silts and sands into the Ordovician and Devonian seas which occupied the area of this field trip. The carbonate rocks were deposited by these seas during periods of quiescence between orogenies.

Geologic History of the Area

(Modified from Fisher, 1965)

During Precambrian time a thick sequence of geosynclinal sediments was deposited. The geosyncline was folded and regionally metamorphosed into a mountain range during the Grenville Orogeny (1,100 m.y. ago). For the next 500 million years the area was eroded and the mountains were beveled down exposing the metamorphic rocks in their roots.

By Late Cambrian time these Precambrian metamorphic rocks were covered by transgressing shallow seas which deposited the Little Falls sandy dolostones and dolostones forming nonconformity. The Precambrian gneiss and Late Cambrian dolostone are exposed in a railroad cut six miles east of Canajoharie along the Mohawk River.

Shallow seas continued to deposit dolomitic rocks into Early Ordovician time. The Chuctanunda Creek Dolostone exposed in the gorge of Canajoharie Creek represents deposition at this time. After deposition of this dolostone the seas withdrew.

In Middle Ordovician time shallow seas again covered the area and deposited the Kings Falls and Sugar River argillaceous limestones on top of the Lower Ordovician Chuctanunda Creek Dolostone forming a disconformity. The thin beds of black shale in the limestones and the dark color of these argillaceous limestones reflect crustal unrest many miles to the east in the area of the present Taconic Mountains. This was the beginning pulse of the Taconic Orogeny. A thick black shale, the Canajoharie Shale, which overlies the limestones represents increased unrest during the Taconic Orogeny. These Early and Middle Ordovician sediments are well exposed in the gorge of Canajoharie Creek.

During Late Middle Ordovician time there was extensive erosion of the Taconic mountains. The detritus was deposited to the west as a thick sequence of shales and sandstones which are exposed in a few scattered outcrops between Sharon Springs and Canajoharie. The best exposure is just north of Sharon Springs.

Throughout most of Silurian time this area was emergent. If there are any Silurian rocks present, they are not exposed along the field trip route.

In Early Devonian time shallow seas encroached into the area and deposited a thick sequence of limestones (Helderberg Group) which are exposed north of Cherry Valley along Route 20 east to the vicinity of Sharon Springs.

Later in Early Devonian time there was uplift and erosion which resulted in the deposition of the Esopus Shale and Carlisle calcareous siltstone. This uplift was followed by another period of submergence when the Onondaga Limestone was deposited. These three formations are exposed on Route 166 north of Cherry Valley .25 mile south of Route 20.

During Middle and Late Devonian time the Acadian Orogeny was taking place in New England. This mountain building episode provided a vast supply of sediments which formed the thick sequence of sandstone, siltstone and shale which are exposed between Cherry Valley and Oneonta. These sediments were deposited as part of the extensive Catskill delta and flood plain deposits.

Since Late Devonian time the area has been subjected to erosion and the development of the Mohawk River, the Susquehanna River and their tributaries. In the Pleistocene, glaciers moved into the area. The ice enlarged the river valleys and deposited morainic material in the valleys. Some of this material was reworked by later advances of the ice to form drumlins or was redeposited by meltwater to form kame terraces along the valley walls. The area is currently being drained by the Mohawk River which joins the Hudson River north of Albany and the Susquehanna River and its tributaries which flows south through Pennsylvania and enters the Atlantic Ocean in Chesapeake Bay.

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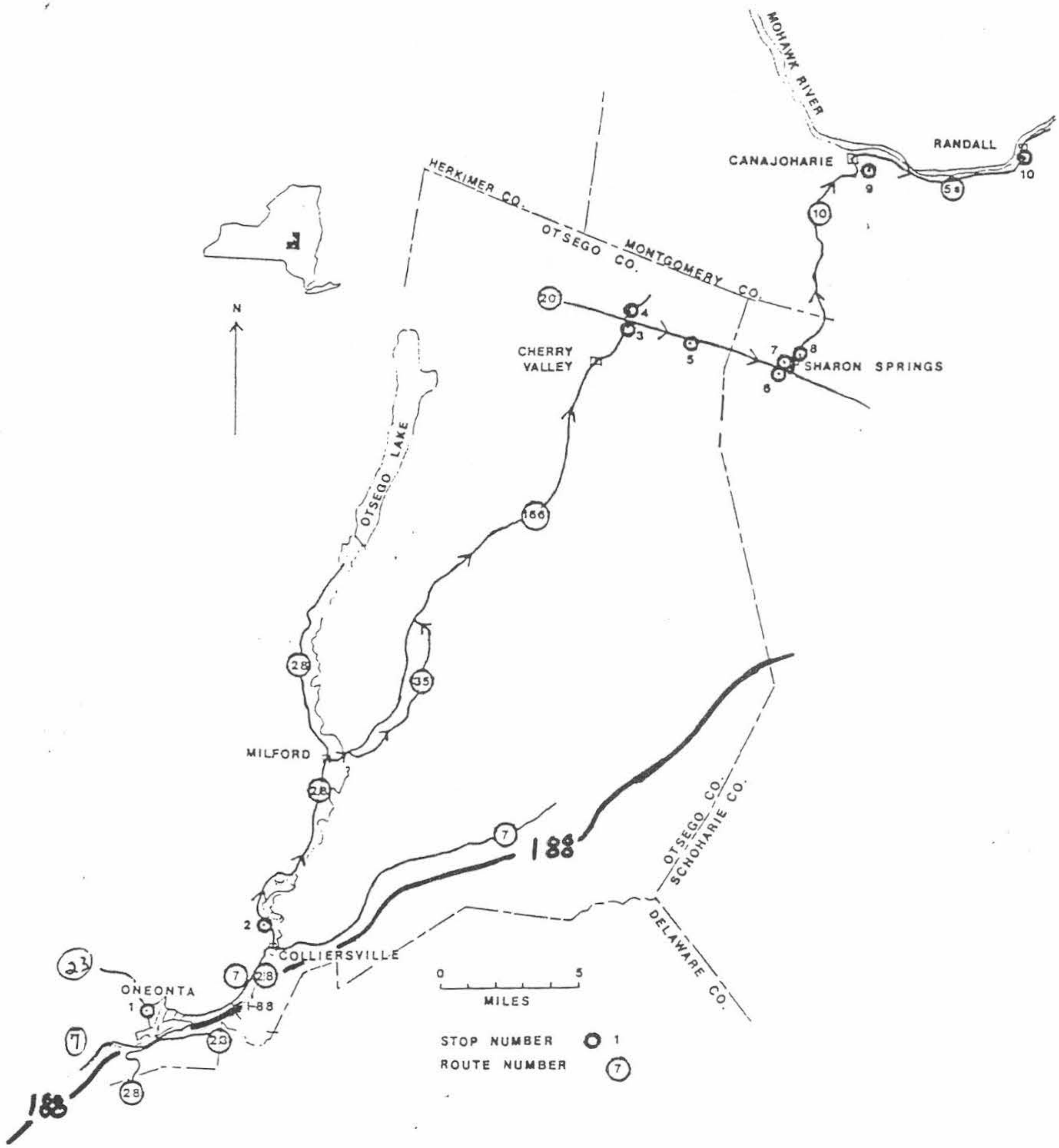
Johnson, K. G., 1970: in Heaslip, W. G. editor, Guidebook to Field Trips of the New York State Geological Association 42nd Annual Meeting, C-1 - C-14.

STRATIGRAPHIC SECTION OF FORMATIONS ON TRIP

(adapted from Rickard and Zenger, 1964 and Fisher, 1965)

	Thickness	
Upper Devonian (Senecan Series)		
Oneonta Formation	200'	Stop 1
Gilboa Formation	460'	Stop 2
Middle Devonian (Erian Series)		
Cooperstown Shale	410'	Drove by (after stop 2)
Portland Point Limestone	5-6'	Not observed
Panther Mountain Formation	800'	Not observed
Solsville Sandstone	290'	Not observed
Otsego Shale	260'	Not observed
Chittenango Shale	150'	Stop 5
Cherry Valley Limestone	5'	Stop 5
Union Springs Shale	25'	Stop 5
Onondaga Limestone	120'	Stop 3
Lower Devonian (Ulsterian Series)		
Rickard Hill Limestone (Schoharie)	0-1'	? Stop 3 ?
- ? disconformity ? -		
Carlisle Center Shale	10-40'	Stop 3
- ? disconformity ? -		
Escopus Shale	0-20'	Stop 3
- ? disconformity ? -		
Oriskany Sandstone	0-2'	Not observed
- unconformity -		
Lower Devonian (Helderbergian Series)		
Kalkberg Limestone	15-50'	Stop 6
Coeymans Limestone	90-100'	Stop 7
Manlius Limestone (lower Thacher member)	30-40'	Stop 4
Upper Silurian (Cayugan Series)		
Cobleskill Limestone	10-12'	Not observed
- ? disconformity ? -		
Brayman Shale	100-200'	Not observed
Vernon Shale	0-80'	Not observed
- unconformity -		
Middle Silurian (Niagaran Series)		
Herkimer Sandstone	0-40'	Not observed
Kirkland Hematite	0-2'	Not observed
- ? disconformity ? -		

Willowvale Shale	0-30′	Not observed
- ? disconformity ? -		
Sauquoit Formation	0-130′	Not observed
Oneida Conglomerate	0-15′	Not observed
- unconformity -		
Middle Ordovician (Mohawkian Series)		
Frankfort Shale	500-800′	Just past stop 8
Canajoharie Shale	200′ (?)	Stop 9
Sugar River Limestone	15′	Stop 9
Kings Falls Limestone		
- unconformity -		
Lower Ordovician (Canadian Series)		
Chuctanunda Creek Dolostone	20′	Stop 9
Tribes Hill Limestone	100′	Not observed
Upper Cambrian (Croixian Series)		
Little Falls Dolostone	500′	Stop 10
- unconformity -		
Precambrian gneisses	?	Stop 10


MOHAWK RIVER
RANDALL
CANAJOHARIE
HERKIMER CO.
OTSEGO CO.
MONTGOMERY CO.
N
CHERRY VALLEY
SHARON SPRINGS
OTSEGO LAKE
MILFORD
188
OTSEGO CO.
SCHOHARIE CO.
DELAWARE CO.
COLLIERSVILLE
ONEONTA
I-88
0
5
MILES
STOP NUMBER
ROUTE NUMBER

ROAD LOG

GEOLOGIC SETTING OF UPPER SUSQUEHANNA AND ADJACENT MOHAWK REGION

Cumulative Mileage	Miles from Last Point	Route Description
0.0	0.0	ASSEMBLY POINT: Hunt Union Parking Lot at State University College at Oneonta (SUCO).
		Departure: At 8:00 a.m.
		Proceed southwest down Ravine Parkway to the main entrance of SUCO campus at West Street.
.7	.7	TURN LEFT (SOUTH) ON WEST ST. until you come to the entrance of Hartwick College.
.9	.2	TURN RIGHT (WEST) INTO THE HARTWICK COLLEGE CAMPUS at Hartwick Drive, proceed up the hill, veer left at the chapel house and continue going through the large parking lot to the outcrop behind the F. W. Miller Science Bldg.
1.3	.4	STOP 1 Upper Devonian Oneonta Formation

This outcrop consists of dark grey shales, thinly bedded siltstones and sandstones. A massive sandstone which fills in part of a Devonian stream channel is well exposed for about 150 feet at the left end of the outcrop. There is an erosion surface between the underlying shale and the overlying massive sandstone. This surface which marks the bottom of the channel rises stratigraphically to the right. Interference ripple marks are exposed just below the road level. Plant materials are abundant in some layers. A few galena crystals about 1 mm across have been found in the thinly bedded sandstone.

Trilobites, marine pelecypods and brachiopods of the Gilboa Formation have been found in beds stratigraphically 75 feet below this outcrop, and marine fossils are found in thinly bedded siltstones and shales stratigraphically 200 feet below stop 1. This outcrop is next to Nick's Diner on Chestnut Street (Route 23). We probably will not stop here today.

A clean sandstone which shows extensive trough-type cross bedding crops out in the woods between Stop 1 and Stop 1A.

STOP 1A Red mudstones and shales with root casts, worm burrows and ripple marks are exposed at the top of the hill just below the Ernest B. Wright Observatory which houses a 16" telescope. Walk or drive (if the sliding chain link fence gate is open) up the hill at the far end (west) of the Binder Physical Education Building parking lot. Continue up the hill through the second gate to the outcrop below the Observatory next to the practice soccer field.

The view down the valley is looking southeast to Mt. Utsayantha in Stamford (about 30 miles map distance). The broad U-shaped valley is the result of Pleistocene glaciation. The Susquehanna River flows in from the north (out of your view) and through the city of Oneonta. The terraces on either side of the valley are kame terraces. Red beds are exposed near the top of the hills across the valley.

LEAVE STOP 1A AND GO BACK DOWN HARTWICK DRIVE TO WEST STREET AT THE MAIN ENTRANCE TO HARTWICK COLLEGE.

1.7	.4	TURN RIGHT (SOUTH) ON WEST STREET and continue going down hill.
1.9	.2	TURN LEFT (EAST) ONTO CENTER STREET which is immediately past the Lutheran Church. (This is called "crash corner" for obvious reasons.)
2.5	.6	CONTINUE ON CENTER STREET FOR .6 MILE TO WALLING AVE. TURN RIGHT (SOUTH). Walling Ave. is the first street to

		the right immediately after you cross Oneonta Creek at the entrance to the park.
2.7	.2	TURN LEFT (EAST) AT FRIENDLY ICE CREAM. You are now on Main Street and Route 7 & 28. (Stay on this road for 4.8 miles to Colliersville.)
2.8	.1	Fox Hospital on your right.
4.0	1.2	Small kettle hole on right. This has been partially filled in to make a parking lot for Pyramid Mall.
4.4	.4	View to right showing kame terraces across valley.
5.0	.6	Access to I 88 (continue going straight).
6.2	1.2	On right (South) across valley gravel pit in delta kame.
7.6	1.4	TURN LEFT (NORTH) ONTO ROUTE 28 IN COLLIERSVILLE AT JUNCTION OF ROUTE 28 AND 7.
8.1	.5	Stop sign; turn left (North).
8.5	.9	STOP 2 Stop directly across from dam at Goodyear Lake. There is pull-over on right (East) side of road.
		The Upper Devonian Gilboa Formation with horizontal sandstones and siltstones is exposed here. The well developed flowrolls are of interest. It is apparent that these are primary structures (formed while the sediment was still "soft"), rather than secondary structures (formed after lithification), but there is some question as to their origin. Some beds contain marine brachiopods, bryozoans, crinoids and pelecypods.
13.4	4.9	Outcrop of Middle Devonian Cooperstown Shale at curve in road. Good view ahead of broad U-shaped valley. The Susquehanna River flows south from its headwaters in Otsego Lake (Cooperstown) in this valley.

15.9	2.5	<u>TURN RIGHT (EAST) IN MILFORD ONTO ROUTE 166</u>.
16.7	.8	Cross Susquehanna River. Clays indicate a glacial lake occupied this area. If you want to study the clays you should go past junction with County Road 35 and park in small area to the right along Route 166. Then to continue road log, turn around and get on County 35 going south.
17.0	.3	<u>TURN RIGHT (SOUTH) ONTO OTSEGO COUNTY ROAD 35</u> and take metal bridge over Cherry Valley Creek.
17.2	.2	Bear left at junction.
18.9	1.7	Meander scar of Cherry Valley Creek is visible to left.
21.0	2.1	Westville Cemetery on left.
21.9	.9	<u>TURN LEFT (WEST) AND CROSS CHERRY VALLEY CREEK</u>. After crossing the creek, NOTE THE EXCELLENT hummocky morainic deposits showing numerous kettle holes and knob and kettle topography.
22.9	1.0	<u>TURN RIGHT (NORTH) AND GO NORTH ONTO ROUTE 166</u>.
31.6	8.7	Town of Roseboom.
35.9	4.3	Cherry Valley - The town was settled about 1740. On November 11, 1778 over 40 people were killed by Tories and Native Americans in the infamous Cherry Valley Massacre.
37.7	1.3	<u>STOP 3</u> (Pull off road to left and park in Highway Department gravel and salt storage area.) Topographically this spot is a break in the Helderberg escarpment. Fairchild interpreted this as a glacial spillway formed at a time when ice filled most of the Mohawk Valley. Meltwater was blocked from draining north and "spilled over" the escarpment eroding the valley (Fairchild, 1925). Fleisher (personal communication) feels that the valley is the result of glacial scouring of

a through valley.

Three formations are exposed here. The lowest formation (exposed about .1 mile down the road) is the Esopus Shale. This is overlain by the Carlisle Center calcareous siltstone which contains numerous worm burrows _Toanurus cauda-galli_ (Rickard and Zenger, 1964). About .1 mile south along the road the Carlisle Center siltstone is overlain by the Middle Devonian Onondaga Limestone which contains abundant crinoids, corals and brachiopods.

The upper part of the Onondaga contains abundant chert. Jointing is very obvious at this outcrop.

LEAVE STOP 3 AND CONTINUE GOING STRAIGHT AHEAD AND GO UNDER ROUTE 20.

38.6 .9 STOP 4 The Lower Devonian Manlius Formation (lower Thacher Member) crops out on the right. This laminated micrite contains some ostracods, tentaculitids and stromatoporids. Some mud crack marks are present. This is the intertidal facies of LaPorte, 1967. The Coeymans Formation rests on top of the Manlius Formation.

TURN AROUND AND GO SOUTH TO ROUTE 20 EAST.

39.4 .8 GO EAST ON ROUTE 20.

42.1 2.7 TURN RIGHT OFF OF ROUTE 20 ONTO BLACKTOP ROAD AND PROCEED TO THE OUTCROP VISIBLE TO THE RIGHT.

42.3 .2 STOP 5 Three Middle Devonian formations are exposed here. The lower most formation is the Union Springs Shale which is a black fissile shale containing calcareous concretions. There is a thin limestone near the top of the shale. This is overlain by the Cherry Valley Limestone which is about 7 feet thick and contains a cephalopod fauna. More than 100 feet of the jet-black Fissile Chittenango Shale rests on top of the Cherry Valley Limestone. At

		the east end (left) of the outcrop the Union Springs Shale has been broken up and sheared indicating some minor faulting.
42.5	.2	*RETURN TO ROUTE 20 AND CONTINUE GOING EAST.*
44.3	1.8	Schoharie County Line
46.1	1.8	*STOP 6* Pull off the road and stop at down-going slop of hill. The Lower Devonian Kalkberg Limestone is exposed in a fresh outcrop. This is a medium grained thin to medium bedded limestone with abundant chert. There are numerous brachiopods, bryozoans, some corals and trilobite fragments. A 2" thick layer of bentonite is exposed near the eastern end of the outcrop. At the east end of grassy field below outcrop there is a small creek that disappears into a sink hole in the limestones.
46.5	.4	*IN SHARON SPRINGS TURN LEFT (NORTH) AT STOPLIGHT ONTO ROUTE 10 AND PROCEED NORTH FOR .1 MILE TO OLD QUARRY NEXT TO BOWLING ALLEY.*
46.6	.1	*STOP 7* (in old quarry) The lower Devonian Coeymans Formation consists of a coarse grained, thickly bedded limestone with abundant brachiopods, crinoids and corals.
		LEAVE THE QUARRY AND PROCEED DOWN THE HILL.
47.2	.6	*STOP 8* (STOP AT PARK NEXT TO OLD BATHS) Stop to look at springs and tufa deposits in the city park at the north end of the village of Sharon Springs. There is a strong odor of H2S from the spring water. This is probably caused by the water passing through the underlying thick black shales.
47.2	.2	Upper Middle Ordovician Shales and Sandstones exposed in cliff at left.
50.4	3.0	Two drumlins are in view to the left.
57.1	6.7	*IN CANAJOHARIE TURN RIGHT AT THE FIRST*

		<u>STOPLIGHT ONTO MONTGOMERY STREET.</u>
57.2	.1	<u>CROSS OVER CANAJOHARIE CREEK AND TURN RIGHT ONTO MOYER STREET. CONTINUE ON MOYER STREET.</u>
57.5	.3	<u>TURN RIGHT ONTO FLORAL AVENUE AND PROCEED TO THE TURN-AROUND AT END OF ROAD.</u>
57.7	.2	<u>STOP 9</u> Four Lower and Middle Ordovician formations are exposed in Canajoharie gorge. The lower most formation is the Chuctanunda Creek Dolostone. This is unfossiliferous except for the "hippopotami backs" which are dolomitized hemispherical stromatolites (algal mounds) (Park and Fisher, 1969). Large potholes have formed in the dolostone (Canajoharie is the Iroquois name for the "Pot that washes itself") (Park and Fisher, 1969). The Middle Ordovician Kings Falls and Sugar River black limestones overlay the dolostone, forming a disconformity. These limestones and the thin black shales in them contain abundant trilobite fragments, bryozoans, brachiopods and crinoids. The limestones are overlain by more than 100 feet of Middle Ordovician Canajoharie Shale.
57.9	.2	<u>RETURN TO JUNCTION OF FLORAL AVENUE AND MOYER STREETS</u>. <u>TURN LEFT AND GO DOWN HILL</u>.
58.2	.3	<u>CROSS MONTGOMERY STREET AND GO ONTO MITCHELL STREET</u> (WHICH IS 30 FEET LEFT AND PARALLEL TO CANAJOHARIE CREEK).
58.3	.1	<u>CROSS RAILROAD TRACKS AND TURN RIGHT ONTO ROUTE 5S AT THE BEECHNUT FACTORY</u>. Continue on Route 5S. En route to the next stop note the cliffs of Upper Cambrian Little Falls Dolostone.
64.8	6.5	<u>STOP 10</u> Rusty weathering Precambrian garnet gneiss exposed along the south

side of the road and in railroad cut.

<u>FOLLOW THE FOOT PATH AT THE EAST END OF THE OUTCROP TO THE RAILROAD CUT. BEWARE OF POISON IVY ALONG PATH AND AT RAILROAD CUT</u>.

The folded Precambrian gneiss is overlain by the Upper Cambrian Little Falls Dolostone. The dolostone is brecciated for several feet above the contact with the gneiss. Apparently there was movement along the unconformity during Ordovician time when the normal faults in the Adirondack Mountains and the Mohawk River Valley were formed. There are also good exposures of the gneiss north of this outcrop across the Mohawk River along route 5 (5.4 miles east of Palatine Bridge).

<u>PROCEED EAST TO</u> OBSERVE THE FAULT-LINE SCARP OF NOSES FAULT FROM THE VEHICLES.

65.3 .5 There is a good view of the fault-line scarp on the overpass of Route 5S over the railroad tracks. Note that west of the Noses fault-line scarp the Mohawk River Valley has steep cliffs of Little Falls Dolostone. These are caused by the down cutting of the Mohawk River on the upthrown western side of the fault. East of the fault-line scarp on the downthrown side of the fault-line scarp there are no cliffs.

65.8 .5 END OF TRIP AT RANDALL.

Note: Time may not permit us to stop at all of the stops. Depending upon the interest of the group, certain stops may be omitted.

ENVIRONMENTAL GEOLOGY AND HYDROLOGY OF ONEONTA AREA

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Cumulative Mileage	Miles from Last Point	Route Description
0.0	0.0	Morris Hall, SUCO Campus. Exit area for Ravine Parkway.
0.4	0.4	Turn RIGHT at stop sign onto Ravine Parkway and drive through campus.
0.8	0.4	Bear RIGHT at fork.
1.0	0.2	Straight through traffic light.
1.5	0.5	Turn LEFT at stop sign onto East Street.
3.1	1.6	Bear LEFT at fork onto Wilber Lake Road.
4.5	1.4	Wilber Lake Dam on right.
4.8	0.3	Turn RIGHT onto dirt road.
5.0	0.2	Park on right. Walk to lake.

STOP #1: WILBER LAKE RESERVOIR
(refer to Figure 1 locator map)

Wilber Lake serves as the principal water supply for the city of Oneonta and adjacent suburban areas. The lake itself covers an area of approximately 90 acres when full and receives runoff and infiltration from a drainage basin of only about 1650 acres. Relief within this drainage basin is 393 feet, with the highest elevation being 1928 feet. Nearly 80% of the drainage basin is forested mainly with mature hardwoods and pine. Some selective lumbering has occurred over the years, but clear-cutting of trees has not, and lumbering has apparently had minimal effect on the drainage characteristics within the basin. The remaining area is comprised of fields, pastures and open bushy areas. The city owns or controls most areas adjacent to the reservoir and no development is allowed. Surface deposits around the lake consist of a silty clay glacial till matrix with shale and siltstone slabs in a layer ranging from about seven to twelve feet thick, decreasing in thickness on the surrounding hillsides to a little over five feet (Walker, 1985).

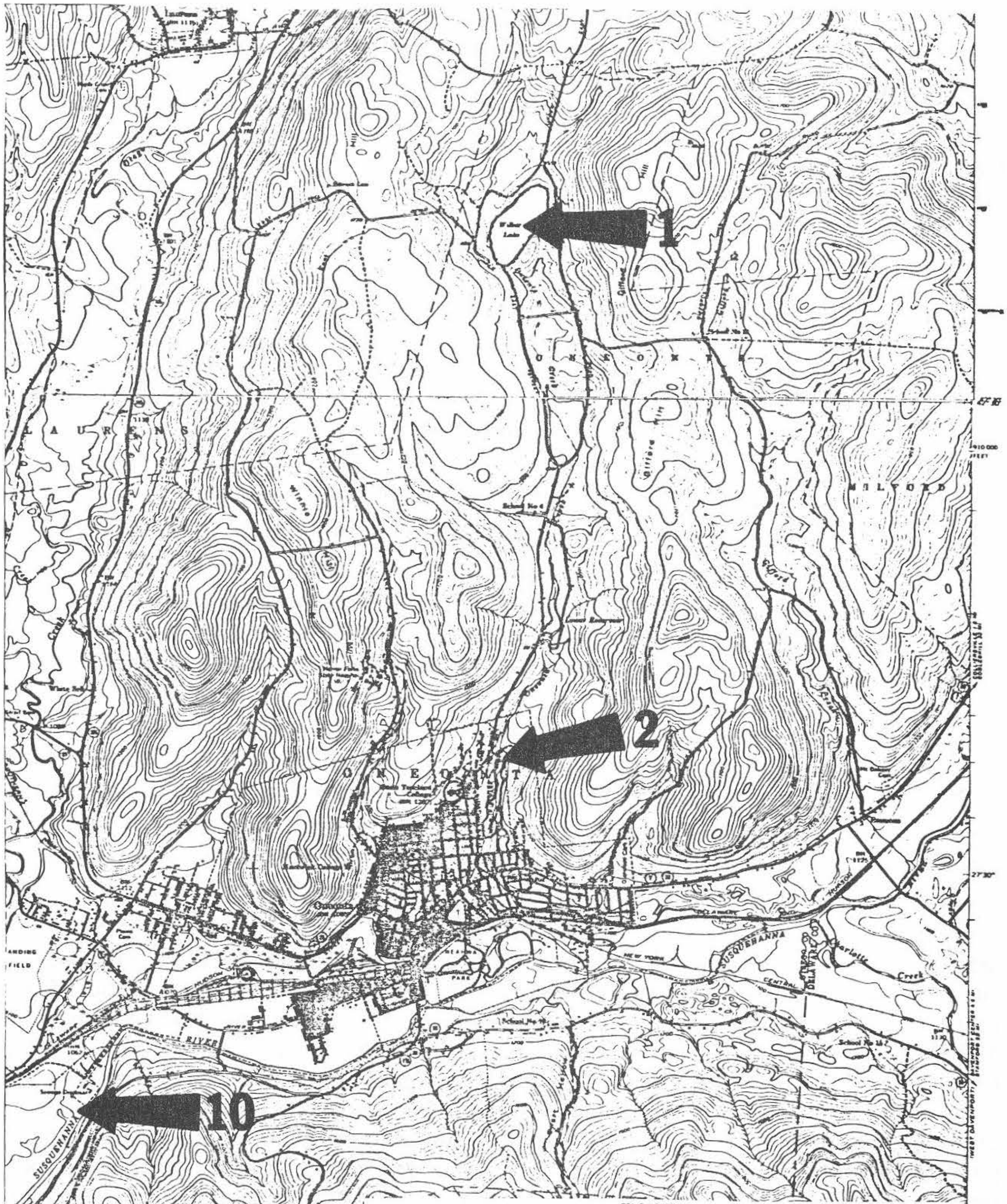

Figure 1: Locator Map

The lake contains approximately 450 million gallons of accessible water, with another 70 million gallons contained below the water supply intake pipes. According to water treatment plant personnel, the present water use is highly variable, but typically averages about 1.5 million gallons per day (roughly 100 gallons/person/day), or approximately 550 million gallons per year. This requires more than one complete recharge of the lake every year, which only occurred for five years between 1958 and 1982 (Walker, 1985). These years were fortunately when water demands were less than present, but increasing demands since that period have necessitated the search for supplementary sources of water. The largest single users of the city's water supply are the colleges (Hartwick College and the State University College) and water usage varies greatly depending on whether or not one or both colleges are in session since the student population is greater than one-third of the total city population.

The lower reservoir holds an estimated 25 million gallons maximum (probably considerably less due to sedimentation). The water treatment plant will not use water from this source deeper than about two or three feet because of the amount of suspended sediment in the water column. Even with this additional supply, it would only equal a few days supply at the present demand. Supplementary sources include a well drilled west of the city in 1941 for industrial uses but was never used. It is now on property owned by the New York State Electric and Gas Company and was leased to the city in 1982. This well has been connected to the city's water main and can provide up to about one million gallons per day (1.5 cubic feet/second) for four months (Arthur Palmer, pers. comm.), but is only used at about half that capacity. This supply is only used when necessary due to the rust scale in the pipes. Water must be pumped back against the normal flow direction in the mains, which increases the rust scale and discoloration problem. Also, any treatment (such as chlorination) must be done at the well head. Another well drilled in Wilber Park can provide up to about 250 thousand gallons per day. This well is near the water treatment plant and is connected to the water supply pipe to the plant, but is presently not in service. A last resort supplementary supply is possible from the Susquehanna River. Problems with this source are discussed later, as this is one of the stops (#3) on this trip.

		Return back on dirt road to Wilber Lake Road.
5.1	0.1	Turn LEFT at stop sign onto Wilber Lake Road and return to city.
6.7	1.6	Bear RIGHT at yield sign.
8.4	1.7	City limit. Continue straight ahead on East Street.
8.8	0.3	Water treatment plant on left.

STOP #2: WATER TREATMENT PLANT
(refer to map in Figure 2)

The city's water treatment plant, located on East Street near the high school, treats an approximate average of 1.5 million gallons of water per day from Wilber Lake with chlorine at the rate of 12-18 pounds/day, the rate dependent upon water demand. Also, the supplementary water supply from the West End well is treated daily with 4 to 5 gallons of sodium chloride solution at the well head when this well is being utilized.

		Continue on East Street.
9.4	0.6	Turn RIGHT at stop sign onto Center Street.
9.5	0.1	Turn LEFT at light onto Maple Street.
9.7	0.2	Turn RIGHT at set of traffic lights onto Main Street. Take immediate LEFT onto Grand Street.
9.8	0.1	Turn RIGHT onto Market Street.
10.0	0.2	Turn LEFT at bottom of grade onto Gas Avenue. Cross RR tracks and bridge.
10.1	0.1	Turn LEFT across bridge, following sign Catella Park and Mill Race Walk.
10.3	0.2	Park on right just before overpass.

STOP #3: MILL RACE PUMPING STATION
(refer to Figure 3)

The mill race, originally constructed to divert water from the Susquehanna River to provide water power to mills located downstream, now serves as a flood control channel and as a possible supplementary source of water for the city. The small brick building contains a pump which can divert water from the intake immediately above the small dam on the mill race to the reservoir via the pipe visible in the streambed which enters the mill race at the dam. However, this water source is highly undesirable and is viable only in an emergency. There is a high algae content which results in discoloration and an undesirable taste. The water may be polluted with other contaminants as well. Perhaps more significantly, during a drought, when supplementary water supplies would be most needed, the river level would be too low to provide water to this channel. Also, during winter months, it is likely that the water in the above ground pipe from the pumping station to the reservoir would freeze.

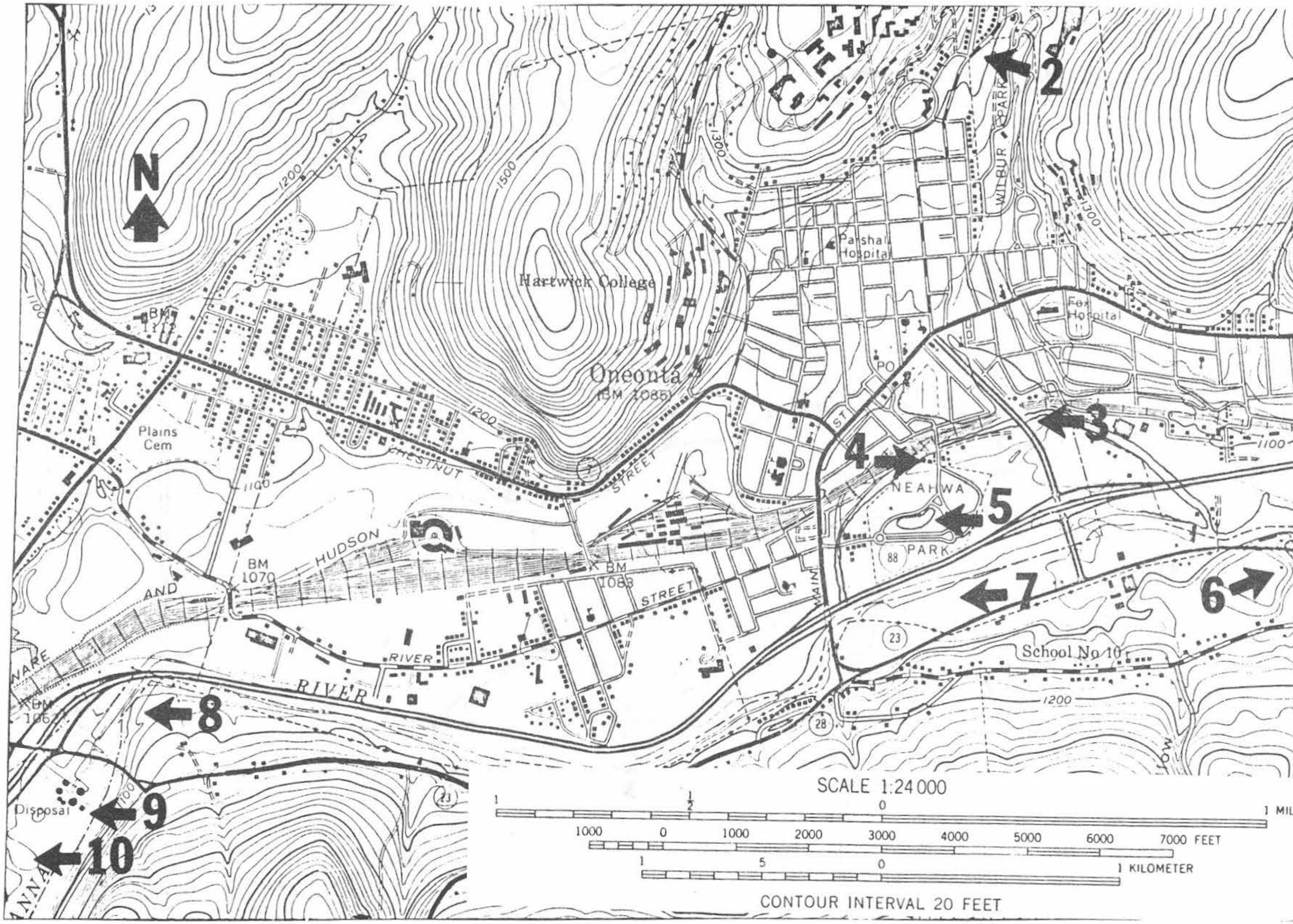

Figure 2: Detail Map of Stops 2-10

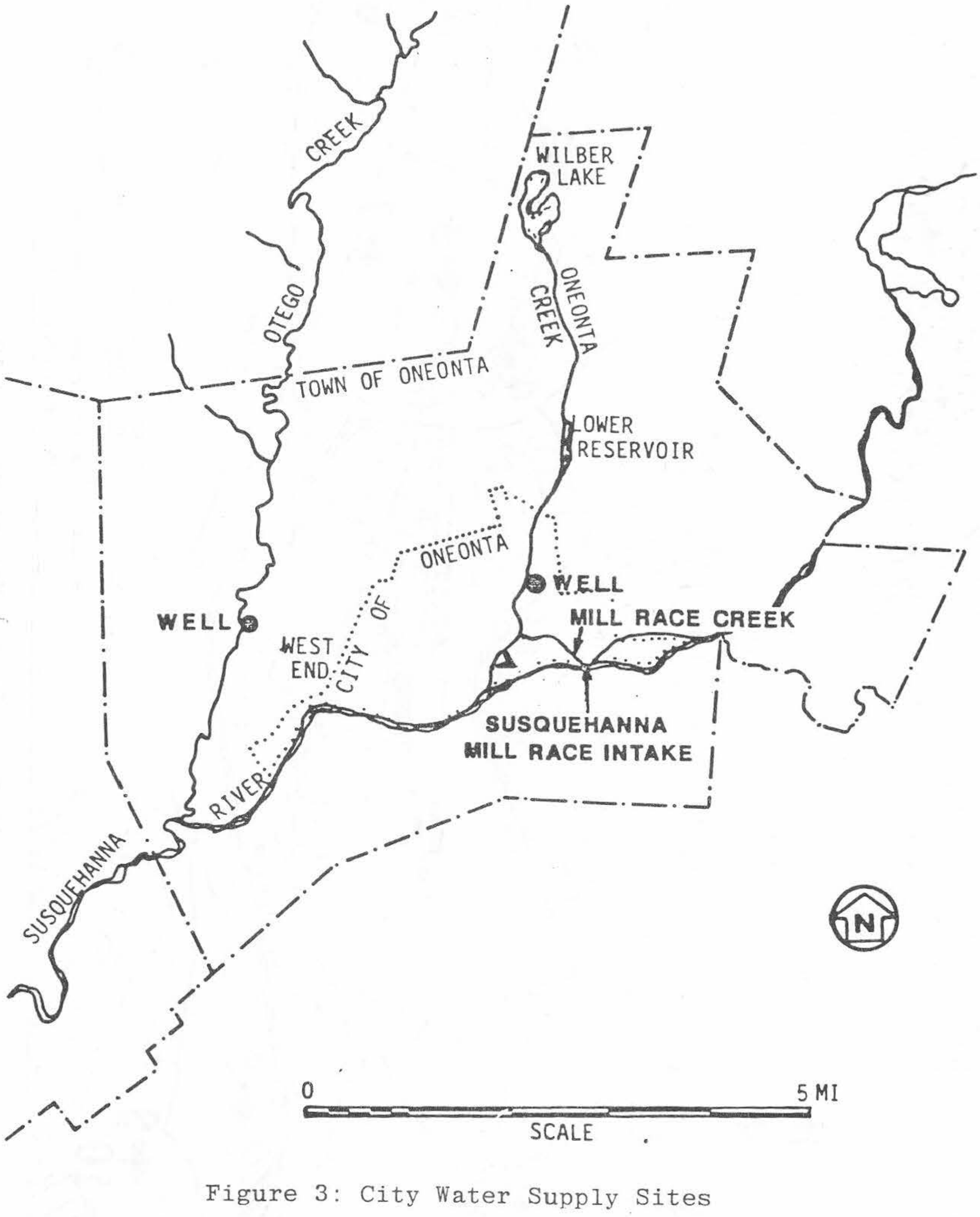

Figure 3: City Water Supply Sites

Return to Gas Avenue.

10.4	0.1	Turn LEFT at stop sign onto Gas Avenue.
10.5	0.1	Park near bleachers at ball park.

STOP #4: <u>NEAHWA PARK-BURIED HAZARDOUS WASTE SITE</u>
(refer to Figure 4)

From 1881 until 1953, a coal gasification plant operated in the general area now occupied by the baseball field and adjacent buildings along Gas Avenue in what is now part of Neahwa Park. The plant was originally constructed in 1881 by the Oneonta Gas Light Company, which merged with the Oneonta Electric Power and Light Company in 1887. In 1918, it became part of the Ithaca Gas and Electric Company, which itself became part of New York State Electric and Gas Company (NYSEG). With the advent of cheaper petroleum, natural gas and electric power, the plant closed in 1953 after 72 years of operation. Most of the buildings were dismantled in 1956 and the remaining buildings and property were sold to the City of Oneonta in 1966. The site has since been used for the city garage and storage of city highway equipment, the dog pound, a road salt storage facility, and presently as a minor league baseball field. While in operation, this plant produced gas for heating and lighting from coal, as well as many by-products which are now known to be hazardous (Figure 5). Some of these by-products were buried in the immediate vicinity and were not discovered until recent excavations to repair the Gas Avenue bridge were begun.

NYSEG hired professional consultants to analyze the soils and groundwater in the area, including those from several drilled test wells and excavated pits, in order to determine any present or potential hazard. The initial site survey was initiated in April 1986 and the final technical report was published in February 1990. Analyses of soil from the test pits indicated the presence of the following material (TRC Environmental Consultants, 1988):

A. Aromatic hydrocarbons

benzene
chlorobenzene
1-2-dichlorobenzene
1-4-dichlorobenzene
ethylbenzene
toluene

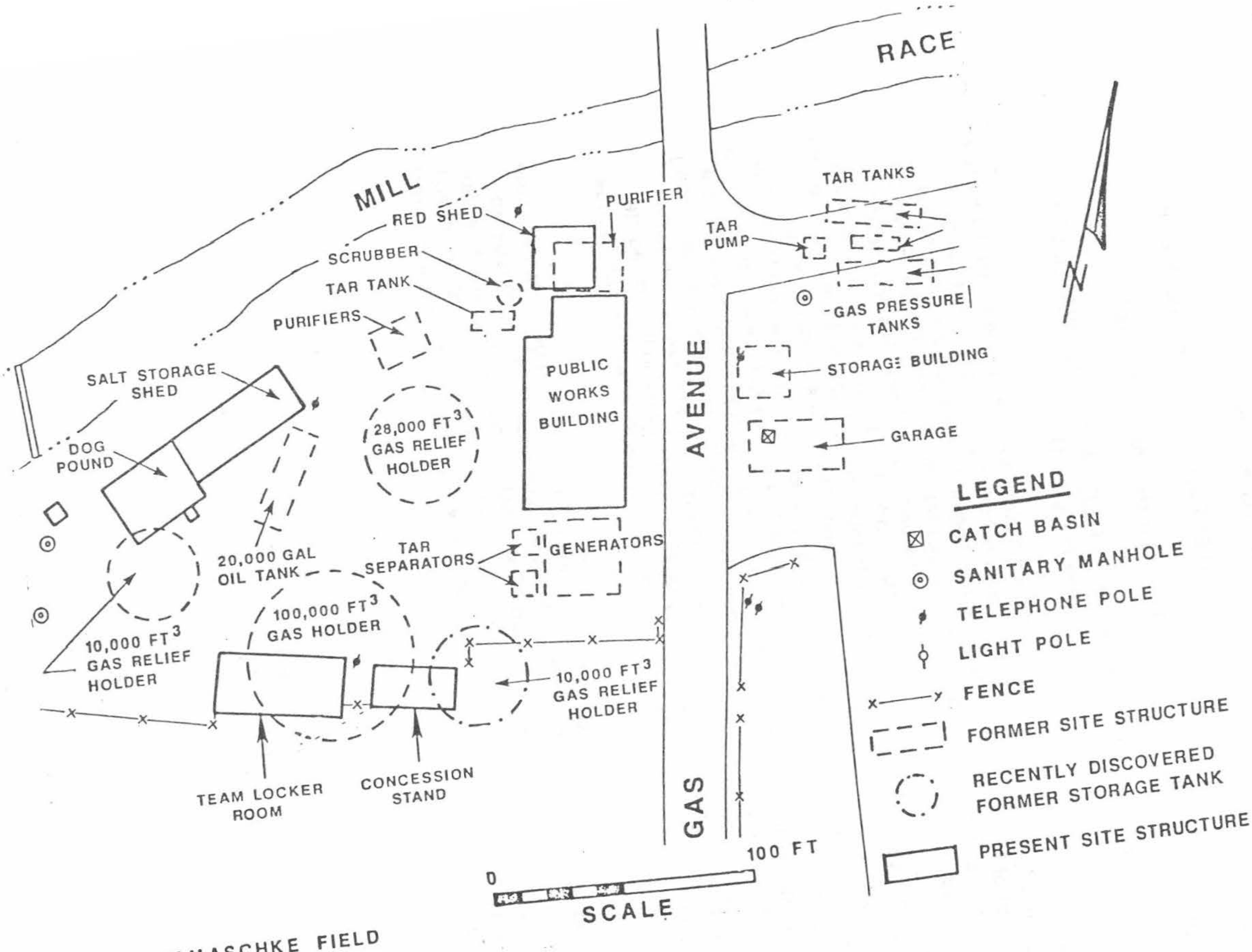

Figure 4: Neahwa Park

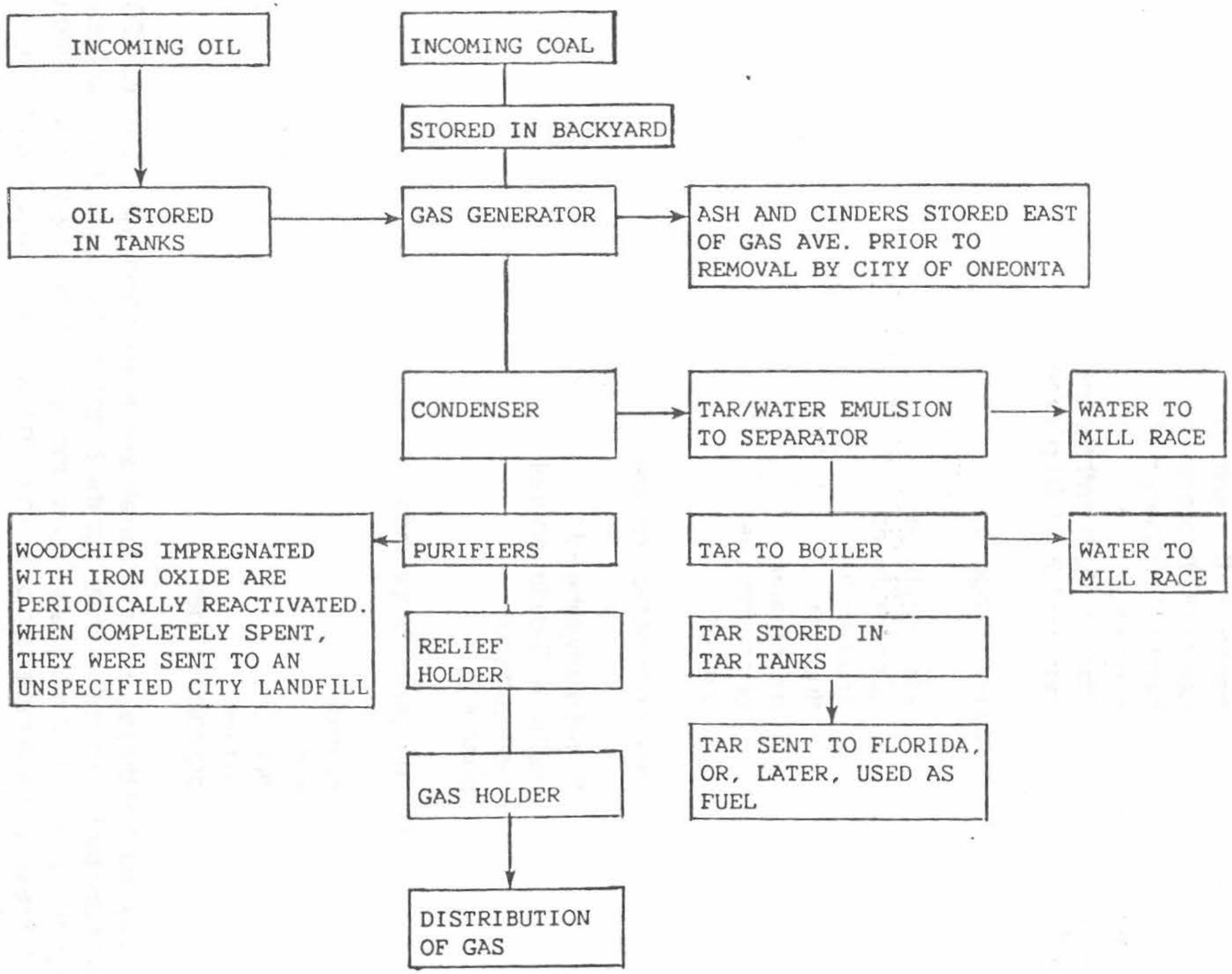

Figure 5: Flow Chart for Gasification Process

B. Polynuclear aromatic hydrocarbons (PAHs)

1. Carcinogens

benzo (A) anthracene
benzo (A) pyrene
benzo (B) fluoranthene
benzo (K) fluoranthene
benzo (GHI) perylene
chrysene
dibenzo (A,H) anthracene
indeno (1-2-3 CD) pyrene

2. Non-carcinogens

acenaphthene
acenaphthylene
anthracene
fluorene
naphthalene
phenanthrene
pyrene

C. Non-chlorinated phenols

2, 4-dimethylphenol
methyl-4, 6-dinitrophenol
4-nitrophenol
phenol

D. Inorganic compounds

cyanide
iron
zinc
sulfate
organic nitrogen

It is assumed that these various materials are derived from the estimated 220,000 gallons of dry tar buried in this area over the history of the gasification plant operation. Although the above list of materials may sound devastating, it was determined that although present, these materials occur in such low concentrations as to be considered safe in their present location deep beneath the surface (TRC Environmental Consultants, 1990). In fact, it was determined that excavation and attempted removal of these materials may present a greater risk while simultaneously presenting the problem of what to do with the material once it is removed. It is doubtful that all of the materials could be retrieved in any case. Groundwater movements are being monitored and excavations in the area are prohibited to prevent this material being exposed. The site has been taken off the hazardous site list by the New York State Department of Environmental Conservation and any remediation will come under the control of DEC's Water Division (rather than the Hazardous Waste Division). However, controversy between City officials and the DEC officials continues. One problem includes the fact that State standards are less strict than Federal standards and if Federal standards were followed, the site would still be on the hazardous site list and may therefore qualify for Superfund cleanup. However, DEC officials state that Superfunds were used only when a site had been abandoned or the responsible party was bankrupt or uncooperative. This is not the case and NYSEG and DEC officials plan to work together to develop a remediation plan if it is determined to be necessary.

Walk to:

STOP #5: HODGES POND

One of the main attractions within the city's Neahwa Park is Hodges Pond. Although swimming is not allowed, it is stocked with small fish (bluegills, etc.) and becomes an ice-skating rink in the winter months. However, recent years have shown dense weed growth in the pond and it was becoming less aesthetically enjoyable as a result. Annual weed removal was expensive and was a temporary solution at best. It was decided in 1990 to drain the pond and dredge much of the bottom mud. Also, to control weed growth without resorting to herbicides, one plan is to lower the water levels in the fall and allow freezing winter temperatures to kill the roots of the shallow water weeds.

Another potential problem is that groundwater movement in this area is generally from the area of the coal tar contamination (STOP #4) toward Hodges Pond. The potential effects of this on the pond are presently unknown.

		Return in vehicle to Market Street.
10.7	0.2	Turn RIGHT at stop sign onto Market Street.
10.9	0.2	Turn LEFT at stop sign onto Grand Street.
11.0	0.1	Turn RIGHT at stop sign onto Main Street. Stay in far right lane and enter Lettis Highway at intersection.
11.6	0.6	Continue straight through next two sets of overhead traffic lights at overpass.
11.8	0.2	Turn LEFT at traffic light by McDonalds onto Route 23.

12.0	0.2	Turn RIGHT into shopping mall and turn left to the far side of the mall parking lot near the hill.

STOP #6: <u>SOUTHSIDE MALL - Planned Mining</u>

The hill of glacial gravels at the east side of the shopping mall is owned by a local developer who plans to remove the hill and use the material to fill in wetlands along the Susquehanna River about one-half mile to the west (STOP #7). The advantages include allowing more room for expansion of the mall and/or other forms of commercial development as well as providing more potential development space in the area to be filled. However, local environmental groups (City Environmental Board, Audubon Society, and others) and private citizens who realize their responsibility for the stewardship of the environment are concerned about the effect of such massive changes to the local area. Questions have been raised not so much in regards to the removal of the gravel hill but more in terms of the effect of filling in the present floodplain/wetland area along the river. These concerns will be explained at our next stop, which is at the planned fill site. Other potential problems concerning the removal of the hillside gravel include the traffic congestion and the impact on the road surface itself from the hundreds of truckloads of gravel that will have to be hauled continuously over several months. As of this writing, the planned operations are "on hold" until all required permits are obtained by the developer from appropriate State offices (Department of Environmental Conservation, Department of Highways, etc.).

		Exit shopping mall and turn LEFT.
12.8	0.8	Continue straight through traffic light near McDonalds.
13.0	0.2	Park any place on right across from Christopher's Restaurant.

STOP #7: <u>SOUTHSIDE WETLANDS - Planned Filling</u>

Most of the area between the highway and the river would be filled with the gravel from the hill adjacent to the shopping mall (STOP #6). As can be seen, most of the trees have already been cut and the developer is apparently only waiting for final approval from State agencies.

One of the principal environmental concerns regarding this operation would be the possible increased potential for flooding. The Federal Emergency Management Agency (FEMA) includes this area within the 100-year floodplain of the river, defined as being at the 1090 foot elevation. Without this area available to help contain floodwaters, it only means that other areas will suffer increased flooding, including city residential and commercial areas. Also, increased stream flow may result in increased erosion downstream. Of particular concern would be the possible erosion of the stream bed adjacent to bridge foundations. Besides the obvious destruction of the wetland habitat, groundwater properties, water tables and drainage characteristics may be adversely affected. A main point of contention is the fact that the developer has not mentioned what his ultimate plans are for this area once it is filled and ready for development. Without knowing what kind of commercial development is planned for this area makes it difficult for even non-environmentally concerned citizens to decide what is best for this area. Most people appreciate the financial benefits of commercial development in their area, as long as they have input on controlling what type of development is planned.

		Continue straight ahead on Route 23.
13.5	0.5	Continue straight through light.
14.2	0.7	Turn RIGHT onto paved road.
15.5	1.3	Turn RIGHT onto paved road.
15.7	0.2	Turn RIGHT into Oneonta fishing access site.
15.9	0.2	Park and walk to river's edge.

STOP #8: RIVER ACCESS - Erosion Control

This river access site is at the downstream end of a section of the Susquehanna River which was straightened in order to allow the construction of the adjacent section of Interstate-88 to proceed without the added costs of building bridges over the original river meanders. This resulted in the river flowing in a nearly straight line for over a mile. Since rivers naturally develop meanders, especially when flowing over land with minimal gradient, man has had to prevent meander development by lining the river banks with large boulders ("rip-rap"). This also tends to increase downcutting in the restricted channel since flow velocity will increase during increased discharge if it is not allowed to widen its channel. A quick glance downstream should point out the potential hazards of increased erosion caused by natural flooding or from removal of upstream wetland areas which can store floodwaters for gradual release. One of the bridge foundation supports is located in mid-stream and may be weakened by erosion.

Notice also how the flow velocity differs across the river at a meander. This should help explain why there are no boulders to prevent erosion on the inside bend of a meander. Rather, sediment is deposited in these areas of lower flow velocity.

16.1	0.2	Exit and drive straight across highway to Oneonta Municipal Complex.

STOP #9: WASTEWATER TREATMENT PLANT

The Oneonta wastewater treatment plant is classified as a secondary treatment process which involves biological treatment to break down the organic materials followed by chlorination to eliminate potentially harmful bacteria. The average daily volume of water treated in 1990 was 3 million gallons but the plant has a maximum capacity of about 3.6 million gallons daily. According to the manager of the plant, one reason for the higher volume of waste water treated than the volume of reservoir water entering the city supply mains is infiltration of groundwater into some of the older permeable sewer pipes. Also, there are some homes in the town that may be connected to the sewer lines but may have their own private wells rather than being connected to the city water supply.

Continue along paved road to:

16.4 0.3

STOP #10: TRANSFER/RECYCLING STATION

The transfer/recycling center was opened in 1988 when recycling became mandatory in the City of Oneonta, one of the first cities in this area to do so. Recyclables are separated by the generator (individuals, businesses, etc.) and either picked up by commercial haulers during the regular garbage pick-up schedule or brought to the station by individuals. Present materials that are removed from the waste stream and recycled include newspapers, certain types of plastic, glass (clear, green and brown are separated), corrugated cardboard, cans, tires, car batteries, used motor oil, and green waste (grass clippings, leaves, branches, Christmas trees, etc.). Newspapers, cardboard, cans and plastic containers are compacted into bales for shipment to available markets for reuse. Glass bottles are crushed before shipment in order to reduce the volume. The remaining garbage is compacted and transferred by truck to a landfill in Montgomery County.

The facility was taken over by the Montgomery - Otsego - Schoharie Authority (MOSA), which controls garbage collection and recycling in the three county area. A program to begin recycling glossy paper, magazines, junk mail, and similar material is planned to begin this fall. Future plans include the mixing of green waste with sewage sludge to produce a composted material which could be used by farmers or other citizens to enrich their fields and gardens. In many larger metropolitan areas, the use of sewage sludge in this way is not feasible due to the high levels of industrial heavy metals or chemicals it contains. It is also hoped that large-scale recycling of other plastics (such as styrofoam and brittle plastics) will be feasible with further advances in recycling technology and the subsequent development of markets for this material.

END OF TRIP.

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FRESHWATER CARBONATE FROM THE UPPER DEVONIAN CATSKILL MAGNAFACIES, DAVENPORT CENTER, CENTRAL NEW YORK

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ABSTRACT

Non-soil carbonates are rare in fluvial deposits of the Upper Devonian Catskill Magnafacies although they are common in other fluvial rocks. A 0.5 m thick dolomitic mudstone bed occurs near the base of the Oneonta Formation at Davenport Center, NY. This laminated mudstone layer contains filament molds, rare ostracode shells, calcispheres and a variety of burrows and mudcracks. Associated strata include sandstones interpreted as channel deposits and interbedded sandstones and mudstones interpreted as crevasse-splay, levee and flood basin deposits. We interpret the carbonate bed as a freshwater-marsh/shallow-lake deposit analogous to the "periphyton" lakes and marshes of the Everglades and interior Andros Island. This marsh/lake deposit developed during local reduction of siliciclastic input. The purpose of this trip is to examine the carbonate bed and associated strata, to compare the sequence with similar coeval sequences in the region, and to discuss our interpretations.

INTRODUCTION

Carbonates are common in many modern and ancient fluvial deposits where they occur both as early diagenetic soil features and as the deposits of a variety of freshwater lakes and swamps. Descriptions of freshwater, non-pedogenic carbonates from ancient alluvial deposits include: Beerbower (1961), Belt and others (1967), Belt (1968), Friend and Moody-Stuart (1970), Berryhill and others (1971), Freytet (1973), Leeder (1974), Ryder and others (1976), Beaumont (1979), Ferm and Horne (1979), Flores (1981), Ordonez and Garcia del Cura (1983), and Anderton (1985). However, only a limited number of thin (< 0.5 m) non-pedogenic carbonate mudstone beds are known from the fluvial part of the Devonian clastic wedge in central New York (Johnson and Friedman 1969, p 461; Demicco and others 1987; Bridge and Willis 1991). Most of these are found in cores drilled by New York State Power Authority in the vicinity of Gilboa, New York, but a few can be found in outcrops of siliciclastic mudstones deposited near the paleoshoreline (Bridge and Willis 1991).

The carbonate mudstone that is the subject of this trip is the exception in that it is apparently associated with wholly non-marine rocks (Demicco and others 1987). It is also the thickest (approximately 0.5 m thick) and best exposed carbonate in the area. The carbonate bed occurs in an abandoned quarry about 1 km south of Davenport Center in south-central New York (Fig. 1). We reckon the quarry is near the base of the Frasnian Oneonta Formation (Fig. 2), but the precise stratigraphic position of this quarry is hard to define because of the scale of geologic maps of this region (Fisher and others 1970) and because the rock types exposed are not unique to the Oneonta Formation. The purposes of this trip are to: (1) describe the sedimentary features of the carbonate bed and associated strata; (2) present our interpretations of the depositional environments; and (3) discuss the implications of the carbonate bed for sedimentation rates and paleoclimate.

DESCRIPTION AND INTERPRETATION OF ASSOCIATED STRATA

The quarry wall is about 130 m long, 20 m high and oriented east-west. Figure 3 is a vertical measured section from the thickest central portion of the quarry. The section comprises a basal 4 meter thick sandstone and an upper 13 m or so of interbedded sandstone and mudstone.

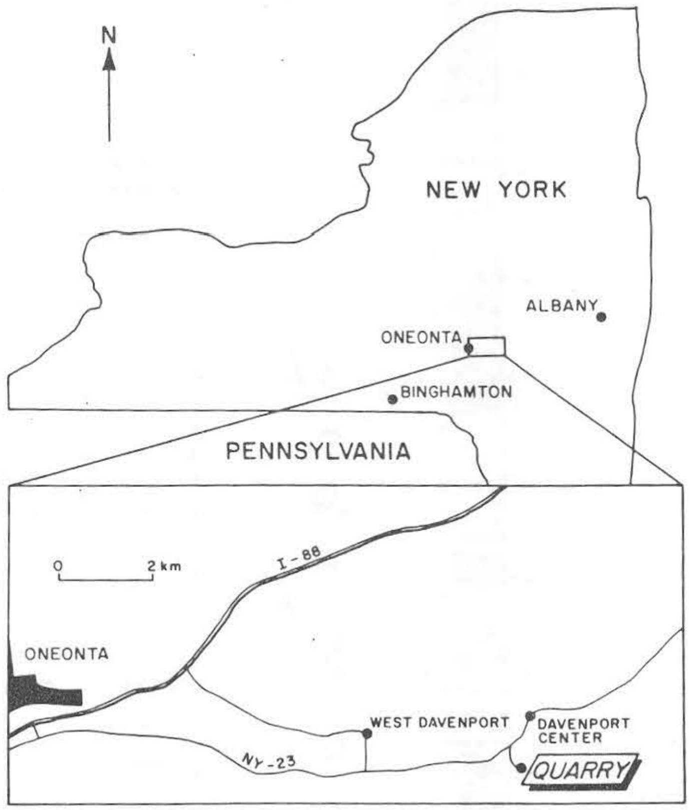


Figure 1. Location Map.

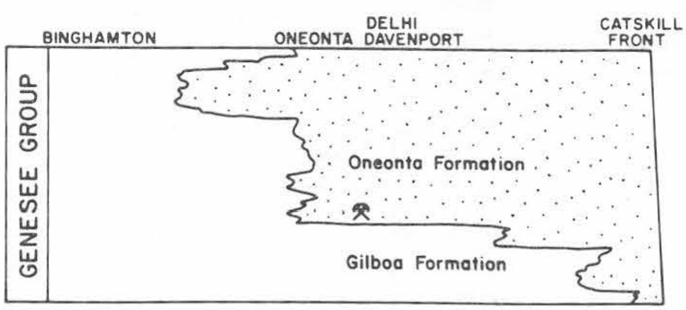


Figure 2. Stratigraphic chart (from Rickard 1975). Quarry marked by symbol. Catskill Magnafacies (alluvial) is stippled. Genesee Group is about 450 m thick in this region (Rickard 1975).

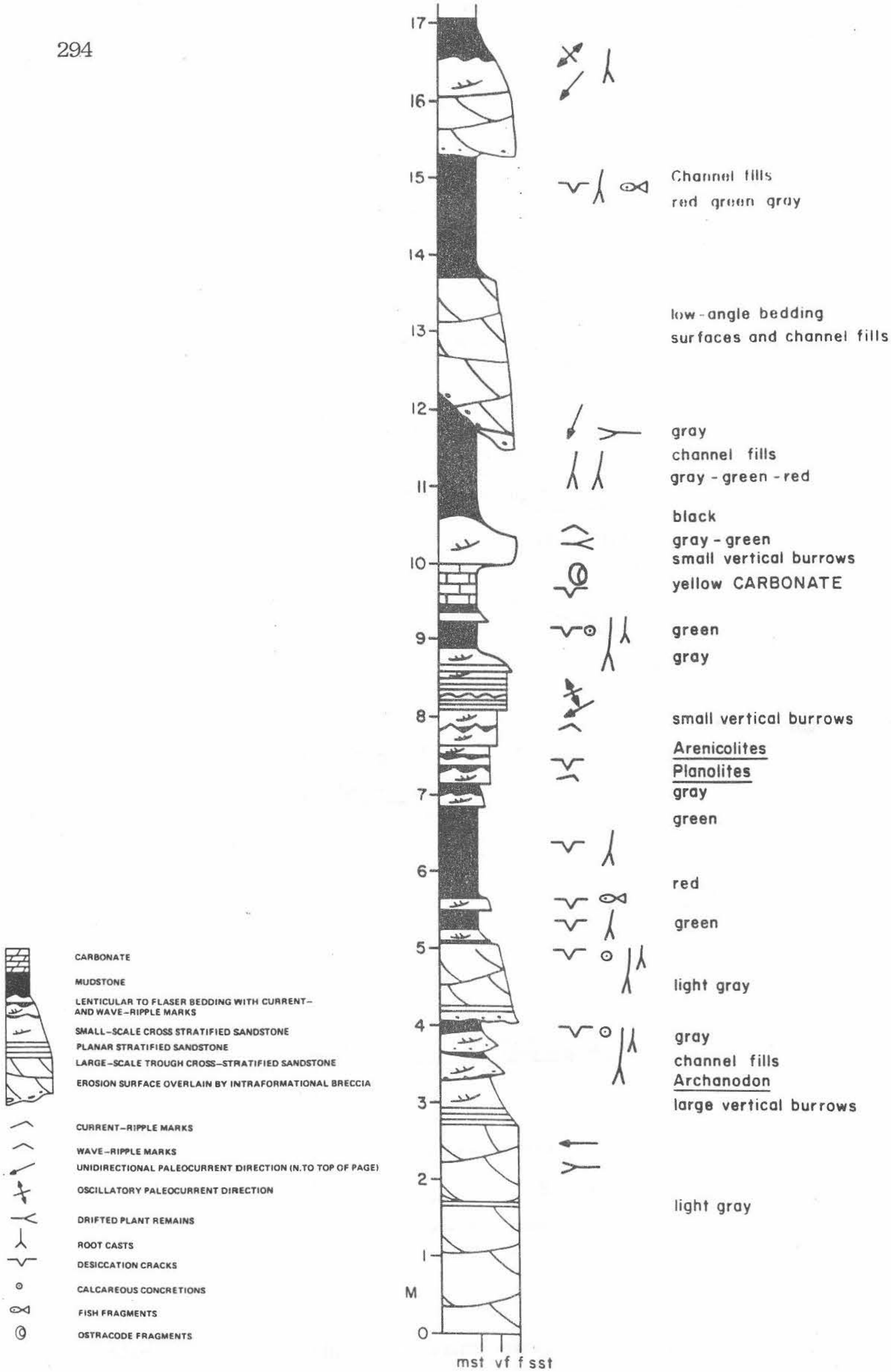


Figure 3. Measured section from the central portion of the quarry. Scale is height above quarry floor in meters. Refer to explanation of symbols. Other sedimentary features noted to right of graphic log.

Description of the basal sandstone: The basal 3 m of the quarry is composed of gray, fine-grained, large-scale cross-stratified sandstone that grades up to very fine-grained, planar-stratified and small-scale cross-stratified sandstone in the uppermost 0.5 m. Overlying this are decimeter-thick bedsets which fine upward from small-scale cross-stratified very fine-grained sandstone to gray mudstone. These bedsets fill meters-wide channels and contain mudcracks, root casts, calcareous concretions, large vertical burrows, and rare impressions of shells of the bivalve *Archanodon*.

Interpretation of the basal sandstone: The thick sandstone at the base of the section is interpreted as a major channel deposit, with the overlying minor channel fills representing chute-channel or crevasse-channel fills. This interpretation is based on its thickness and sedimentary features compared to better exposed deposits described by Bridge and Gordon (1985a, b).

Description of the interbedded sandstones and mudstones: The remainder of the section containing the carbonate bed is composed of decimeter- to meter-thick beds of fine- to very fine-grained sandstone interbedded with red, green and gray mudstone in beds up to a meter thick. *Sandstones* have erosional bases, are generally sheet like, and most fine upward. Thicker sandstones are large-scale cross-stratified whereas small-scale cross-stratification is more common in thinner sandstones. Wave- and current-ripple marks occur on top of some sandstone beds. Unidirectional currents are to the southwest whereas there is no systematic orientation of wave ripple marks. The uppermost 0.5 m of some thick sandstone beds are disrupted by root casts, desiccation cracks, calcareous concretions, and burrows. Drifted plants remains and fish fragments also occur. The thick sandstone bed about 12 m above the quarry floor contains low-angle bedding surfaces that define broad, shallow channel and bar forms. At about 8 m a distinctive, 2 m thick *coarsening-upward bedset* occurs. Mudcracked, wave-ripple cross-stratified and small-scale cross-stratified heterolithic beds (lenticular -> wavy -> flaser beds) grade up into planar-stratified and wave-ripple cross-stratified very-fine grained sandstone. A variety of burrows (including U-shaped *Arenicolites*, horizontal *Planolites*, and small vertical burrows) disrupt the bedset. Wave ripple marks are common on bedding surfaces here. Green mudstone overlies the coarsening-upward bedset, and is sharply overlain by the carbonate bed. The mudstone contains brown carbonate concretions, desiccation cracks, and root casts that penetrate the underlying sandstone. Other *mudstone* beds contain abundant desiccation cracks, burrows, rare fish fragments, and root casts up to 0.5 m long that are rarely encased in micritic carbonate. Channel fills up to 1.1 m deep occur in the mudstone above the carbonate bed.

Interpretation of the interbedded sandstones and mudstones: The interbedded sandstones and mudstones are interpreted as overbank flood deposits associated with crevasse splays, levees, flood-basins and lakes. Bridge and Gordon (1985a, b) described better-exposed examples from nearby roadcuts in the Oneonta Formation. *Sandstone* sheets are generally interpreted as the bedload deposits of individual sheet floods over flat flood-basin surfaces whereas the *mudstone* layers are interpreted as suspended load deposits. The erosional bases and fining-upward character of the sandstones record initial erosion of the flood-basin surface followed by deposition from waning flows. The disrupted upper portions of sandstones and the general disruption of mudstones record colonization by plants, burrowing by organisms, desiccation and calcareous paleosol development. Fine-grained channel fills in the upper mudstones may be flood-basin drainage channels. The sandstone 12 m above the quarry floor is interpreted as a crevasse-splay deposit based on its overall wedge-shaped geometry and internal channel and bar bedding structure. The *coarsening-upward bedset* at 8 m in the section is reckoned to record the progradation of a levee or crevasse splay into a flood-basin area. The progradation is initially recorded by the heterolithic strata which indicate periodic bedload deposition followed by suspended load deposition in ponded water, culminating in desiccation. The planar stratified sandstone in the upper portions of the coarsening-upwards bedset record upper plane bed deposition on the levee or crevasse-splay proper. This levee or crevasse splay may have built into a perennial lake. A lake is inferred from the common vortex ripples and extent and nature of the burrowing in this interval. The abrupt fining at the top of this bedset may be due to avulsion of the main river channel and abandonment of the levee or crevasse splay system (e.g. Bridge 1984).

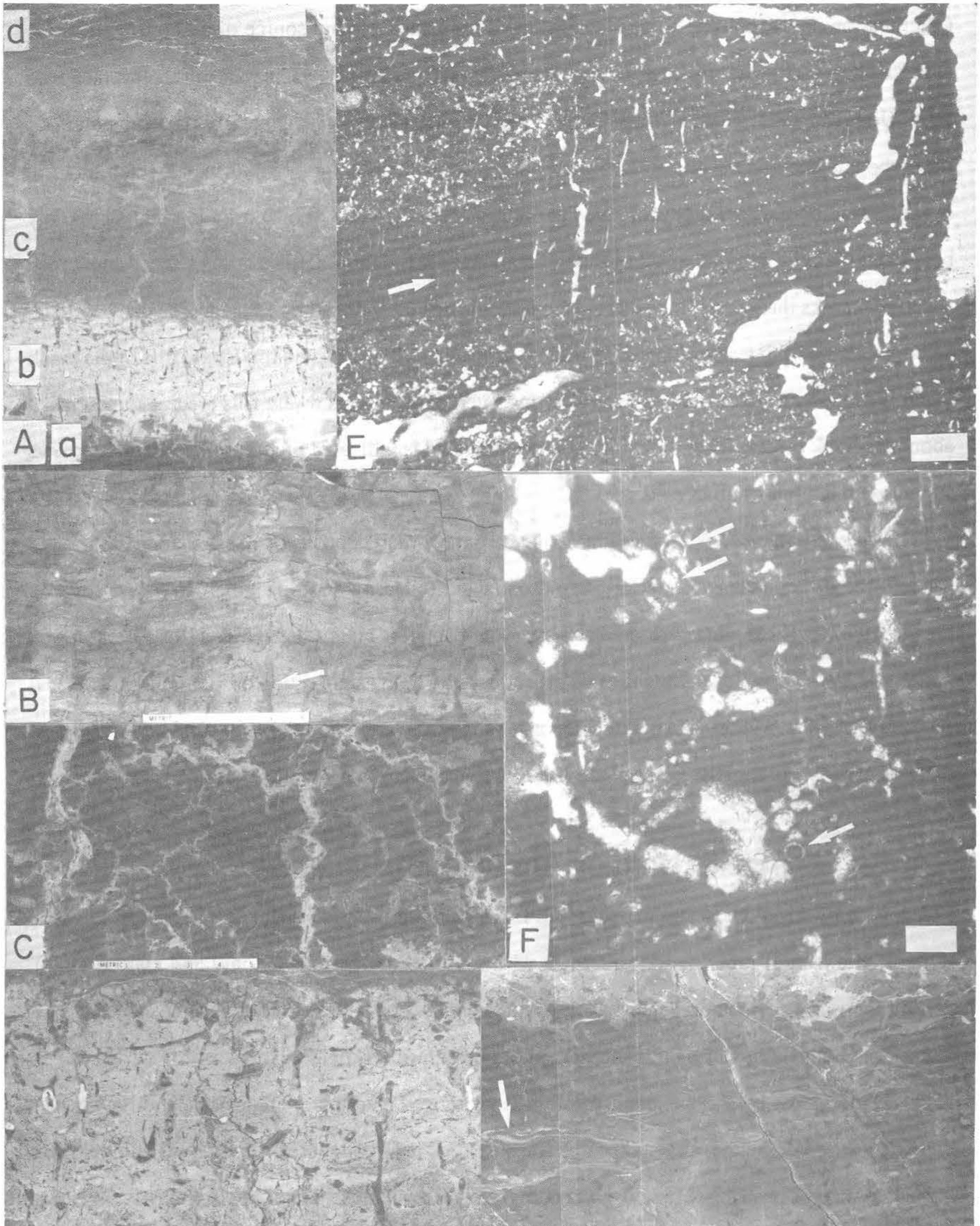
DESCRIPTION AND INTERPRETATION OF THE CARBONATE BED

The carbonate bed is located about 9.5 m above the base of the section, It is exposed more or less continuously along the 130 m length of the quarry.

Description of the carbonate bed: The carbonate bed varies in thickness from 0.4 - 0.5 m, and is a laminated to massive dolomitic peloidal mudstone that contains 6-23% insoluble residue comprising quartz silt, illite and chlorite with no microfossils. The bed can be divided into four laterally continuous layers each about 0.1 m thick (labelled a to d in Fig. 4A). The sedimentary features of the various layers are similar but vary in degree of burrow disruption and mudcrack disruption (Fig 4B, C, D). Layers are separated by brecciated zones a few centimeters thick as a result of numerous vertical and horizontal mudcracks. Laminae (Fig. 4B) are planar to undulatory and are either millimeter-thick, normally graded lenses of silt-sized peloids and quartz grains or submillimeter-thick mudstone layers traceable for hundreds of millimeters. Laminae commonly contain abundant, spar-filled, vertically oriented tubes 5-15 microns in diameter and up to 1 mm long (Fig. 4E). Laminae in layer b also contain calcispheres up to 0.1 mm in diameter (Fig. 4F) and ostracode shell fragments that become more abundant toward the top of the layer. Mudcracks originate from many surfaces within each layer and many are bridged by horizontal sheet cracks. Toward the top of the carbonate layer, mudcracks are filled with a finely laminated dolomitic mudstone, laminae being parallel to the crack walls. In some cases, crack-fill laminae are symmetrically arranged as in a vein filling (Fig. 4G). Burrows are horizontal and vertical tubes up to 20 mm in diameter (Fig 4D). Tubes up to 3 mm in diameter have simple blocky-spar mosaic cements. Larger diameter tubes commonly have laminated linings, peloids and blocky-spar central fills. Rare coarsening-inward drusy cements leave central voids.

Interpretation of the carbonate bed: The carbonate bed is interpreted to be the deposit of a groundwater-fed, freshwater marsh to shallow lake that developed within a flood basin during a time of local, reduced input of siliciclastic detritus. The carbonate is thought to have originated as mud-sized crystals precipitated around filaments within a cyanobacterial ("blue green algal") mat. Indeed, the micron-diameter tubes preserved in laminae are probably cyanobacterial filament molds.

Figure 4. Sedimentary features of the carbonate bed. A) Polished, etched slab of nearly complete carbonate bed, layers a-d labelled. Scale bar is 50 mm long. B) Preserved undulatory laminae (from layer a). Note large, compacted mudcrack in center of photograph, smaller mudcracks, and burrow (arrow). Scale bar is 50 mm long. C) Bedding plane section of mudcracked laminites. Scale bar is 50 mm long. D) Layer b. Note rare laminae, vertical and horizontal burrow tubes, and brecciated upper surface. Scale bar is 50 mm long. E,F) Thin-section photomicrographs of layer b. E) Alternating laminae of peloidal silt and dense mudstone. Note the micron-scale, vertical tubes (at arrows) and larger, spar-filled burrow tubes. The micron-scale tubes are interpreted as cyanobacterial filament molds. Scale bar is 1 mm. F) Calcispheres (arrows) in burrow-mottled mudstone. Scale bar is 0.2 mm. G) Laminated, horizontal crack fills which are symmetrically arranged ar arrow. Burrow tube cuts crack fill at the extreme right. Scale bar is 50 mm long.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Calcification of charophytic algae may also have been an important source of carbonate sediment as calcispheres are generally interpreted as the calcified Gyrogonites of Characean or Dasycladacean algae (Johnson 1961; Horowitz and Potter 1971; Flugel 1982). The generally massive nature of the carbonate bed and the preserved burrows suggest it was thoroughly bioturbated which, in turn, suggests that there were fairly prolonged periods of standing water. The burrowing fauna probably pelletized the original mud-sized cyanobacterial and algal debris. However, the small mudcrack and sheet cracks throughout the bed suggest periodic drying and the large mudcracks that brecciate layer boundaries suggest prolonged periods of exposure. The laminated mudstone fills of the crack network are interpreted as micritic cements and internal sediments where fills are not symmetrical about the crack center. That they were not too indurated is indicated by the fact that large burrows cut across them but small one do not.

This carbonate bed is very similar to the carbonate mud deposits of fresh-water marshes and lakes of the Everglades and interior of Andros Island (Gleason and Spackman 1974; Monty and Hardie 1976). These areas are covered by cyanobacterial mats ("Periphyton" of Gleason and Spackman 1974) in which carbonate mud is produced by light calcification of filaments due to local extraction of CO_2. Modern periphyton deposits range from complexly mudcracked and brecciated muds in marshy, seasonally-exposed areas to massive and laminated pellet muds in semipermanent lakes.

IMPLICATIONS OF THIS UNIQUE CARBONATE BED

Paleoclimate: Cyanobacterial marshes and lakes develop in rather specific subtropical climatic settings where short rainy periods alternate with seasonal dry periods (Monty and Hardie 1974). During the rainy periods, marshes become flooded as water tables rise above the surface. Cyanobacterial mats thrive during this submergence, and contribute their seasonal sediment. However, these periods must be short enough to prevent prolonged standing water. Where this happens, higher plants colonize the area and peat is formed. The paleoclimate of the Catskill alluvial plain in New York was probably tropically wet and dry, with a paleolatitude of approximately 20° south of the equator as deduced from a range of sedimentological, paleontological and paleomagnetic evidence (Woodrow and others 1973; Heckel and Witzke 1979; Woodrow 1985; Gordon and Bridge 1987). By way of comparison, average rainflall in the Everglades modern periphyton marshes is about 150 cm/yr with about two-thirds to three-quaters of this rain falling between June and October (Ginsburg 1964).

Deposition rate: The carbonate bed represents a decrease in terrigenous deposition rate, possibly due in part to the diversion away from the area of the major river channel that was supplying the sediment. However, it appears from the degree of bioturbation and calcareous paleosol development in the beds immediately below the carbonate that deposition rates were already low compared to those following carbonate deposition. The carbonate bed cannot be traced beyond the quarry so it is not possible to assess whether the reduced deposition rates were of local or regional importance. The Givetian-Frasnian boundary in New York is marked by a major transgression which had been associated with eustatic sea-level rise and reduced coastal deposition rates (Rickard 1975; Ettensohn 1985b; Johnson and others 1985). However, the carbonate bed was apparently deposited during or immediately after a period of marked regression (Fig. 2) and presumably high regional deposition rates relative fo subsidence and eustatic sea-level rise. The deposition rate of the carbonate bed cannot be directly ascertained. By way of comparison, as much as 3 m of uncompacted algal marsh deposits have accumulated in the Everglades over the last 5,000 years (Monty and Hardie 1976). The carbonate bed, therefore, may have been deposited within a few thousand years.

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ROAD LOG FOR DEVONIAN (CATSKILL MAGNAFACIES) FRESHWATER-CARBONATE FIELDTRIP

Route description: Travel underneath I88 at exit 15 and turn left onto NY 23, heading east. Go east approximately 7 miles. Immediately past the sign for Davenport Center, turn right sharply (hairpin bend) onto a dirt road. There is a cemetary on the hill above the dirt road. Davenport Quarry is 0.4 miles up the dirt road.

ACTIVE AND STAGNANT ICE RETREAT: DEGLACIATION OF CENTRAL NEW YORK

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INTRODUCTION

The depositional landforms of glacial origin on the Appalachian Plateau of central New York are atypical of classic continental environments. They consist of an unconventional landform assemblage formed by retreating ice-tongues at the margin of the Laurentide Ice Sheet. Landforms representative of backwasting occupy most through-valleys,while widespread stagnation and downwasting occurred in non-through valleys. A modified ice-tongue model includes valley tongues 20 km long and emphasizes the significance of inwash as a primary source of sand and gravel, and subglacial sediment flow within tunnels as the major source and transport mechanism for silt.

Previous work

This region received little published attention between Fairchild's 1925 description of the glacial landscape (kame and kettle topography, pitted plains, terraces, proglacial lakes and hanging deltas) and Coates' definition of the till shadow concept in 1966. Woodfordian facies of "bright" and "drab" drift were the subject of study in the Binghamton/Elmira area, (Denny, 1956; Moss and Ritter, 1962; Coates, 1963), but consideration of depositional environments along the main Susquehanna Valley remained unreported until described by Fleisher (1977a) and Melia (1975). Fleisher (1984, 1985) discussed the distribution of landforms as ice-marginal indicators and proposed an upper Susquehanna Lake chain, including glacial lakes Cooperstown, Davenport, Middlefield, Milford, Oaksville and Otego. Krall (1979) mapped and named the Cassville-Cooperstown Moraine (5 km south of the Otsego Lake at Index), which he interpreted as a evidence for readvance (circa 14,000 years BP) based on topographic correlation with moraines in the Hudson Valley. Sales, et al., (1977) provides a general summary of Otsego Lake geology, including bedrock, glacial and bottom samples. The first report of systematic sampling of lake sediments was by Yuretich (1979, 1981), who documents shallow stratigraphy, mineralogy and geochemistry in 25 short cores(30-60cm) from representative locations throughout the lake basin.

MacNish and Randall (1982) used well and boring log data to establish Quaternary aquifer properties and to provide generalized stratigraphic associations between surface data and landforms. They suggested various geomorphic and hydrologic settings to depict deglacial conditions related to rate and mode of retreat (active vs. stagnant).

The first suggestion of significant stagnant ice sedimentation was by Fleisher (1986a), who recognized anomalously large floodplain areas confined within high outwash terraces and valley train. These "mega-kettles", known as dead-ice sinks, developed in early post-glacial time by retarded downwasting of detached and buried remnant ice masses.

Regional Setting

The region is characterized by a coarsely dissected plateau of Middle Devonian strata that dip gently to the south-southwest at less than 10 degrees. Major north-south oriented valleys consist of broad, U-shaped troughs separated by wide, low-relief divides. Local topographic relief ranges between 250-300 m, but valley floor well data (Randall, 1972) indicate bedrock relief is considerably greater, averaging 300 m but reaching 400 m in places.

The Susquehanna River flows south, then southwest through a glacially modified, deeply incised, meandering valley (Figure 1). This valley and its strongly asymmetric tributary pattern of south-flowing streams are considered vestiges of a pre-glacial drainage system of great antiquity (Fleisher, 1977b, Sarwar and Friedman, 1990). Larger north-south oriented valleys are invariably asymmetric, with steeper slopes facing west and more gently inclined slopes to the east, from which the majority of tributary runoff originates. This topographic configuration is completely unrelated to bedrock strike and dip. While the cause of the valley asymmetry is not well understood, it appears to have played a significant role in late-glacial deposition of inwash-sourced sediment.

GLACIAL DEPOSITION

Upland diamict on divide areas is generally thin, ranging from a discontinuous veneer to 5 m, except in till shadows in the lee of divide ridges and hills, where test borings indicate thicknesses reach 40-60 m (Resource Engineering, 1986). Valley floor deposits consist of sand and gravel interstratified with sand, extensive silt and some clay in thicknesses that commonly reach 70-120 m. By correlation with deposits in adjacent drainage basins, the chronology is interpreted to be Woodfordian and range from 16,000 to14,500 years BP (Cadwell, l972b). Fleisher (1987) interpreted stratigraphy and deglacial landforms to be consistent with single-stade development. Consistent with this are generalized diagrams of MacNish and Randall (1982) that suggest single-stade settings based on regional well data and general topographic expression.

Listed below are landform assemblages for through and non-through valleys. These are interpreted to indicate active ice retreat (backwasting) in through valleys and stagnation (downwasting of detached ice masses, as well as ice-tongue collapse) in non-through valleys.

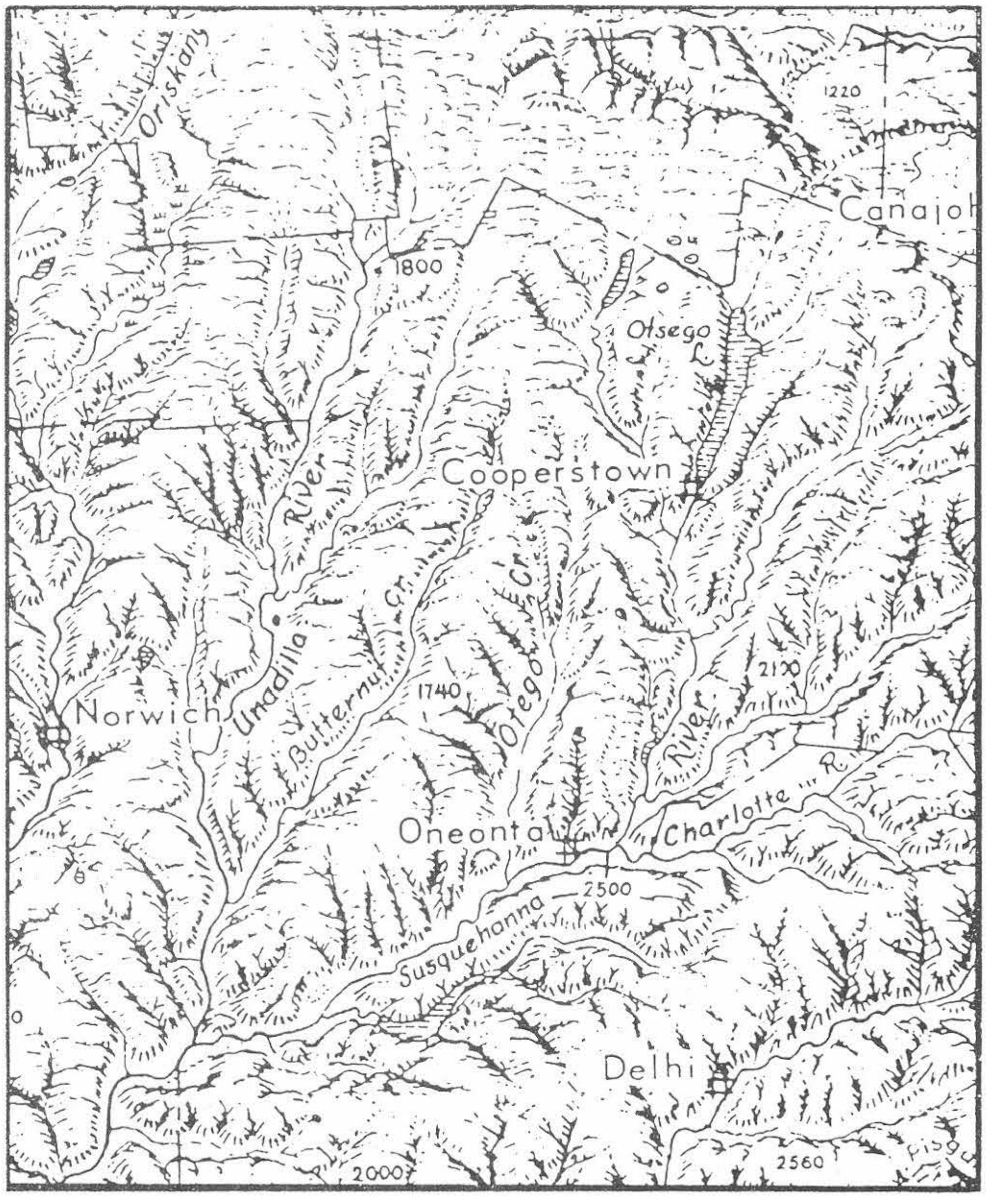

Figure 1 Index Map

THROUGH VALLEY ASSEMBLAGE	NON-THROUGH VALLEY ASSEMBLAGE
(produced by backwasting)	(produced by downwasting)
kame moraines/valley train	kames and kame fields
high gravel terraces (deltaic)	discontinuous gravel plain remnants
lacustrine plains	dead-ice sinks
kame fields	eskers
dead-ice sinks	

Kame moraines general occur in association with valleys trains that grade upvalley to the base of the moraine. A conspicuous kame and kettle topography, with local relief of 10-15 m, identifies moraines. Most moraines still occupy the full valley width, although breached by modern drainage and rise 25-30 m above the floodplain. They are poor morphostratigraphic units owing to the loss of topographic expression on valley walls and are essentially absent from upland areas. The moraine/valley train landform association is interpreted to be indicative of an active ice-marginal position (actively flowing ice) because it is assumed that a high-discharge, hydrological connection with meltwater from the main ice sheet was required to aggrade thick sand and gravel in the form of a massive valley train. Local, clusters of small scale ice blocks were occasionally incorporated within some gravels as indicated by the pitted nature of valley trains. The cause of ice block concentration in some areas and not others remains to be determined, but may be related to clusters of grounded ice-islands and bergs in emptied proglacial lakes across which inwash accumulated. This may account for why 20 m of pitted outwash lies over 110 m of silt in the vicinity of Milford Center on the Susquehanna near Oneonta.

Paired and non-paired "kame terraces" are common and typically contain conspicuous deltaic foreset and topset bed. They stand 25-30 m above the floodplain and are graded remnant patches of other planar gravel deposits and hanging deltas. These features are commonly referred to as kame terraces, but actually appear to have originated by meltwater streams discharged from lateral englacial tunnels where ice-tongues were in contact with proglacial lakes. In this sense, they are deltas formed by continuous aggradation during slow, active ice retreat.

Lacustrine plains occur commonly throughout the region and occupy large segments of many through and non-through valleys. They are underlain by deposits of silt and "quicksand" (driller's term for saturated coarse silt and fine sand that liquifies when vibrated) reported in driller's logs to exceed thicknesses of 100m (Randall, 1972). However, the sparse occurrence of well developed hanging deltas and paucity of other strandline features seem inconsistent with the extent of lakes suggested by thick silts and extensive lacustrine plains. Perhaps this indicates very rapid rates of lacustrine sedimentation in short-lived lakes.

Kame fields are found more commonly in non-through valleys, although not exclusively. They are of limited lateral extent (seldom exceeding 1-2 km across), with local relief similar to that of moraines, yet found independent of valley train. Invariably, kame fields are found at the confluence of upland tributaries and main valleys. As a landform of ice-cored origin, downwasting of remnant ice masses appears to be the primary mode of deposition.

Another landform of ice-cored origin is the dead-ice sink. It's presence is apparent only when viewed in the context of adjacent landforms. Conceptually, it is an exceptionally large kettle that occupies the full valley floor width. Although formed in much the same way as a kettle, sinks are many times larger. They appear as anomalously broad floodplain areas (3-4 km in diameter) bounded upvalley and down by terraces, valley train, kame fields or moraines. Implicit to their occurrence is the burial of large stagnant ice mass remnants of collapsed ice-tongues or detached from the terminus of retreating active ice tongues. Downwasting ultimately creates a valley-floor depression that serves as a sediment sink during late-glacial time, or post-glacial time if melting is sufficiently retarded.

Very few eskers are know within the upper Susquehanna. However, those that have been mapped all are short (less than a km) and of low relief (10-14m). They typically occur as part of the non-through valley landform assemblage and are assumed to indicate stream deposition beneath dead-ice.

VALLEY ICE-TONGUE MODEL

The valley ice-tongue model, proposed by Moss and Ritter (1962), was used by Cadwell (1972a) to correlate the valley facies of ice-marginal positions with isolated stratified drift on adjacent slopes in the Chenango drainage basin. Fleisher (1977a) applied ice-tongue depositional environments to account for the common occurrence of deltaic sand and gravel kame terraces and explain patterns of ice-marginal landform distribution within valleys of the Susquehanna drainage basin. MacNish and Randall (1982) proposed rate of ice-tongue retreat as a controlling factor in the development of landforms and stratigraphy. Although discussion is limited to descriptive generalities, their schematic diagrams clearly relate rate of retreat to types of deposits. Slow retreat favored the formation of deltaic kame terraces in much the same way described by Fleisher (1977a). Hesitation during retreat permitted aggradation to occur across full valley width to form a recessional moraine, although no specific moraine forming process (i.e. sediment source or transport mechanism) is identified. Conversely, rapid retreat precluded lateral accretion by outwash streams and glaciofluvial sand and gravel was interbedded with lacustrine sediments. A completely different set of landforms, including eskers, and heterogeneous stratigraphic units of limited lateral continuity are attributed to stagnant ice conditions.

Shortcomings to the earlier ice-tongue model

Although the earlier valley ice-tongue model has been useful, it fails to account for several curious aspects of regionally extensive deposits and does not portray sufficient details pertaining to depositional environments. For example, the current model does not address the following observations:

1. landforms in through valleys differ significantly from those in non-through valleys
2. kame-moraines
   a. lose topographic expression on valley walls and across divides
   b. consist of crudely sorted and stratified sand and gravel, not till
   c. are graded to thick valley train through valleys
   d. are difficult to correlate from one valley to the next
3. thick silts occupy valleys that lack well developed strandline features, such as hanging deltas and beach deposits

Furthermore, it is difficult to apply this model to deglacial conditions that would account for:

1. an adequate source of coarse and fine sediments
2. sediment transport mechanisms to account for the contrast between stratified valley deposits and till-like upland drift
3. a single-stade environment that would result in 20 m of pitted sand and gravel over 80 m of silt (lacustrine ?)

While the ice-lobe model is helpful in understanding the distribution of glacial landforms within valleys and provides a conceptual model to guide regional correlation Fleisher (1986b), it does not apply well to upland areas where end and recessional moraines are virtually lacking. Therefore, the valley ice-lobe model must be modified to better accommodate field evidence.

Modified ice-tongue model

As a starting point in the modification of the ice-tongue model, it was assumed that "the glacier profile is related to the hydraulic and strength properties of potentially deformable bed material" (Boulton and Jones, 1979). In central New York, the subglacial zone of the Laurentide ice sheet would have been charged with saturated, fine sediment of low permeability derived from the erosion of lower Paleozoic shales and siltstones as the ice moved southward onto the Appalachian Plateau and into the Susquehanna drainage basin. As indicated by Boulton and Jones, when "bed transmissibility is low, water pressure builds up, the bed begins to deform, and a lower equilibrium profile will develop". Proglacial, ice-contact lakes into which subglacial meltwaters discharged during deglaciation would detain free downvalley flow, thereby contributing to the development of positive pore pressure build up within saturated bed material. This in turn reduced basal shear strength, facilitated deformation and resulted in lower ice-tongue gradients.

Two methods of gradient calculation were considered. The Mathews (1974) method is summarized by the general expression,

$$h = A\,x^{0.5}$$

where h = ice thickness, A = a coefficient that varies with different values of basal shear stress, and x = distance from glacier terminus.

The Ridky and Bindschadler (1990) method is based on Nye (1952) and Hughes (1981) and involves a much more sophisticated approach, which was used to develop ice thickness profiles and flowlines for Late Wisconsinan ice in the Finger Lakes region of central New York. Their calculations, based on basal shear stress values of 0.5, 1.0, and 1.5 bars, compare favorably with profile data generated by using the general Mathews equation (Table 1).

Table 1 shows how these gradients may be used to estimate the length of ice-tongues that rise gradually upvalley before joining the ice sheet margin on adjacent divides. Because local bedrock relief (from beneath valley fill to outcrops on divides) averages 300 m, the length of an ice-tongue would be approximated when ice thickness equals bedrock relief, as shown in Table 1. General agreement between ice-tongue height values that span 300 m, as determined by both methods of calculation (at comparable basal shear stress), suggests limited extrapolation to shear stress values less than 0.5 bars is valid. This technique suggests ice-tongues may have reached lengths in excess of 20 km.

TABLE 1. ICE-TONGUE SURFACE GRADIENTS

	Mathews*					Ridky & Bindschadler**		
	Basal shear stress (bars)							
	0.10	0.30	0.50	1.00	1.50	0.50	1.00	1.50
Distance from terminus (km)	Height of ice surface (m)							
2	57	100	134	182	223	142	204	251
4	81	141	189	257	314	191	278	345
6	100	172	223	314	385	221	327	409
8	115	199	268	363	445	294	417	511
10	129	223	287	406	498	295	432	538
12	141	244	314	445	545	397	547	663
15	157	273	352	498				
20	182	314	423	575				
30	223	385	518					
40	257	445						
50	287							
60	314							

Underlined sets of data indicate height values that bracket 300 m, which is required to exceed local bedrock relief of divides above valley floor. Each set corresponds to a predicted range of ice-tongue lengths.

* Mathews method

$h = A\,x^{0.5}$, where h = ice thickness, A = a coefficient that varies with different values of basal shear stress, and x = distance from glacier terminus.

** Ridky and Bindschadler Method
(see original reference for derivation of formula)

These calculations assume specific quantitative conditions that may be more precise than accurate for the purpose of developing a conceptual working model. Therefore, it should be emphasized that the most important aspect of the modified model is the extensive ice-free uplands on which inwash processes could function, not the finite interpretation of ice-tongue length (Figure 2).

An additional significant modification of the ice-tongue model involves the dynamics of longer ice-tongue movement leading to stagnation on the scale of an entire ice-tongue (20 km long). Ice-tongue stagnation has been proposed as the result of the progressive development of a negative ice budget within through valleys caused by restricted internal flow within thinning ice on headward divides (Fleisher, 1986a). Figure 3 (schematic longitudinal profiles) illustrates how ice-tongue starvation developed in non-through valleys, whereas active ice movement within through valleys (open-to the-north) was sustained by continued internal flow.

The occurrence of dead-ice sinks in through valleys implies still another cause for stagnation that involved the detachment of large ice masses (a few km in diameter) from the ends of actively flowing ice-tongues during backwasting, as indicated by Fleisher (1986a). A possible detachment mechanism, proposed by Mulholland (1982) suggests *"tongues of ice reached a critical level of reduced compressive strength after which they could no longer transmit directed basal shear stress. When this critical thickness was achieved, a line of failure would develop some distance up glacier, separating thick, strong, active ice from thin, weak, stagnant ice"* (see Figure 4).

The terminus of the Bering Glacier, Alaska, contains such structures and provides a modern analog for this detachment mechanism. Here, low angle shears rise from within the glacier to separate tabular plates of ice a few meters thick (Figure 5). Similar remnant ice plates are buried beneath several meters of outwash gravel and lacustrine sediment 2 km from the glacier terminus. Here,ice foliation and multiple shear planes are oriented semi-parallel to those observed in the Bering (Figure 6). This suggests detachment occurred during retreat from a 1966 surge terminus (Fleisher, 1991).

SEDIMENT SOURCES

Evenson and Clinch (I987) document the significance of meltwater in moving sediment from upvalley sources to the glacier terminus. Their study identifies specific mechanisms responsible for down-glacier transport of inwash derived from ice-free tributary valleys, ice-dammed lakes, slope processes and re-worked older deposits, as well as inwash from tributaries beyond the existing ice limit.

The modified ice-tongue model proposed here combined with the inwash concept advanced by Evenson and Clinch lead to mechanisms of sediment transport that account for the origin of all sediment associated with Laurentide deglaciation. Central to the modified model are ice-tongues 20 km long within valleys between ice-

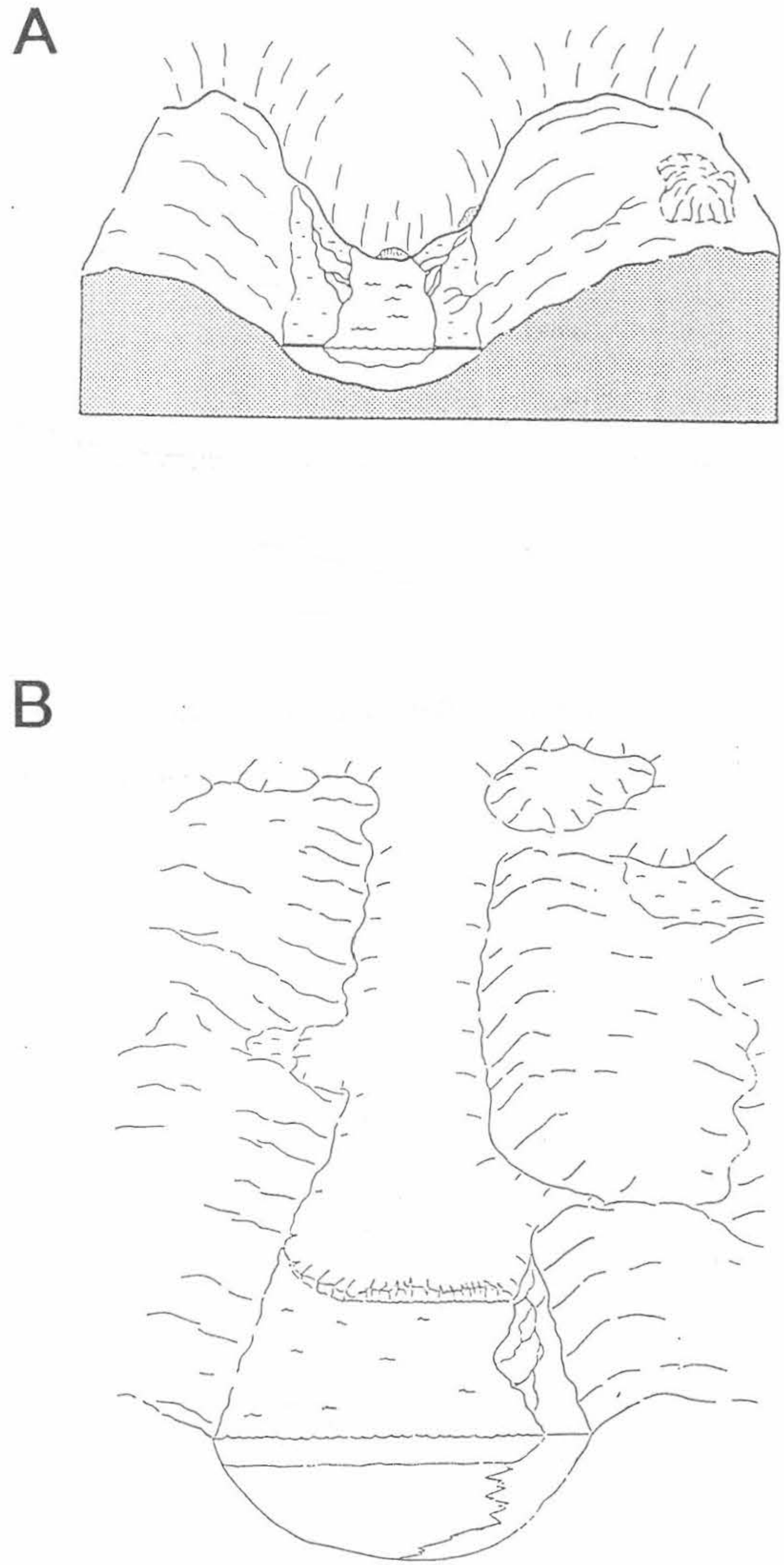

Figure 2 Modified Ice-Tongue Model. The modified model (B) depicts depositional conditions that differ significantly from those of the earlier model (A) and includes several possible sediment sources and debris transport mechanisms.

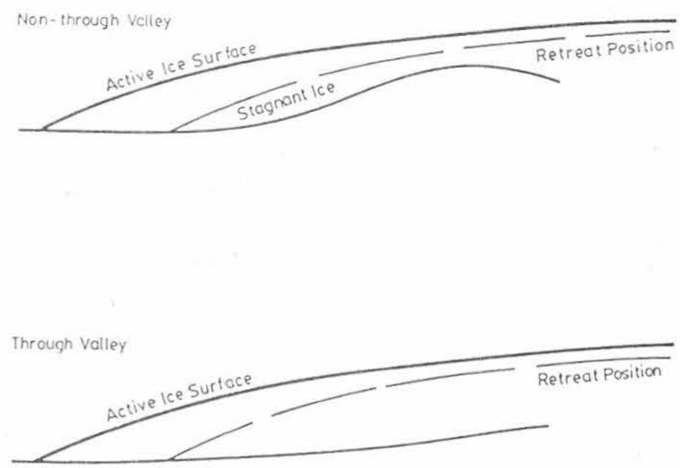

Figure 3 Schematic longitudinal profiles

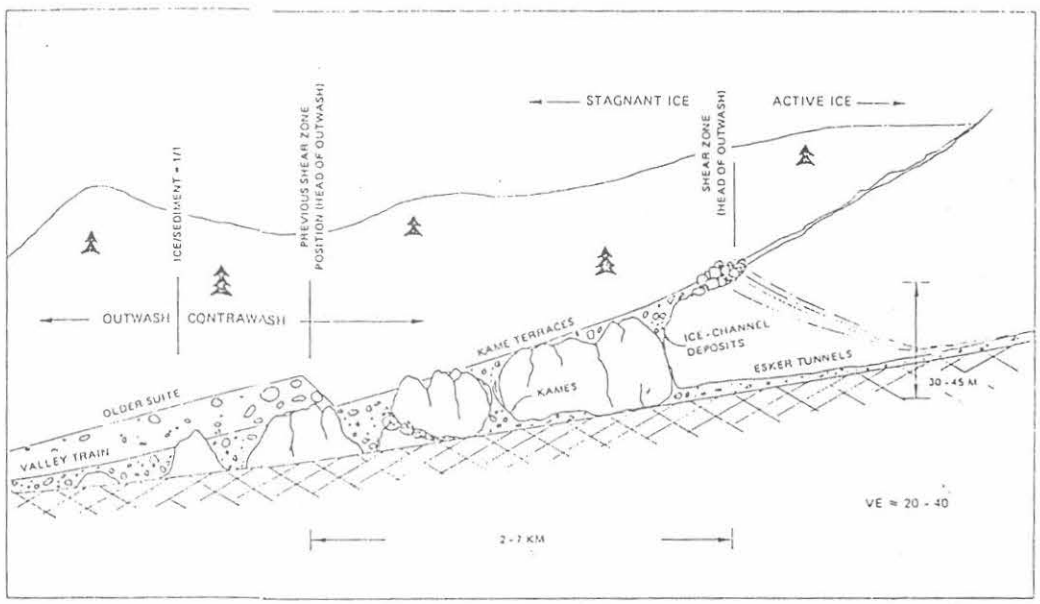

Figure 4 Detachment of stagnant ice masses (from Mulholland, 1982)

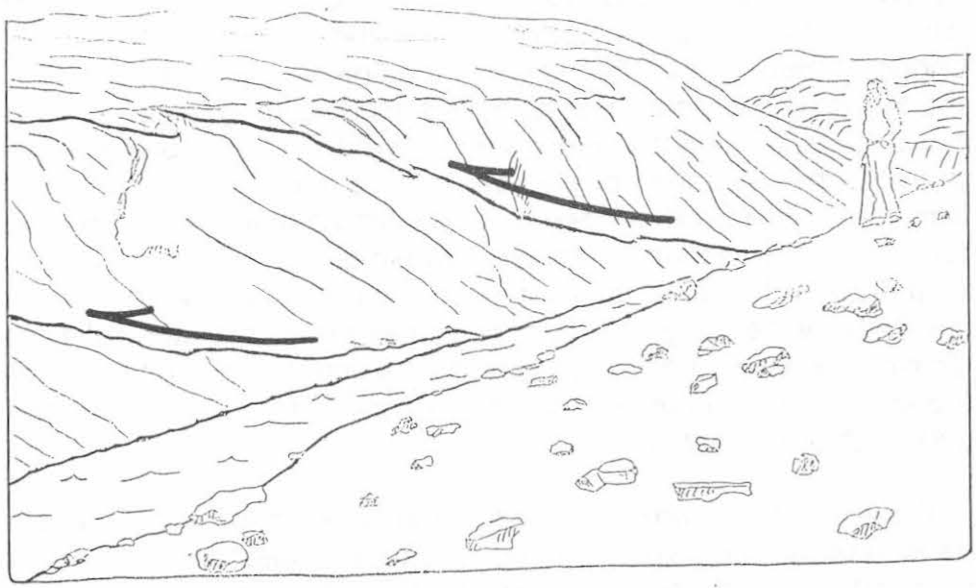

Figure 5 View up-glacier of rising shear planes on vertical canal wall at the Bering Glacier terminus.

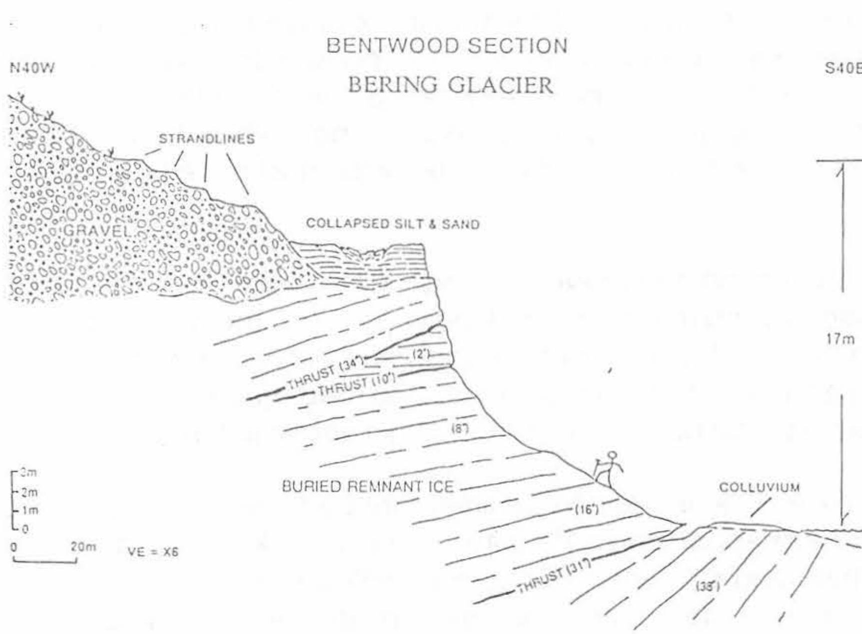

Figure 6 Shear planes separate four tabular ice plates (Bentwood Section). Foliation in each plate dips in the upglacier direction and is truncated by shears. Overlying sediment includes lacustrine silt and sand disturbed by subsidence as remnant ice below melts (from Fleisher, 1991).

free uplands and deglaciated tributaries. Consequently, bouldery till originally deposited in tributary valleys was re-worked and transported to the surface of adjacent ice-tongues, where re-sedimentation sorted fine from coarse material. Downglacier transportation by supraglacial streams fed accumulating deltas in ice-contact proglacial lakes. Coupled with uniform, slow retreat this process formed continuous lateral deltas that aggrade upvalley as space is made available during retreat. These join with kame terraces deposited by associated streams along the glacier/valley wall interface. Fluvial transport of tributary inwash directly into a proglacial lakes formed deltas at grade with deltaic terraces. Active ice-tongue flow concentrated supraglacial inwash at stable ice margins to form moraines and associated valley trains, whereas less active flow allowed debris to remain concentrated in areas near tributary confluences. Where protected from supraglacial streams, such deposits formed kame fields through downwasting.

While tributary inwash seems to be an adequate source of coarse sediment, a comparison of sediment volume to upland area suggests the inwash process could not have yielded sufficient fines to account for the volume of fine sand and silt within the valley fill. Uplands are mantled by silt-rich lodgement till that is generally thin, except in till shadows where it reaches thicknesses in excess of 40-60 m. To test the possibility that upland drift may have served as a significant source of fines, the volume of fines in the valleys was compared to the area of adjacent uplands.

Well data from the upper Susquehanna and all major tributaries indicate the valleys contain approximately 58 cubic km of fine sediment fill. Distributing this volume as a uniform blanket over the entire upland area (4455 sq. km) would add l6 m to the existing till mantle. This amount of upland erosion in rills, gullies, mass wasting and sheet wash would have produced conspicuous evidence of dissection, which does not exist. Furthermore, winnowing fines from that much till would produce widespread lag gravel, which is also lacking. While these processes may account for some fines, the bulk of proglacial lake silts must have been derived from other sources.

Most sand and gravel is interstratified within fine sand and silt, which establishes contemporaneous deposition. Yet, the source of fines does not appear related to inwash sources of the sand and gravel. Assuming minimal silt from supraglacial and englacial sources, and inadequate amounts from uplands, the only possible remaining source is the subglacial sediment load of valley ice-tongues.

Extensive fine clastic sedimentary bedrock was exposed to Laurentide ice on the northern Appalachian Plateau. Here, a thick lower Paleozoic section contains 60-70% shale and siltstone. Subglacial bedload derived from these rocks would consist of fine-grained, deformable sediment incapable of draining a constant influx of meltwater. These conditions favor development of subglacial channels and conduits through which excess water would drain (Boulton and Hindmarsh, l987). As water discharge increases, so does the piezometric gradient, which in turn raises water pressure values within the sediment to equal ice overburden pressure. As Boulton and Hindmarsh suggest, this leads to sediment liquification in the terminal zone and the "flow of liquified sediment into the proglacial environment."

Similar conditions are proposed to have existed beneath Laurentide ice-tongues on the Appalachian Plateau. The saturated subglacial sediment, driven by englacial water, served the dual function of providing; l) low basal shear stress leading to low gradient, long ice-tongues and 2) large discharge of silt-rich sediment directly into ice-contact, proglacial lakes. Therefore, it is proposed this process provided a major source of fine sediment by a transport mechanism of subglacial sediment flow. Because non-through valleys were incapable of maintaining steady-state ice flow conditions, ice-tongues collapsed, downwasted and left non-time-transgressive inwash in an assemblage of dead-ice landforms (Hughes, l987; Kaszycki, l987, Mullins and Hinchey, 1989).

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CASE STUDIES

The following case studies discuss examples of deglacial processes and resulting landforms at specific locations within the upper Susquehanna drainage basin.

Case #1: A Case for remnant ice, inwash and sediment source, Glacial Lake Cooperstown, Cooperstown, New York

Case #2: Remnant ice as base level controls for stratified drift aggradation: Otego Creek, New York

Case #3: Inwash sediment sources, Valley of Charlotte Creek, central New York

Case #4: Implications of pebble count data; confluence of Unadilla River and Tallette Creek, Columbus Quarters, New York

Case #1: A Case for remnant ice, inwash and sediment source, Glacial Lake Cooperstown, Cooperstown, New York

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INTRODUCTION

Otsego Lake is the headwaters for the Susquehanna River near the northern extent of the eastern Appalachian Plateau (Figure 1). The historic village of Cooperstown is situated on its southern shore and the hamlet of Springfield Center is a few kilometers from the northern shore. Otsego Lake lies within one of several through valleys in the region. To the west is Canadarago Lake, to the east Cherry Valley, all oriented N10E as part of a regional stream pattern, along which ice flow was facilitated. Local relief varies from 700' in the vicinity of Cooperstown to 900' northward. Bedrock relief is observed to be greater at the southern end of the lake where valley fill is thickest.

Oaks Creek drains Canadarago Lake and joins the Susquehanna River 4 km south of Cooperstown near the hamlets of Index, Hyde Park and Phoenix Mills. Red Creek enters the Susquehanna Valley 1.5 km south of Cooperstown from a non-through valley on the divide east of Otsego Lake. The Susquehanna Valley at Otsego Lake is conspicuously asymmetric, as are most through valleys in this region. Much steeper west-facing slopes are unrelated to the gentle southern dip of Devonian strata. Short streams drain small, first-order catchment on the eastern slopes, whereas streams off western slopes flow from second order drainage basins and cover a significantly larger area than those from the east.

GLACIAL LANDFORMS

The conspicuous kame and kettle topography at Index (5 km south of Cooperstown) is the Cassville-Cooperstown moraine, named and interpreted by Krall (1977) as the terminal deposit of a readvance here and at Cassville, 50 km to the northwest. The moraine grades downvalley into a short, deeply incised valley train. Both consist of sand and gravel, which is only slightly less well sorted and stratified in the moraine. Local relief on the moraine ranges from 7-20 m, with kettle frequency diminishing toward the valley train. Krall mapped the northwest trend of the moraine along Oaks Creek valley to Fly Creek and Oaksville, where it served as a dam for Glacial Lake Oaksville, the precursor to Canadarago Lake (Fleisher,1977). To the east, the moraine is correlated with kames in the valley of Red Creek.

The terrain upvalley from the moraine, between Index and Cooperstown, is anomalous. It consists of a unique assemblage of landforms unlike those typical of either active or stagnant ice retreat, as defined by MacNish and Randall (1982). Although the narrow floodplain here is gentle enough to support tight river meanders, the adjacent valley floor is 7-10 m higher and much too irregular to qualify as a

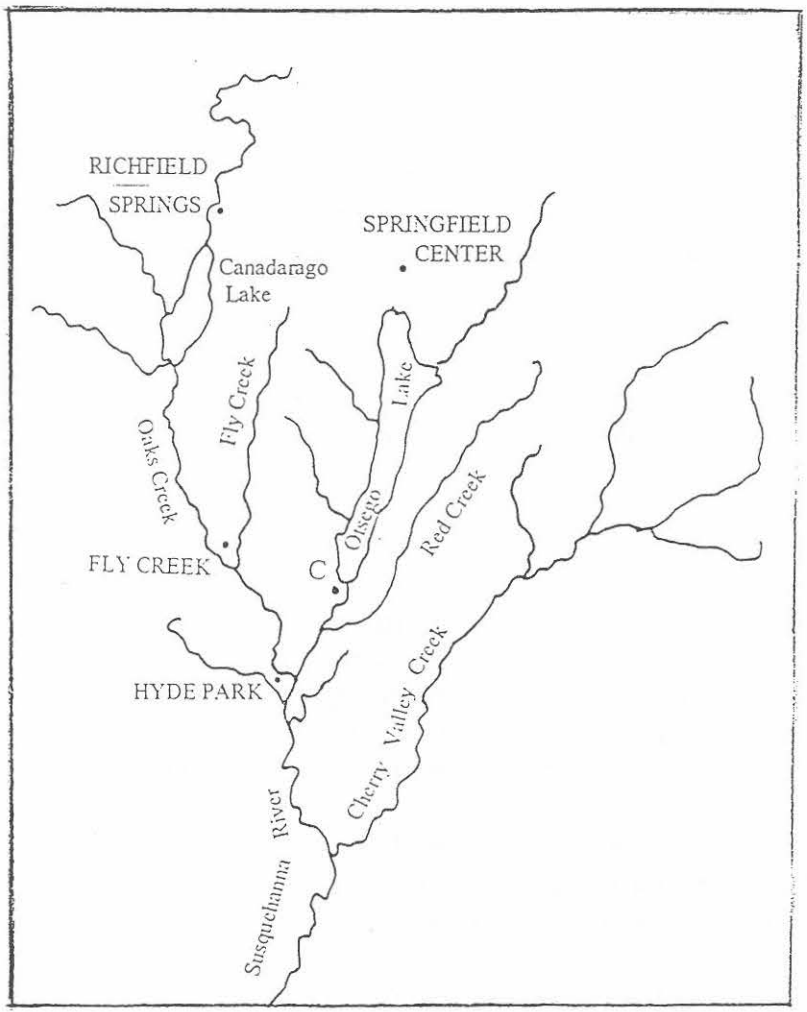

Figure 1 Index map

lacustrine plain, although composed of silt at depth. It rolls gently, suggestive of subdued terrain, but is not consistent with landforms immediately upvalley or down.

This *misfit terrain* gradually yields upvalley to gravel terrace remnants at the mouth of Red Creek and others at the mouth of a few small first-order streams on the valley walls immediately south of Cooperstown. The village of Cooperstown is situated upon the dam for Otsego Lake (possibly a moraine), referred to here as the *"Doubleday Ice Margin"*.

ON-LAND GLACIAL STRATIGRAPHY

A clear record of valley fill stratigraphy is represented by published well and boring data (Randall,1972), technical reports (Resource Engineering,1986, unpublished) and through personal communication with local water well drillers. Subsurface data shown in Figure 2 represents a semi-continuous stratigraphic record of deposits formed during retreat from the Cassville-Cooperstown margin between Index and Cooperstown.

While silt behind a moraine might suggest lake sedimentation, the lack of a lacustrine plain and the absence of strandline features (i.e. hanging deltas at tributary mouths) indicate otherwise.

FORMATION OF MARGINAL-ICE CLEAT AND DEAD-ICE SINK

Projection of a low ice gradient (consistent with a basal shear stress of 0.3-0.5 bars) northward from the Cassville-Cooperstown margin depicts an ice-tongue surface rising to the main ice margin along north-facing upland slopes north of Otsego Lake. This means the entire lake basin held an ice-tongue approximately 20 km long, while adjacent slopes were ice-free and subject to inwash processes.

Retreat from the Cassville-Cooperstown margin produced landforms and stratigraphy atypical of normal active ice retreat. In order to account for this, the concept of a *"marginal-ice cleat"* is introduced.

The cleat concept is based on field observations at the terminii of many active glaciers in Alaska, where active ice appears to rise from within the glacier on deep seated shears. Such is the case for structures along the retreating terminus of the Bering Glacier, central coastal Alaska. Here, discretely separate, semi-horizontal plates of ice, bound above and below by shear planes, form a cleat of passive ice on which active ice has risen. Several km from the retreating ice front, remnant ice plates have been detached and buried beneath foreland drift (Figure 3) during the last 20-30 years (Fleisher, 1991).

It is suggested here that similar rising structures developed during retreat from the Cassville-Cooperstown margin (Figure 4a) in a series of detachment planes (shears), as proposed by Mulholland (1982) for New England deglaciation.

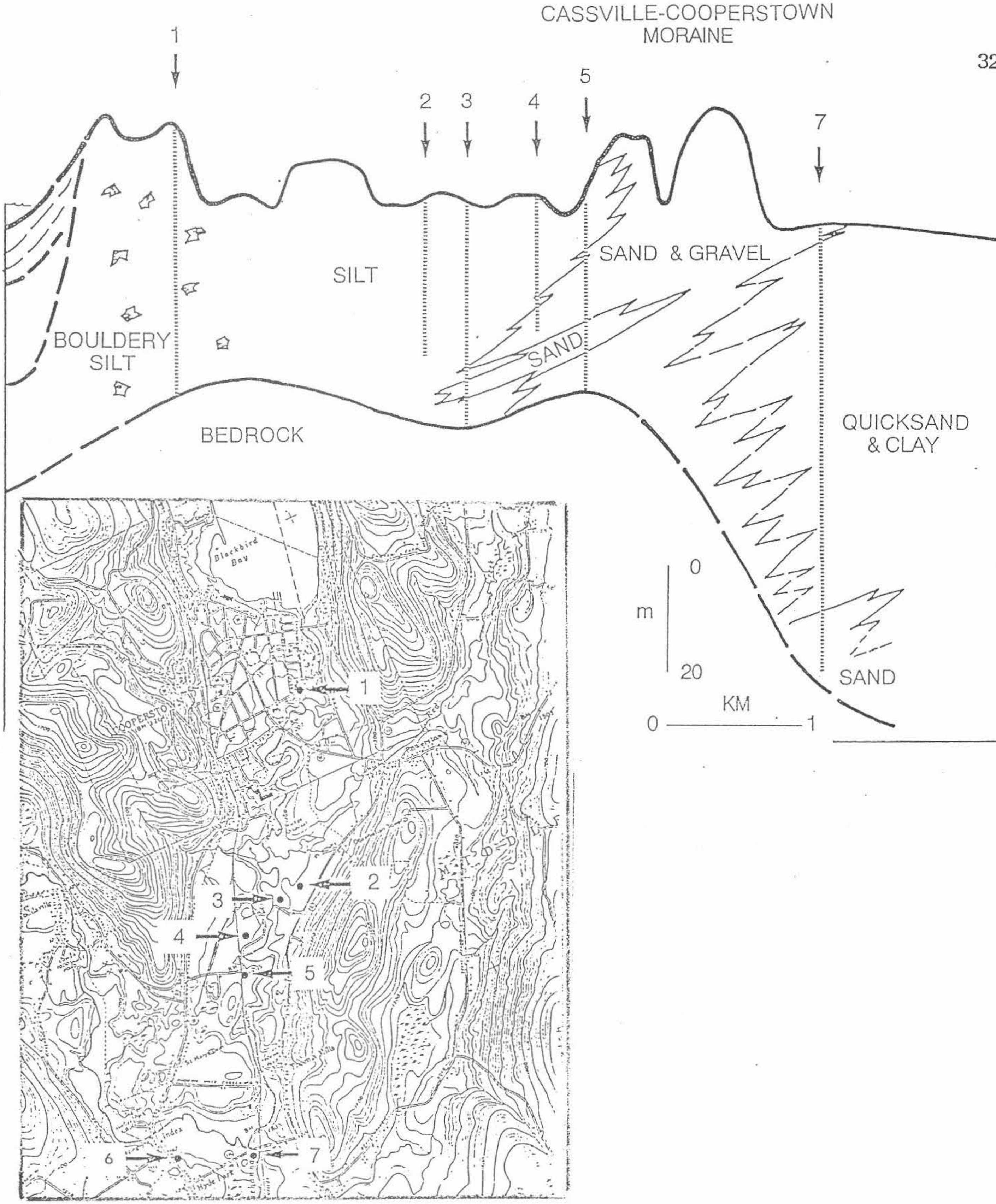

Figure 2 Well data and cross section. On-land subsurface data are derived from several locations between Cooperstown and Index, as shown on the index map. The correlation of stratigraphic units and their occurrence beneath associated landforms is illustrated. (from Fleisher, et al, 1990)

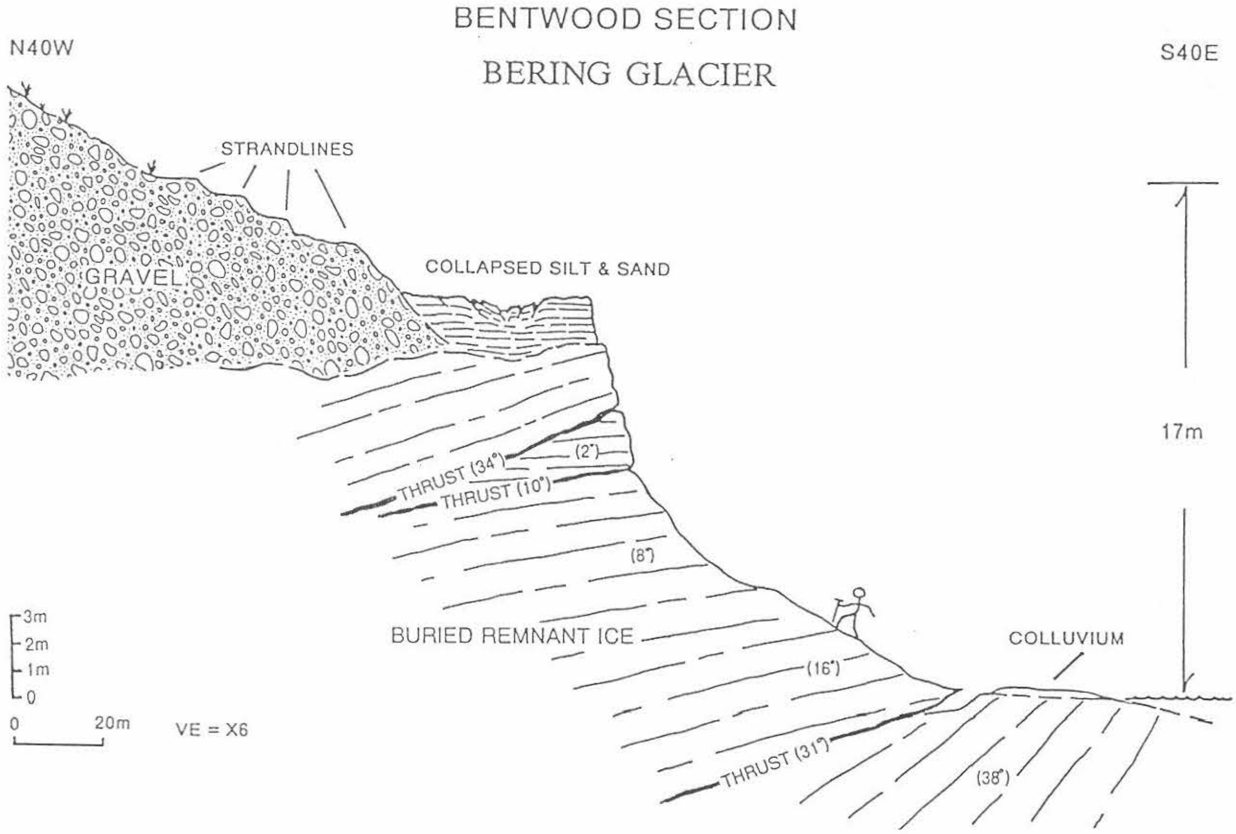

Figure 3 Detached remnant ice beneath foreland drift several km from retreating ice front. (from Fleisher, 1991)

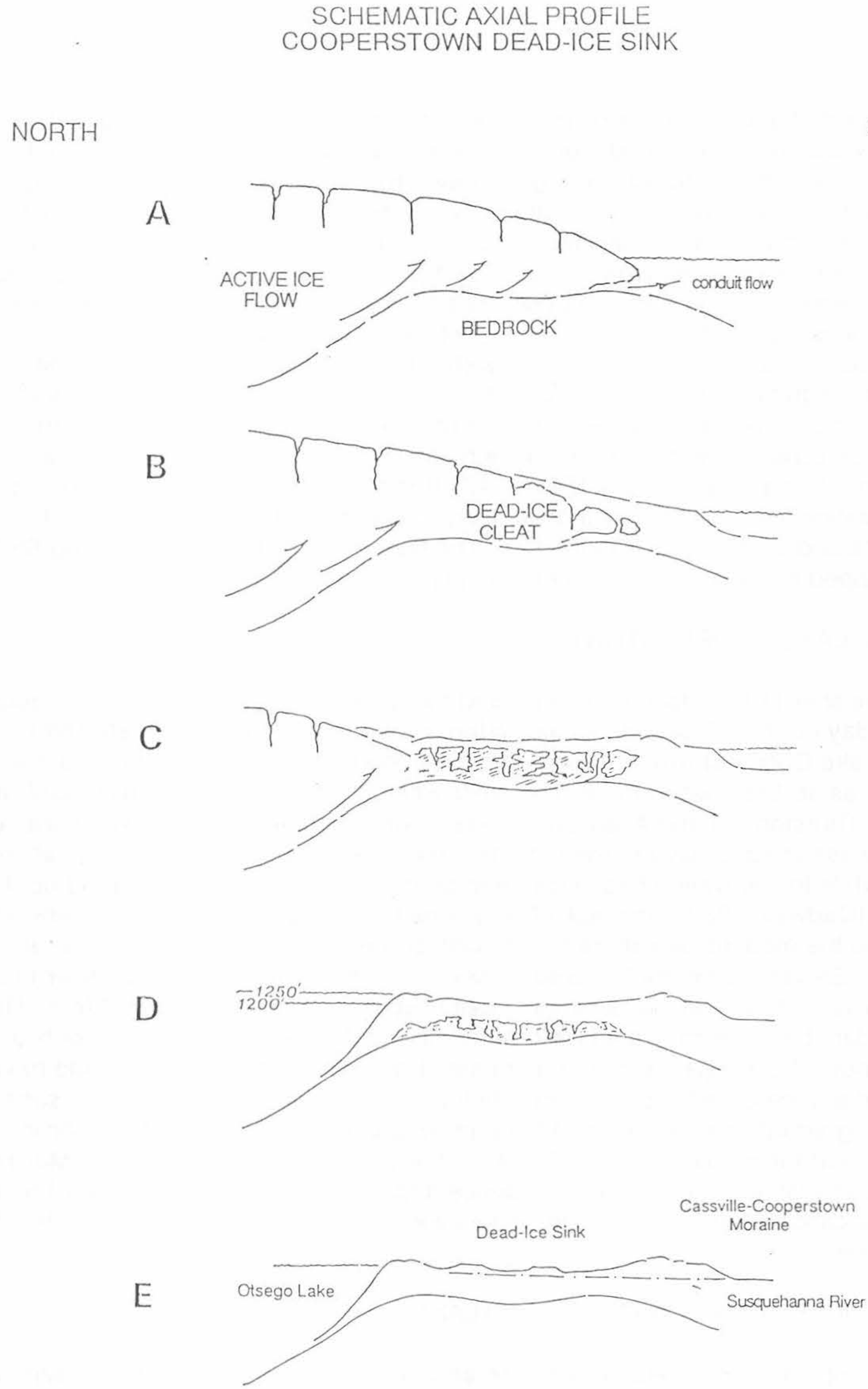

Figure 4 Dead-ice sink development. Schematic diagram of rising structures in the terminal zone of the retreating ice-tongue leads to detachment of marginal-ice cleat and subsequent development of dead-ice sink. (from Fleisher, et al, 1990)

Detachment of a marginal ice-cleat (a stack of imbricate ice plates) 65-70 m thick, 1 km (3,000') wide and 3 km (9,000') long filled the valley, as active ice rose against its upvalley side at the Doubleday margin (Figure 4B). Inwash, primarily from Red Creek and Willow Creek, spread an insulating cover across the cleat causing retarded melting and slow, gradual subsidence. Coarse sediment from these tributary sources was supplemented by silt-ladened overflow from incipient Glacial Lake Cooperstown. Local ephemeral ponds and lakes collected silt over buried ice, analogous to ice-contact lakes reported by Cadwell, et al.,1990 and Gonsalves, et al., 1991. Continuously changing water levels adjusted to new outlets in downvalley ice-cored deposits. High turbidity favored rapid silt accumulation in a constantly adjusting lake system, which precluded the development of distinct strandline features. Semi-continuous subsidence over a buried ice mass of this dimension qualifies this environment as a dead-ice sink (Figure 4C) (Fleisher, 1986). However, ongoing sedimentation filled the sink as it developed, creating a unique association of landforms and stratigraphy (Figure 4D). The final result is a valley containing 65-70 m of silt capped by inwash gravel, yet lacking lacustrine landforms (Figure 4E).

GLACIAL LAKE COOPERSTOWN

Fleisher (1977) documented incised hanging deltas and near-surface lacustrine silt and clay north of Otsego Lake as evidence for a higher lake stage referred to as Glacial Lake Cooperstown. Hanging deltas are found at the mouths of all west-slope streams, as at Brookwood Point, Threemile Point, Fivemile Point, Sixmile Point, and from the Thurston Hill and Allen Lake areas. The Cassville-Cooperstown moraine was originally assumed to have formed the dam for Lake Cooperstown at an elevation of 1250'. While this elevation has since been confirmed by strandline mapping north of the lake (Cadwell, 1988), the lack of lacustrine landforms between the moraine at Index and the modern lake shore to the north suggests the dam was actually farther upvalley. Elevations on the Doubleday Margin (at the village of Cooperstown) range between 1220-1240', whereas hanging deltas stand at a consistent 1250 feet. How can the dam be lower than the lake? Although the Doubleday Margin is not high enough now, it is suggested that at the time of original impoundment it would have been partially ice-cored and, therefore, higher. As buried ice slowly melted, surface elevation gradually diminished and the dam was eventually breached. A conspicuous alluvial fan at the mouth of Willow Creek on the western end of Doubleday Margin indicates inwash kept pace with subsidence and forced the spillway eastward to its present location where incision ultimately lowered lake level to the modern elevation of 1200 feet.

SEISMIC INVESTIGATION OF OTSEGO LAKE

Complementing on-land evidence and further documenting the depositional history of Glacial Lake Cooperstown are the high-resolution seismic reflection profiles obtained by Mullins and Hinchey (1988) along twenty-one east-west oriented transverse profiles one north-south oriented longitudinal (axial) profile (Figure 5).

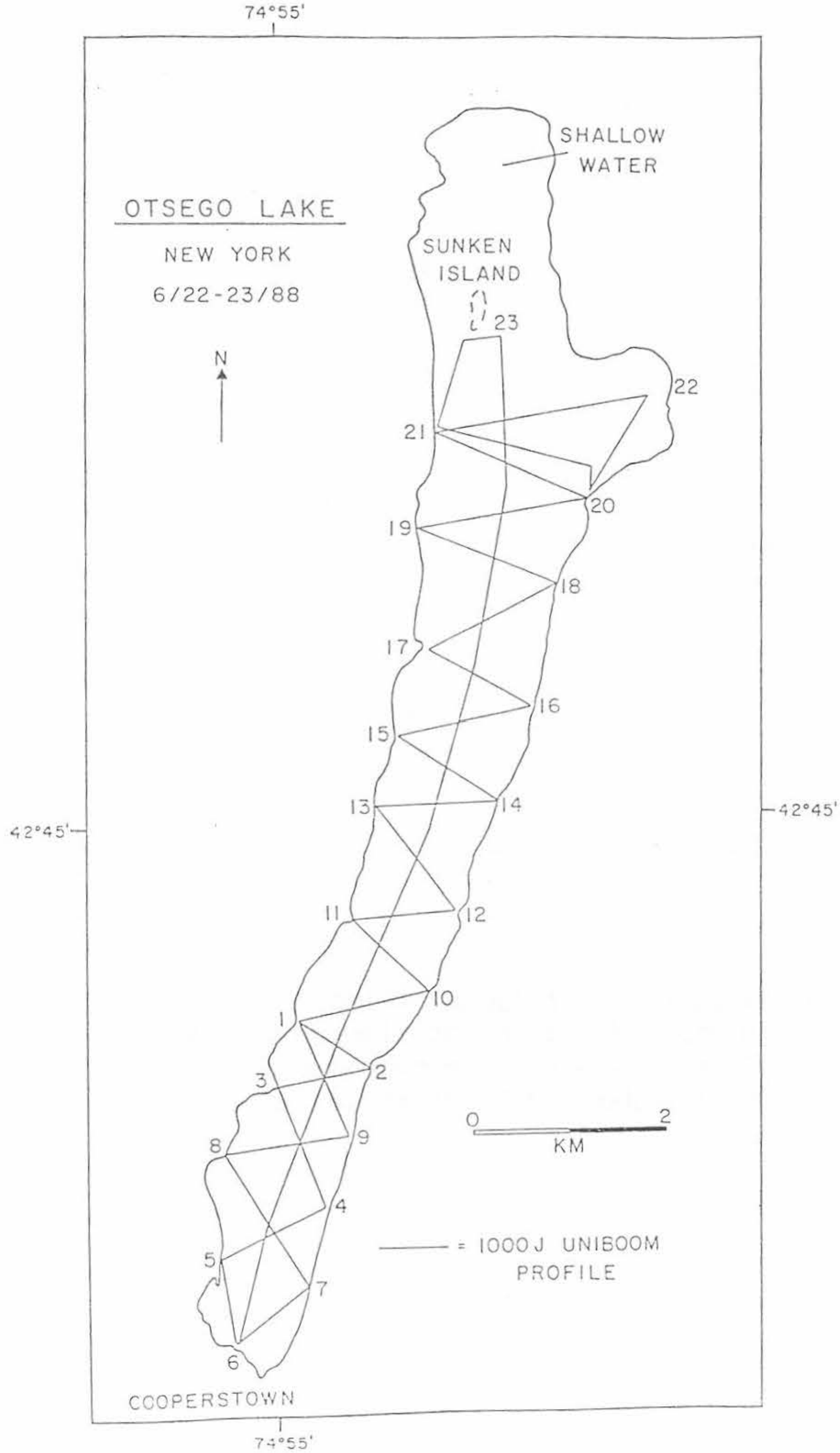

Figure 5 Index map of seismic traverses. Trackline map illustrating location of uniboom seismic reflection profiles. (from Fleisher, et al, 1990)

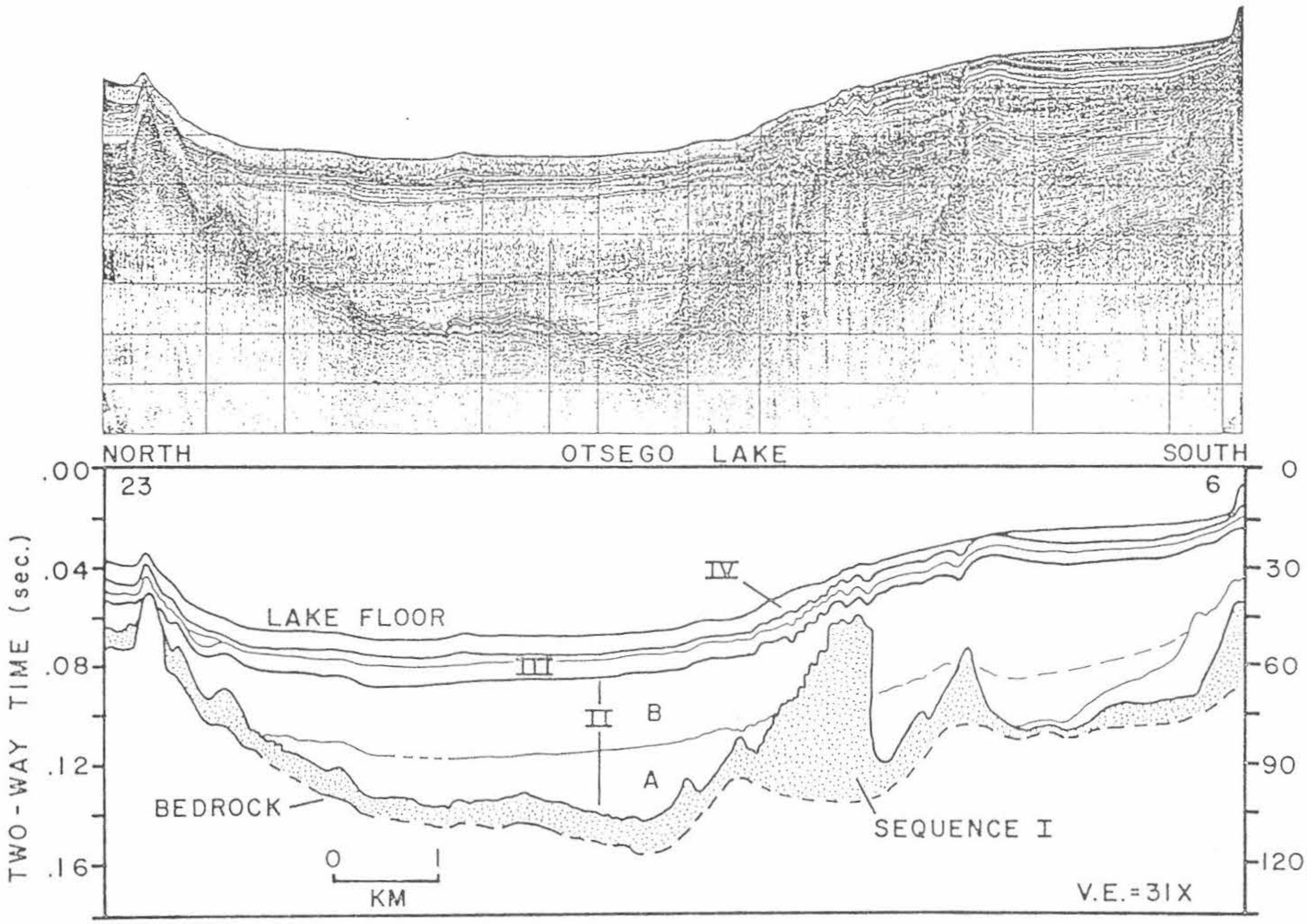

Figure 6 Axial profile. Photographic (top) and line drawing interpretation (bottom) of axial seismic reflection profiles from Otsego Lake. Roman numerials label depositional sequences. Vertical exaggeration = X31. (from Fleisher, et al, 1990)

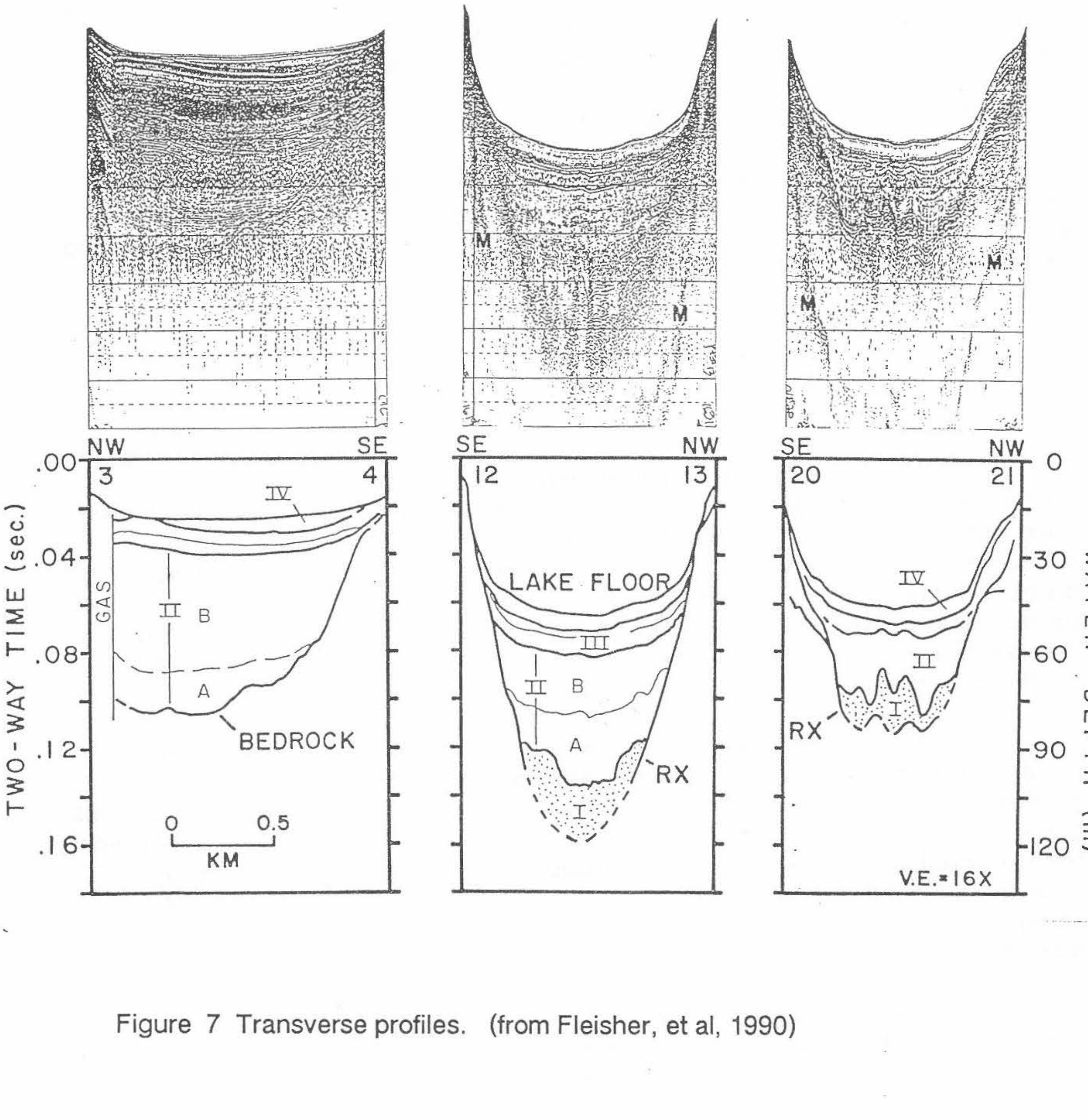

Figure 7 Transverse profiles. (from Fleisher, et al, 1990)

Figure 6 is the north-south axial line and Figure 7 contains transverse examples from the southern, central and northern portions of the lake. From these data measurements of water depth, sediment thickness and depth to bedrock were made. Figure 6 includes interpretations of depositional sequences within the sediment column (Fleisher, Mullins and Yuretich, 1990).

A maximum of 88 m of Quaternary sediment rests upon the bedrock basin of Otsego Lake. The thickest sediments are found in the southern-third of the lake basin and thin by more than 50% to the north. This information correlates with water-well log data that shown 55 m of bouldery silt beneath the Doubleday Margin.

Bedrock lies as much as l32 m below the modern lake level in the central portion of the basin. Bedrock rises at both ends of the lake to define a closed basin.

OTSEGO LAKE STRATIGRAPHY

Based on a combination of reflector terminations and vertical variations in seismic facies, Four depositional sequences have been identified within the sediment column of Otsego Lake (Fleisher, Mullins and Yuretich, 1990). The oldest unit, sequence I, has a seismic character interpreted to be coarse-grained, poorly sorted diamict. It is relatively thin (20 m) throughout much of the basin but forms three distinct ridges (possibly eskers or suballuvial fans) up to 60+ m thick in the southern third of the lake basin. Sequence I rises above lake level beneath Cooperstown to form the modern dam for Otsego Lake.

Sequence II is the thickest and appears to dip slightly northward in Figure 5. It is subdivided into two sub-sequences, IIA and IIB. Sequence IIA is characterized by high-frequency continuous reflections, whereas sequence IIB displays a seismic facies change that grades from continuous reflections in the south to a transparent (reflection-free) seismic facies northward. The continuous, high-frequency reflections suggest wide-spread, but temporally variable deposits, such as rhythmites, whereas the transparent seismic facies implies massive, non-stratified sediments of uniform texture that may have accumulated very rapidly.

Sequence III is a relatively thin unit that maintains a more or less uniform thickness along the north-south axial profile. Sequences II and III account for as much as 68 m of the total 88 m (77%) of sediment fill within Otsego Lake.

Sequence IV, the youngest, is not entirely continuous and generally appears as a layer less than l0 m thick. It is thickest along the flat lake floor, where it is about 6 m thick. This sequence is interpreted to be a massive, unstratified deposits, and represents modern (post-glacial) lacustrine deposition.

ANALYSIS OF DIP DIRECTION

Apparent dips measured on crossing and axial profiles were resolved at profile intersections for true dip direction through the use of the standard Schmidt net technique. The true dip is less than 2.2 degrees (most less than 1.0 degree), and vary only slightly from place-to-place. Generally, magnitude of dip appears to diminish up-section and northward.

A conspicuous eastward dip is shown on several transverse profiles (Figure 8). Analysis of dip direction in Sequence IIA and B indicates azimuths lie between N 79 E and N 88 E in profiles 1-2 and 2-3 in the vicinity of Brookwood Point and Leatherstocking Falls. Similarly, the dip direction in Sequence IIB along profiles 10-11, 11-12 and 12-13 lies between N 74 E and S 75 E northward of Threemile Point for more than a km . Consistent eastward dips continue in IIB south of Brookwood Point in profiles 4-5 and 8-9, where azimuths range from N 84 E to S 68 E, all of which is summarized in Figure 9. Furthermore, bedding-plane troughs marking the deepest part of the depositional basin shift eastward in progressively younger layers. These data indicate a consistent sediment source from the west, which is taken as evidence for tributary inflow. Several unnamed streams occupy drainage basin on the western valley wall and appear to have supplied a source of sediment to the southern-third of the lake basin.

A less conspicuous eastward dip direction is present in the central portion of the basin between Threemile and Fivemile Points along profiles 12-13, 13-14 and 14-15. Here, stratification is well represented in Sequence A, while IIB lacks bedding. This condition continues northward, where general symmetry is seen in profiles 15-16 and 16-17 within 1 km north and south of Fivemile Point. Here, the axial profile reflects the true dip of northward onlap in Sequence A (Figure 6) indicating continued sediment movement and infill from the south. Bedding symmetry within crossing lines in the vicinity of Fivemile Point suggests infill from Shadow Brook via Hyde Bay equaled that from streams flowing from western slopes.

Asymmetry is seen again in the northern part of the lake with increasing significance from Fivemile Point northward along profiles 17-18, 18-19, 19-20 and 20-22. However, dips here favor inclination to the west, which indicates sediment was primarily derived from Shadow Brook to the northeast.

INTERPRETATION OF SEISMIC RECORD

Seismic character and pattern of occurrence suggest sequence I is a late glacial diamict implaced at the base of a retreating ice-cliff. Deposition is thought to have involved sediment flow through subglacial conduits and channels that delivered suspended silt to the ice-contact, proglacial lake environment of Glacial Lake Cooperstown. Fountain discharge driven by englacial hydrostatic head (Gustavson and Boothroyd, 1987) created local thickening above bedrock pinning points in the southern part of the basin.

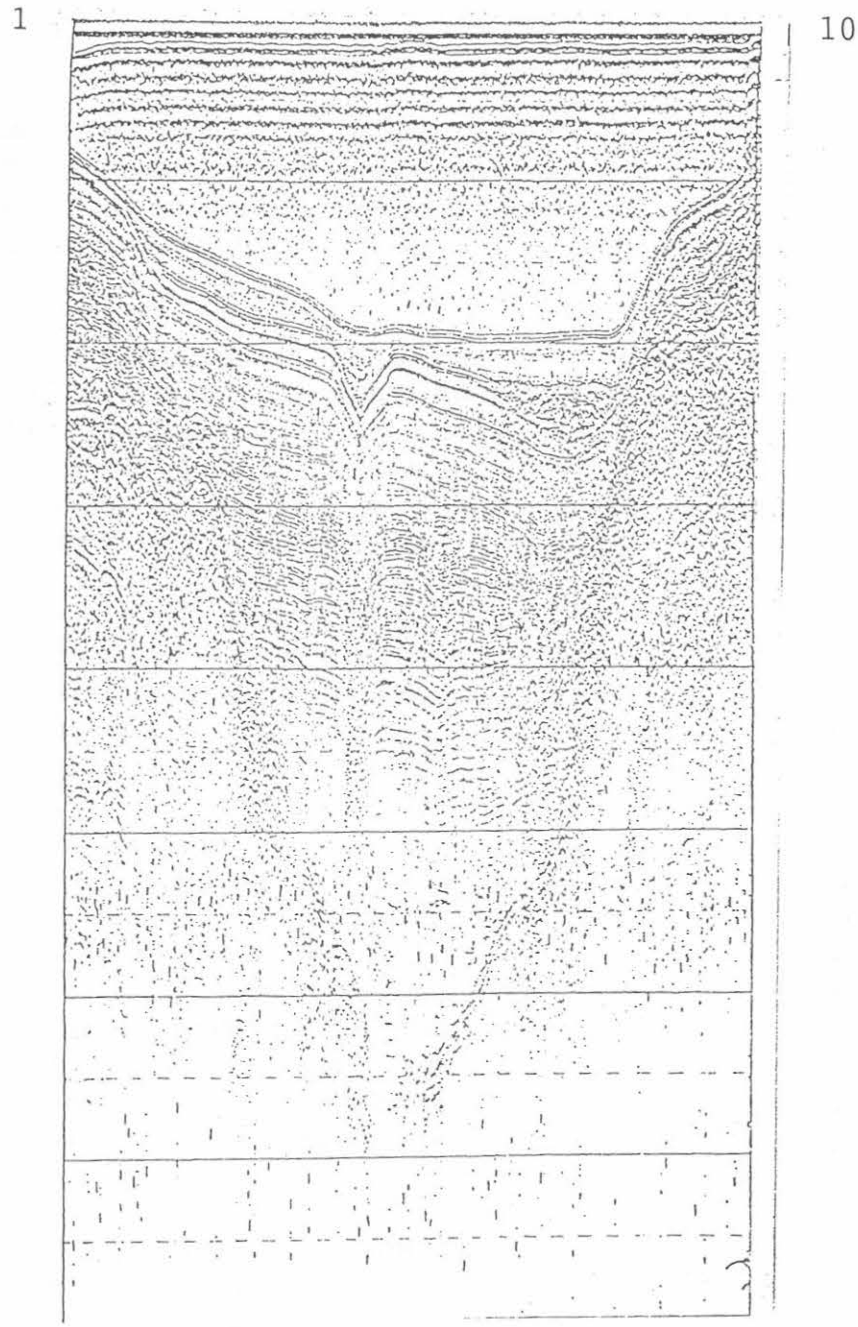

Figure 8 Asymmetric bedding. Transverse profile between stations 1 and 10 shows conspicuous eastward dip direction. Vertical exaggeration = X16. (from Fleisher, et al, 1990)

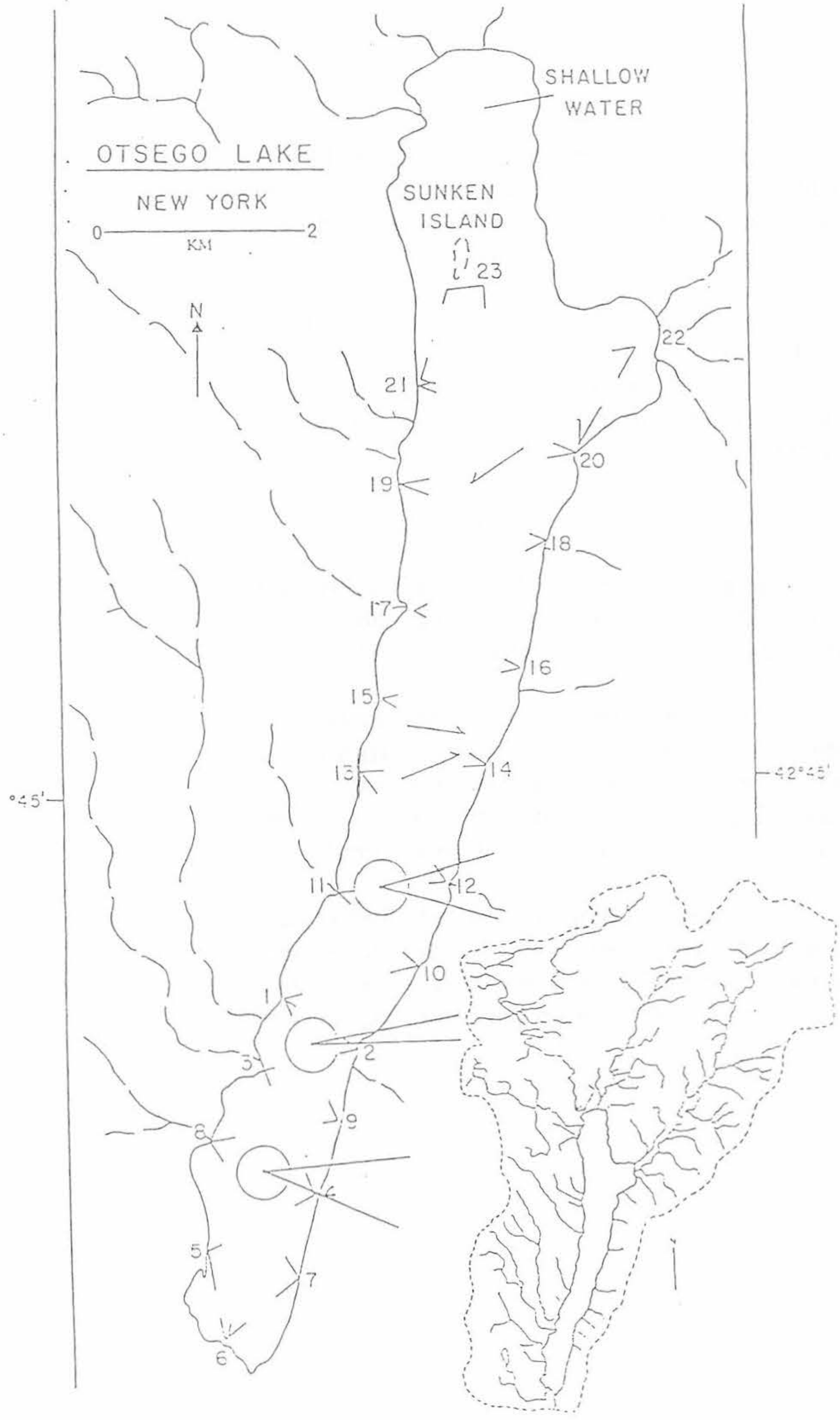

Figure 9 Dip direction for sequence II and tributary inflow sources. Small arrows indicate true dip direction calculated from apparent dips shown on seismic profiles. Data from two or more crossing profiles are summarized in "pie segments". Inset map shows asymmetry of drainge basin. (from Fleisher, et al, 1990)

Sequence II is attributed to sediment transport mechanisms that involved turbidity flows, suballuvial flows and debris flows, possibly related to flashy tributary streams, associated with extremely rapid silt sedimentation in a highly turbid ice-contact lake.

Thickening of A and B in the southern portion of the basin appears associated with tributary inflow that initially introduced sediment from the west, and was followed by northward bottom flow around "highs" in Sequence I. Disturbed bedding in sequence II is thought to be related to subsidence, collapse and compaction of glacial and post-glacial units.

CONCLUSIONS

Glacial Lake Cooperstown was dammed at 1250' by the Doubleday ice-margin at Cooperstown, not behind the Cassville-Cooperstown moraine 5 kilometers downvalley, as originally postulated by Fleisher (1977).

Depositional landforms and stratigraphy between the Cassville-Cooperstown margin at Index and the Doubleday margin at Cooperstown are inconsistent with standard interpretations of active ice retreat (fast or slow). Instead, the anomalous features suggest stagnation and inwash. This report suggests that retreat included the formation of a marginal-ice cleat and ultimate development of a dead-ice sink.

The eastward dipping beds of Glacial Lake Cooperstown indicate fluvial sources to the west. However, the primary source of silt in sequence II is thought to have been subglacial sediment flow in basal conduits that discharged directly into the lacustrine environment from the retreating ice-tongue terminus.

Case #2: Remnant ice as base level controls for stratified drift aggradation: Otego Creek, New York

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State University of New York, College at Oneonta

INTRODUCTION

The Otego Creek valley is a non-through valley with a glacial landform assemblage typical of downwasted ice, which includes kame fields, discontinuous gravel plain remnants, dead-ice sinks and eskers (Fleisher, 1991). Morrow (1989) and Fleisher (1986a, 1986b) described the processes responsible for these landforms.

Morrow's mapping units include ablation moraines (equivalent to kame field), outwash and kame deltas (similar to gravel plain remnants), local lake plains, dead-ice sinks, alluvial fans, an esker, and modern flood plain as shown in Figure 1.

Morrow's map is used to identify and locate upper-level planar landforms. The purpose of this study is to display and describe their longitudinal gradients and postulate the base level to which they are graded. The possibility that remnant ice masses served as temporary base levels during ice-tongue collapse is considered.

METHOD

Once planar landforms were identified on 7.5 minute quadrangles, longitudinal projected profiles were constructed within the stream valley between the hamlets of Laurens and Hartwick, a distance of 9.7 miles (Figure 2). The profile represents the highest planar landforms along a strip consisting of contiguous 1000-foot wide Block (pixels) aligned along the general valley thalweg. The maximum elevations of planar landforms, with a minimum dimension of 1000 feet within each block, were plotted at the centers of each block and used to construct the profile. Otego Creek was also plotted.

DISCUSSION OF OBSERVATIONS

Three general elevation groups were observed: 1140-1160' between Laurens and just south of Blood Mills Road, 1200-1260' from Blood Mills Road to about 1500' north of Laurens town line, and 1300-1340' from Jones Crossing to Hartwick (alluvial fans and patchy kame moraines were not included) (see Figure 1).

The profile lends itself two possible interpretations. The first involves a single profile (Figure 2A) on which all high planar surfaces fall. This is graded to an elevation of 1140 +/- 10', which corresponds to Glacial Lake Otego within the Susquehanna Valley at the mouth of the Otego Creek (Melia, 1975; Fleisher, 1977). The second (Figure 2B) suggests the projection represents three different profile segments at 1300-1340', 1200-1260', and 1150-1160', each graded to different local base levels. The

FIGURE 1. QUATERNARY LANDFORMS, MT. VISION QUADRANGLE (FROM MORROW, 1989)

Surficial Mapping Units

Abm	Ablation Moraine: Extensive, hummocky terrain, not associated with massive outwash. Consists of silt, sand and gravel in both well sorted and poorly sorted intervals. Potential for large scale aquifer development is variable depending on permeability and compaction.
Ow	Outwash: Well sorted, stratified sand and gravel, deposited by proglacial fluvial action. Contains a variable amount of silt, but is generally very permeable and has high potential for aquifer development.
kd	Kame Delta: Stratified sand and gravel that has been deposited into a proglacial or ice contact lake and has internal deltaic structures. Very permeable and has high potential for aquifer development.
lp	Lake plain: Primarily deposits of sand, silt and/or clay deposited in a lacustrine environment. Poor potential for aquifer development.
DIS	Dead-ice sink: Variable texture (size and sorting), identified as an anomalously broad flood plain that is bounded by high outwash and/or ablation moraines. These confining units often show collapse structures adjacent to the sink.
t/r	Till/bedrock: Till (e.g. clay, silt-clay, boulder clay), resting on bedrock, thickness variable.
mfp	Modern flood plain: Post-glacial flood plain deposits of silt, low permeability, thickness 1-5 meters.
Alf	Alluvial fan: Fan shaped accumulations of poorly stratified silt, sand and boulders, at the foot of steep slopes, generally permeable.

NOTE: Where two units are separated by a slash, the first unit overlies the second. Example: afp/DIS, modern flood plain deposits overlie a dead-ice sink.

Map Symbols

Esker

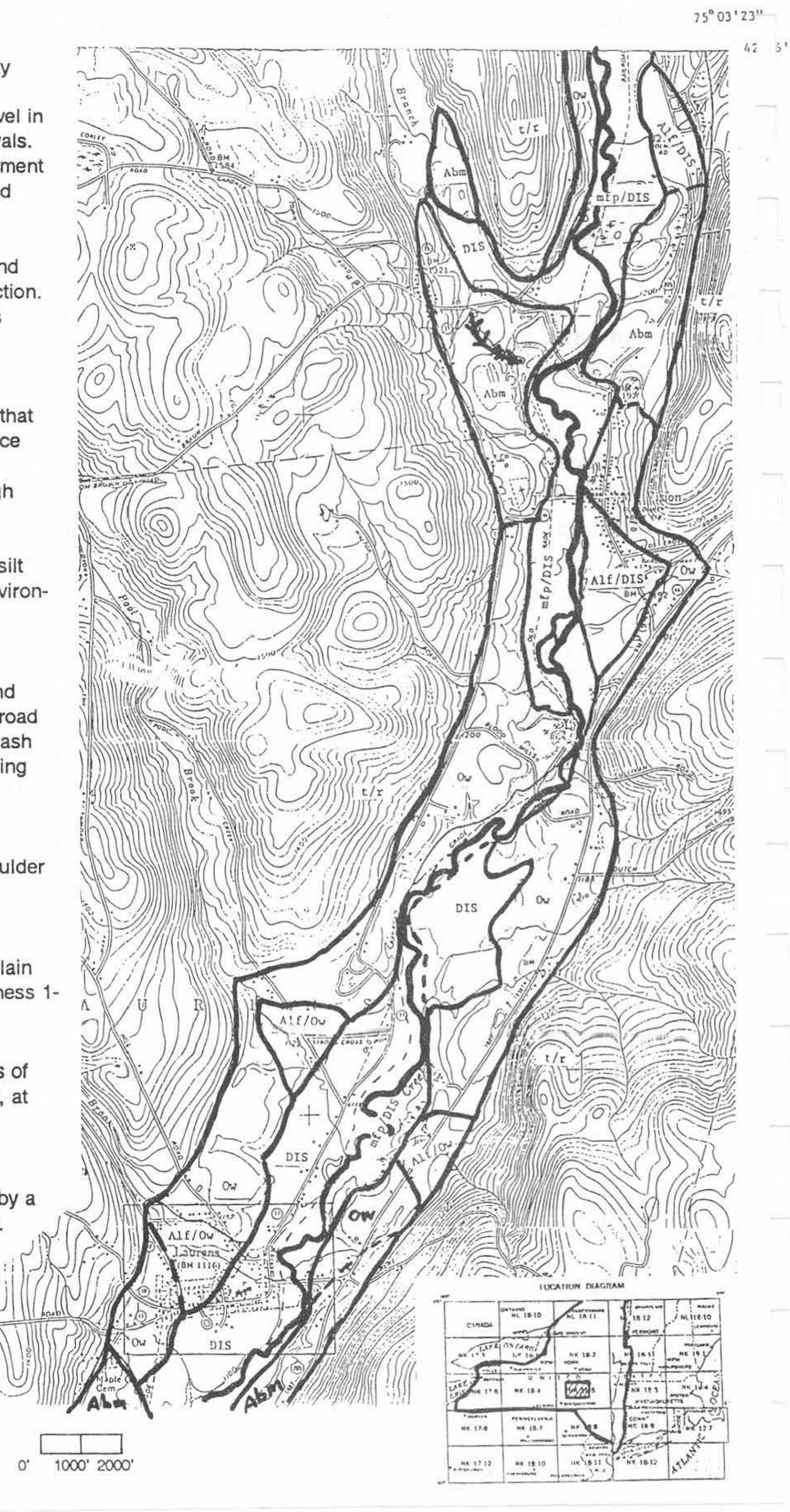
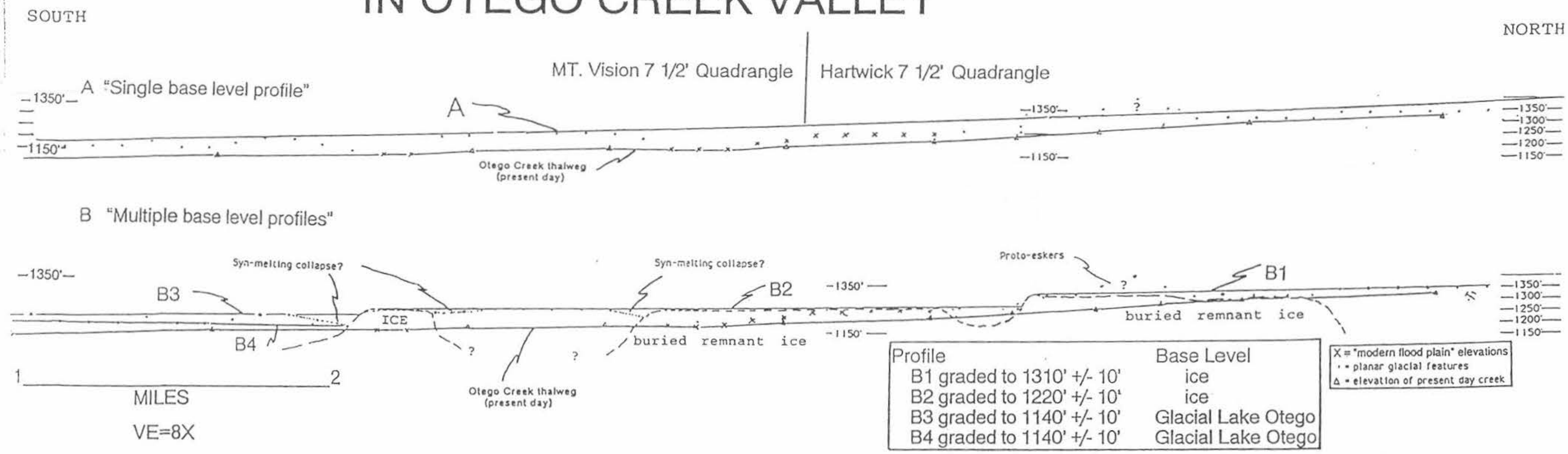


FIGURE 2. PROJECTED PROFILES: PLANAR FEATURES IN OTEGO CREEK VALLEY



surfaces represented by these profiles do not resemble a "shingled sequences" configuration, as presented by Mullholland (1982) in New England or morphosequences similar to those suggested by Koteff (1974). An alternative of local application is that several remnant ice masses occupied the valley at various times during ice-tongue collapse and obstructed meltwater movement during the deposition of planar sand and gravel features.

Air photo observations support the occurrence of one or more small-scale sinuous gravel ridges (proto-eskers) between Jones Crossing and Hartwick. This supports the notion of englacial fluvial transport within stagnant and downwasting ice.

CONCLUSIONS

The best fit for the data suggests that detached remnant ice masses impeded meltwater flow, which in turn formed ponds and lakes into which deltaic terraces were deposited. Problems with this model stem from the need to develop a time-frame which would account for both the sequence of deposition and source of meltwater necessary to move the sediment. It is also suggested that englacial and subglacial tunnels and conduits provided avenues for sediment transport.

Case #3 Inwash sediment sources, Valley of Charlotte Creek, central New York

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INTRODUCTION

Remnant-ice landforms and the moraine at West Davenport are thought to have developed in association with ice buried beneath inwash. Here, it appears as though tributary inwash from Kortright Creek provided significant supraglacial aggradation to bury large segments of the valley ice-tongue and produce extensive ice-cored deposits that ultimately downwasted in place.

GLACIAL DEPOSITS OF CHARLOTTE CREEK VALLEY

Glacial landforms and stratigraphy of the Charlotte Creek valley (Figure 1) are consistent with a regional pattern seen elsewhere throughout central New York. Deglaciation was characterized by retreating ice-tongues in valleys dammed by moraines, ice-cored inwash and thickly aggraded valley train (Fleisher P.J., et al., 1990). Consequently, many ice-contact lakes formed and their associated deposits are well developed and widely distributed. Such is the case for Charlotte Creek valley.

Four general types of depositional landforms shown in Figure 2 are:
1. Moraine and Pitted Valley Train
2. Dead-ice Sink Complex
3. Pitted Hanging Deltas
4. Lacustrine Plain

Moraine and Pitted Valley Train (Outwash/Inwash)

The moraine at West Davenport spans the full valley width at a maximum elevation of 1280' and is breached by Charlotte Creek at a floodplain elevation of 1180' through the moraine. Local relief on the kame and kettle topography of the moraine is about 60'. Contiguous with the moraine to the west is a dissected, valley train at a general elevation of 1220'. It is pitted by shallow kettles and spotted by occasional kames. The floodplain of Charlotte Creek is 60' lower, widening westward to its confluence with the Susquehanna. A north-south profile across the valley and through the moraine illustrates the association of landforms with stratigraphy at depth (shown in Figure 3), and shows no indication of a break in the depositional record. Therefore, these deposits, as well as those of the entire valley, are thought to be the product of a single deglacial event.

Dead-ice Sink Complex

The term "dead-ice sink" (Fleisher,1986) describes a landform similar to a kettle, but of much larger scale. As with a kettle, a block of ice is buried within valley-floor

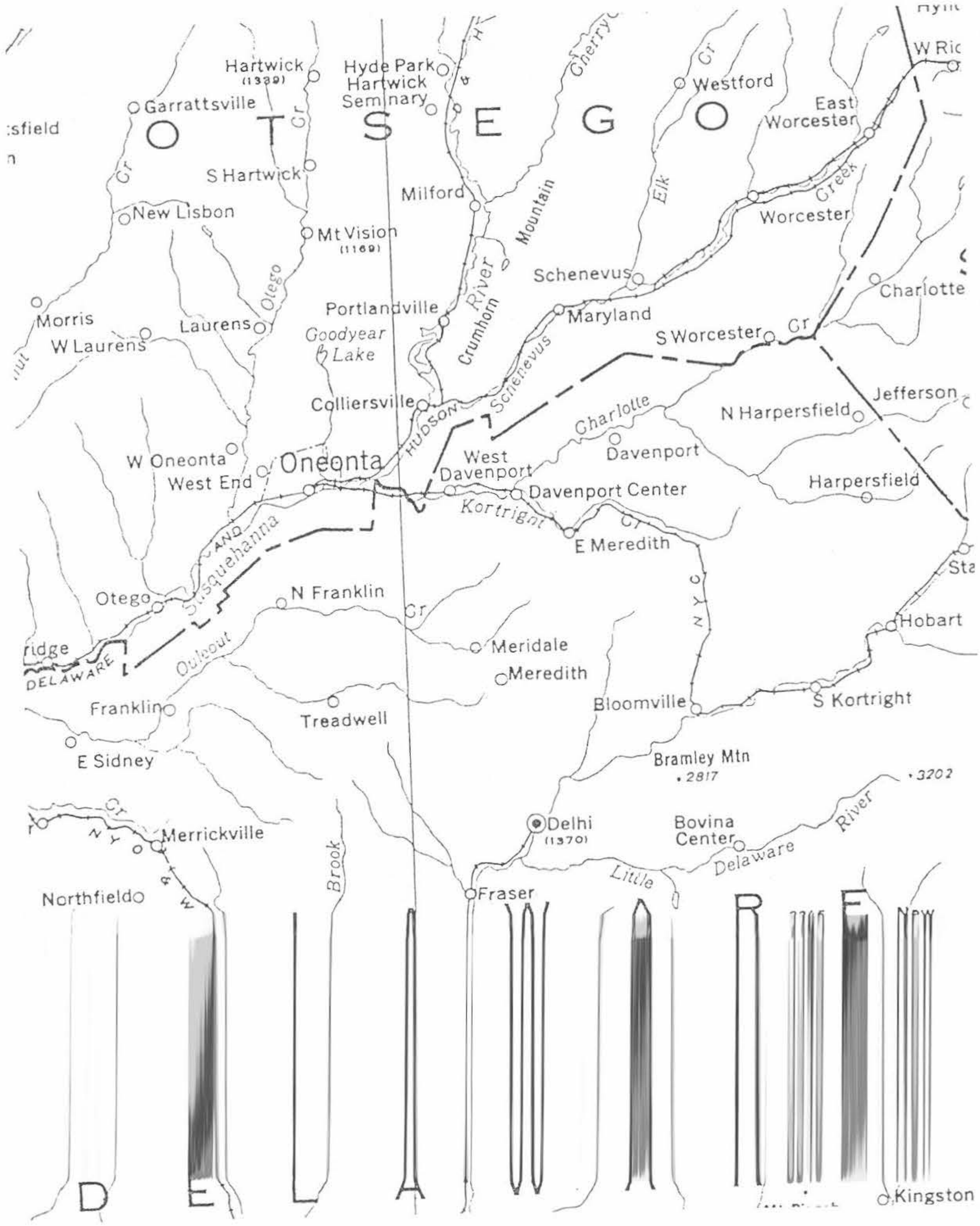

FIGURE 1 Index Map. The West Davenport/Davenport Center area is located near the mouth of Charlotte Creek Valley, 4 miles upstream from the Susquehanna confluence.

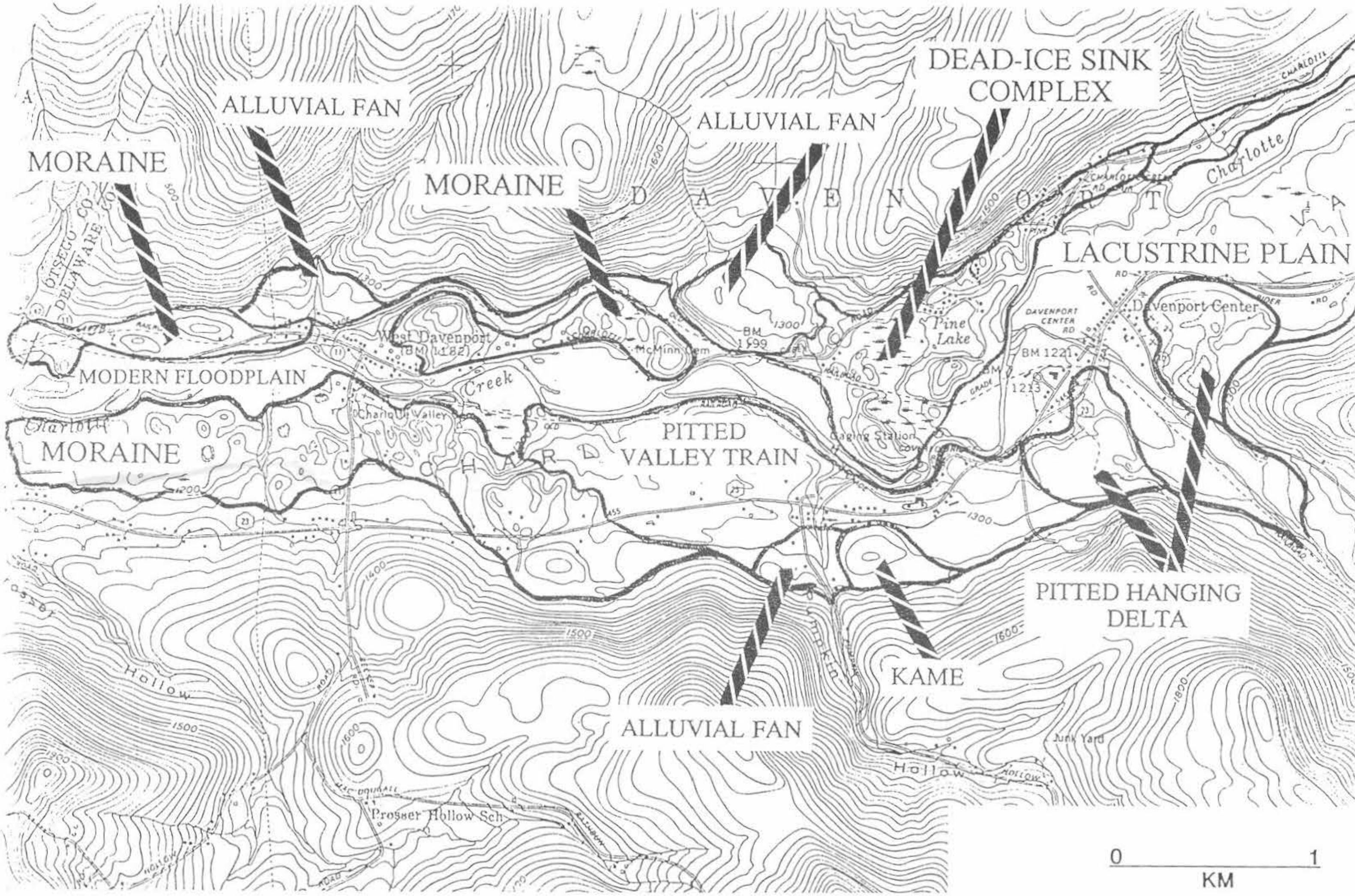

FIGURE 2 Map of Glacial Landforms. Landforms in the West Davenport/ Davenport Center area between the moraine at West Davenport and lacustrine plain of Glacial Lake Davenport.

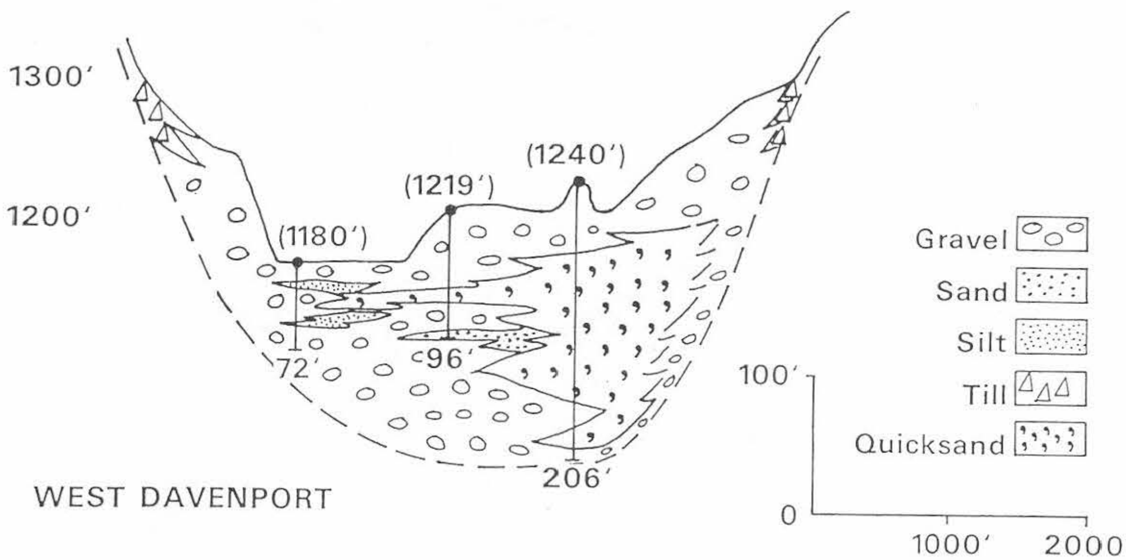

FIGURE 3 Cross section of Quaternary deposits within moraine at West Davenport. Upvalley view of interstratified gravel, sand and "quicksand" (coarse silt). Well data from Randall, 1972.

material (usually stratified drift) and ultimately melts to create an exceptionally large closed depression. However, the distinction between a kettle and a sink lies in the size of the resulting depression. Dead-ice sinks commonly occupy a significant portion of the entire valley floor and are recognized by anomalously broad floodplains within otherwise continuous valley trains and/or paired outwash terraces. Deglacial conditions leading to sink formation involve the detachment of a large ice mass from the retreating ice-tongue, with subsequent burial beneath a valley train of outwash and inwash deposits. Retarded melting due to insulation by overlying debris results in the slow development of a valley-floor depression in which highly diverse deposits accumulate. Where more than one large ice mass is involved, a complex of juxtaposed sinks and kames form. A dead-ice sink complex at Pine Lake lies between the West Davenport moraine and Davenport Center lake plain. Here, local relief exceeds 100' and pitted valley train (outwash/inwash) occurs along the south side of the valley adjacent to the sink complex at elevations of 1280' to 1300'. The floodplain forms the floor of the sink at 1200' (BM on abandoned railroad bridge at 1213').

Associated with the moraine/dead-ice sink complex are several landforms at elevations between 1300-1380'. These include an incised hanging delta/fan complex at 1300' on north side of the valley between West Davenport and Pine Lake, an alluvial fan remnant at 1360' at the mouth of Pumpkin Hollow on the south side of the valley and a dissected alluvial fan at 1260-1280' on the north side of the valley at West Davenport. These are interpreted to be indicative of significant inwash accumulation.

Pitted Hanging Delta

Possibly the most significant landform within the entire valley is the pitted hanging delta that stands at 1280' at the mouth of Kortright Creek in Davenport Center. Comparable elevations on the moraine downvalley suggest the moraine served as the dam for Glacial Lake Davenport (Fleisher, 1977), into which the delta grew. More importantly, the surface of the delta contains several distinct kettles that are in excess of 40' deep. This indicates the delta aggraded onto and across grounded ice in Glacial Lake Davenport immediately adjacent to the dead-ice sink complex and that remnant-ice and ice-cored drift accumulated within the valley while upland slopes were ice-free. These ice-contact, alluvial deposits are interpreted as inwash from upland tributaries.

Another pitted hanging delta can be seen at Butts Corner, where Middle Creek enters Charlotte Creek valley. Borrow pit excavations reveal 20-30' of alluvium from Middle Creek covers deltaic gravels containing a topset/foreset contact near 1320', thereby establishing a downvalley spillway at 1300-1320'. Smaller hanging deltas at similar elevations occur elsewhere within the valley. East of Butts Corner a variety of landforms appear to have served as local dams for ponding upvalley as suggested by hanging deltas that rise in steps through Fergusonville, Simpsonville and South Worcester. Remnant-ice masses may have also provided temporary base levels for scattered strandline deposits.

Lacustrine Plain

A semi-continuous lacustrine plain rises from a valley floor elevation of 1210' at Davenport Center to 1240' at Davenport 2 miles to the east. In the main Charlotte Creek valley, the lacustrine plain is interrupted by various forms of outwash, inwash and previously ice-cored, landforms, which appear as subdued, gently rolling terrain 20-40' above the lacustrine plain and a comparable elevation below a few valley train remnants. The upvalley floodplain climbs to 1280' at Fergusonville, 1320' at Simpsonville and 1400' in South Worcester. These abrupt rises support the notion of a lake and pond chain within the valley. In some places the lacustrine plain is covered by modern floodplain sediments five feet thick, as seen in archaeology dig site excavations a few hundred meters south of Pine Lake (Fleisher, 1990, unpublished report).

CHRONOLOGY OF GLACIAL EVENTS

Summarized below are the events interpreted to have occurred in the vicinity of Pine Lake during deglaciation, beginning with late glacial and continuing into post-glacial time.

Glacial Time

Deglaciation of Charlotte Creek valley pre-dated full retreat of the Susquehanna valley ice-tongue. Separated and detached ice masses remained within the valley long after upland slopes and tributary valleys were ice-free. The Susquehanna ice-tongue and associated ice-cored deposits dammed the mouth of Charlotte Creek valley and impounded an early phase of Glacial Lake Davenport at 1300-1320'. Graded to this elevation are hanging deltas near West Davenport, Butts Corner and Fergusonville. Dissected remnants of a delta-like feature at 1300' extend across the valley at the mouth of Dona Brook and may have at one time divided Lake Davenport into eastern and western segments.

Inwash from Kortright Creek buried remnant ice masses at Davenport Center and effectively covered the downvalley ice-tongue to the west. This contributed to the debris mass that ultimately formed the moraine at West Davenport .

Late-Glacial Time

As the Susquehanna valley ice-tongue retreated from its position behind the Oneonta Moraine, an ice-cored dam remained at the mouth of Charlotte Creek buried beneath inwash primarily from Kortright Creek (Figure 4). In adjustment to local base level changes within the Susquehanna, the spillway for Lake Davenport was lowered to 1280', where it remained while several deltas formed, including the ice-cored delta at Davenport Center.

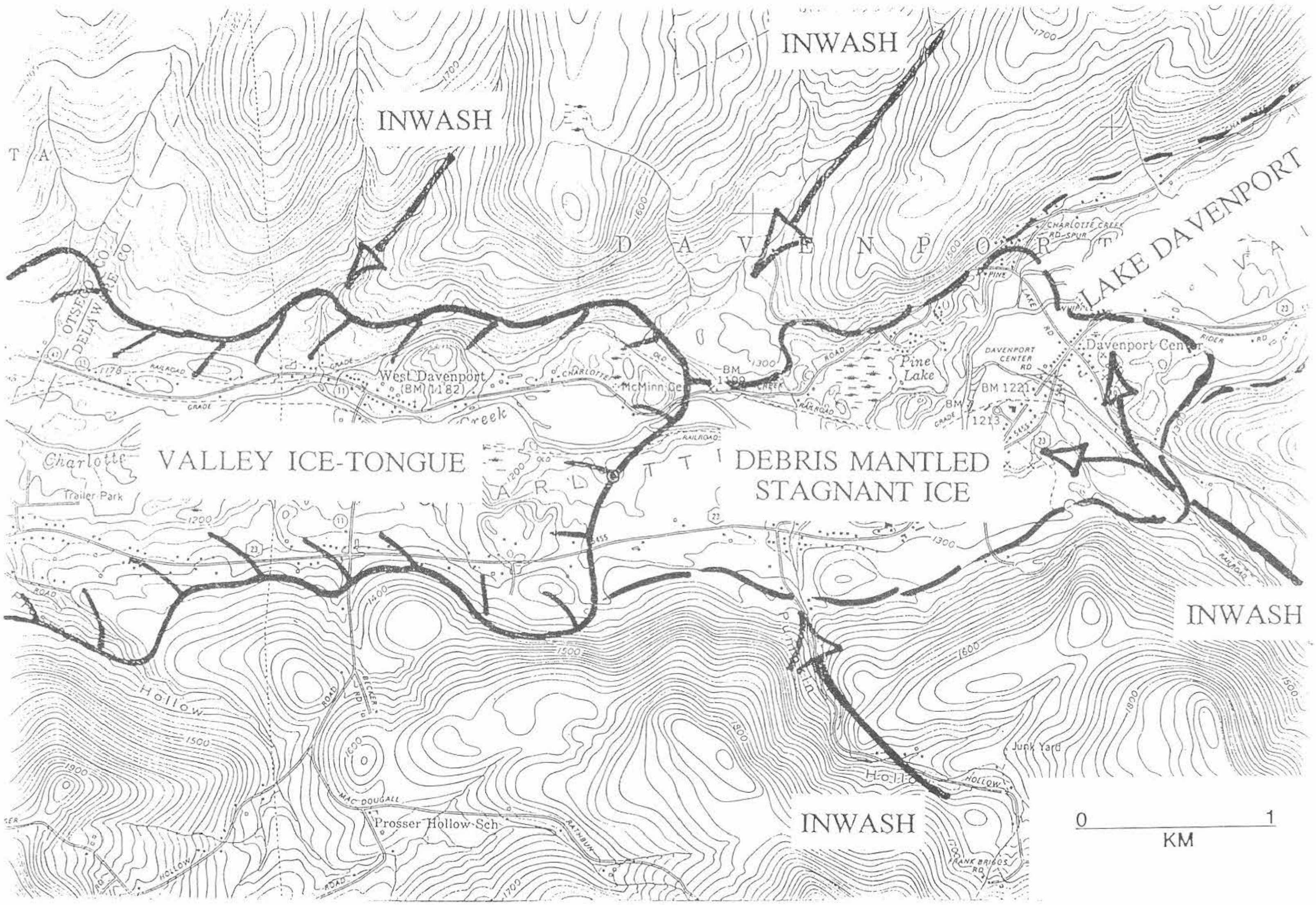

FIGURE 4 Late Glacial Retreat Configuration.

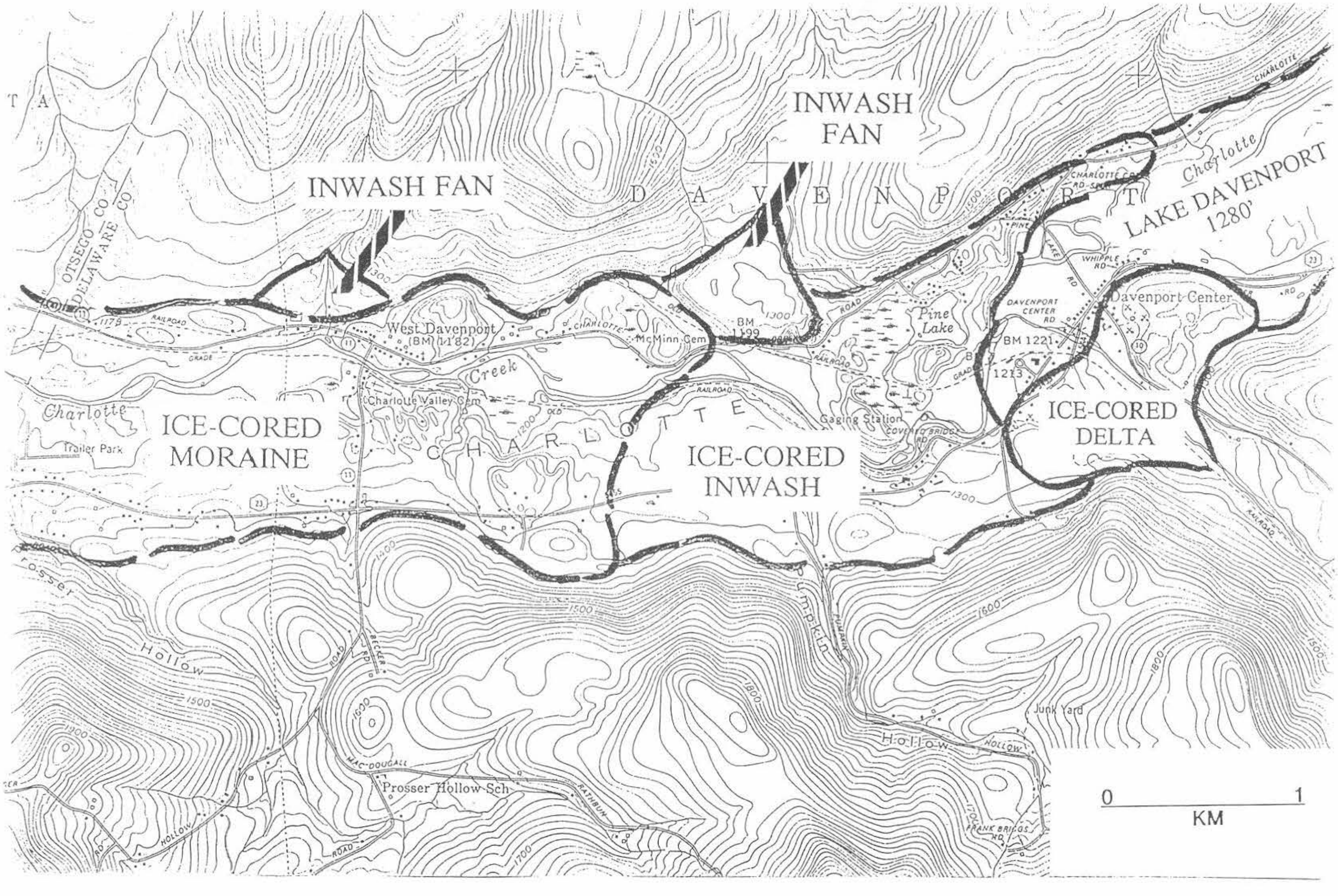

FIGURE 5 Early Post-Glacial Deposits.

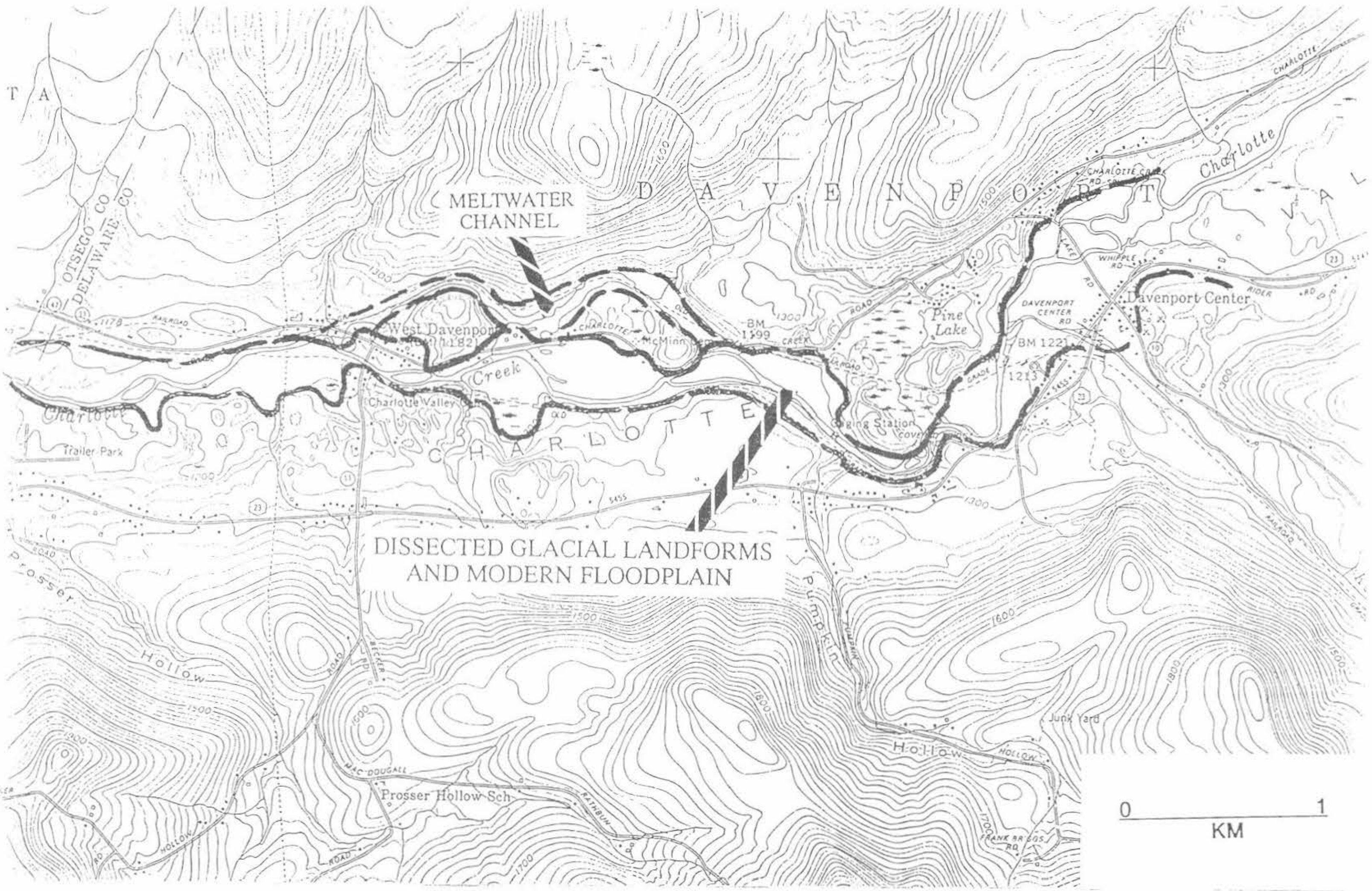

FIGURE 6 Holocene Dissection

Downwasting of all ice-cored material progressed, while the moraine and sink began to develop. As the spillway elevation was lowered, Lake Davenport emptied, as delta/fan complexes were incised and the lacustrine plain emerged.

Early Post-Glacial Time

Retarded melting of partially buried ice masses eventually produced a variety of collapse features in association with the newly-formed lacustrine plain. The terrain at Pine Lake began to form as downwasting lead to the development of pitted valley train, kame and kettle topography and a dead-ice sink complex (Figure 5). These landforms developed during this period of time as downwasting slowly progressed and ice-cored valley deposits were lowered to river level, as Holocene sedimentation kept pace with subsidence.

Holocene Time

As Charlotte Creek established its course on the newly-formed valley floor, floodplain aggradation began (Figure 6). Lateral planation trimmed the edges of kames and terraces, as overbank discharge spread blankets of silt on glacial gravel and lacustrine silt, sand and clay. Through the course of aggradation and meandering, the creek re-occupied some locations several times and floodplain sediment accumulated. Silt layers and channel sands were observed in trench excavations at the Pine Lake archaeology dig of 1990.

APPENDIX A

PEBBLE COUNTS IN THE CHARLOTTE CREEK DRAINAGE BASIN

Steve Carley and P. J. Fleisher, Earth Sciences Department, SUNY-Oneonta

INTRODUCTION

In a paper dealing with the geology of the Charlotte Creek Valley, Fleisher (1980) interpreted many of the landforms as inwash features formed in association with remnant ice. Ice-free uplands provided a source for alluvium washed onto remnant ice blocks between Davenport Center and West Davenport (Figure 1). Subsequent downwasting formed a moraine and dead-ice sink complex. A major source of inwash is thought to have been Kortright Creek. This study tests the usefulness of pebble count data to determine sediment provenance in an area where exotic pebbles comprise less than 5% of the total sample. If this method of analysis is valid for samples such as these, then the source of the gravels associated with the dead ice complex and the moraine may be substantiated. Halter, et al., 1984, suggest that sediment transported by Laurentide ice in Maine was deposited within 10 kilometers of its source. If this applies to the Appalachian Plateau as well, a high percentage of exotics beyond 10 km would suggest transport mechanisms other than glacial were significant.

The possibility of selective lithologic attrition by weathering or transport mechanisms is important in this area because susceptible siltstone and shale are common. This is an important factor when evaluating the meaning of pebble count data.

PEBBLE COUNT SAMPLE SITES (Figure 2)

Kortright Creek

The modern stream flows over glacial deposits from the valley head to the village of East Meredith, where it encounters a bedrock knick point. All samples taken in this valley (Sample 1-9) were obtained upstream from the bedrock knick point from holes dug in the drift.

Pitted Hanging Delta

Kettles in the surface of this delta and prominent foreset beds indicate that it prograded from the Kortright Creek Valley, across grounded ice, and into glacial Lake Davenport (Fleisher, 1990). The upper elevation of the dissected delta lies at 1280 feet, 40 feet above the modern creek. Sample 10 was taken from an active borrow pit within the core of the delta adjacent to country Route 10 along Kortright Creek. Sample 11 was taken from foreset beds in an active borrow pit on the distal northern side of the delta a few hundred yards from sample 10.

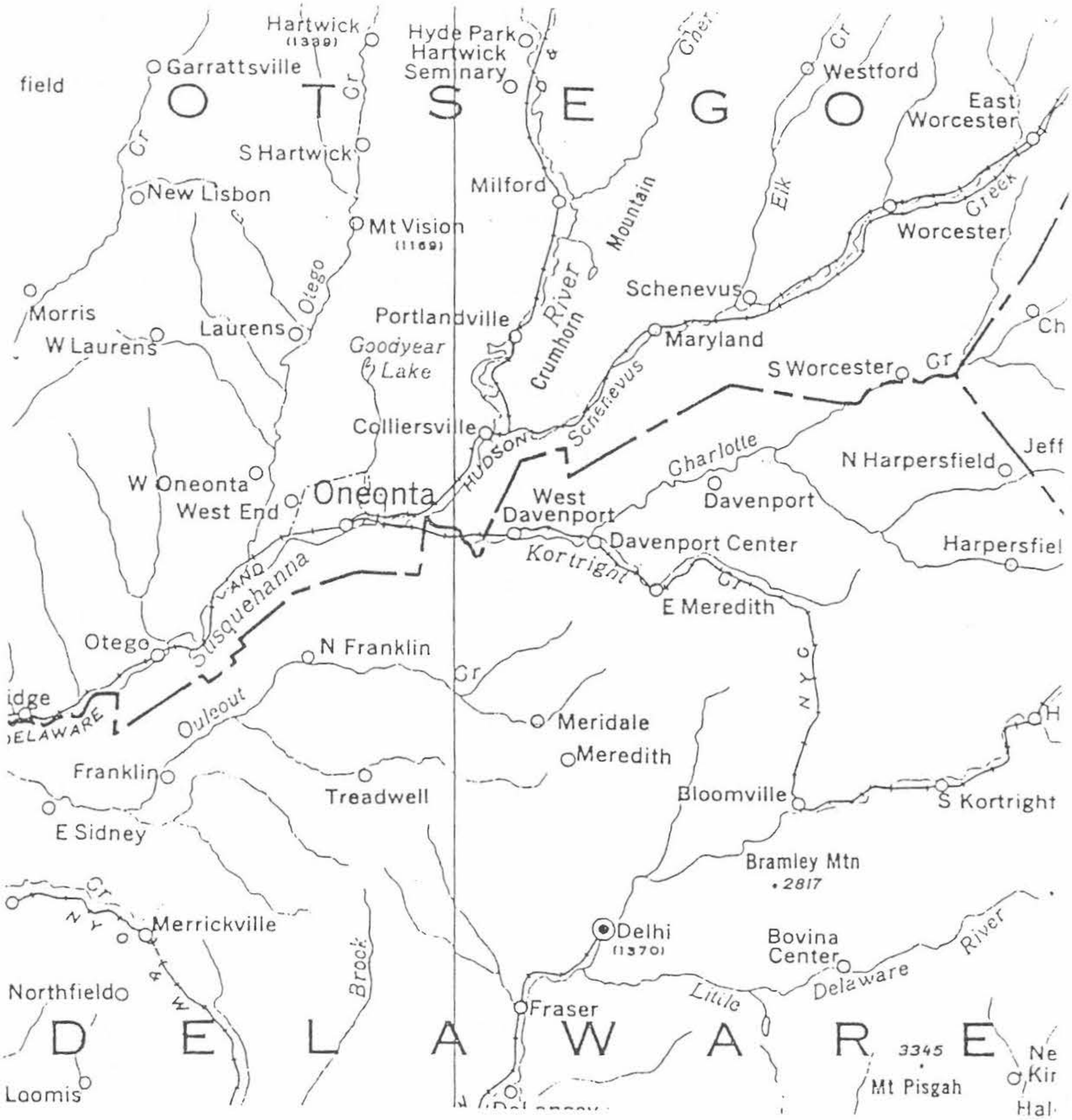

Figure 1 Index Map

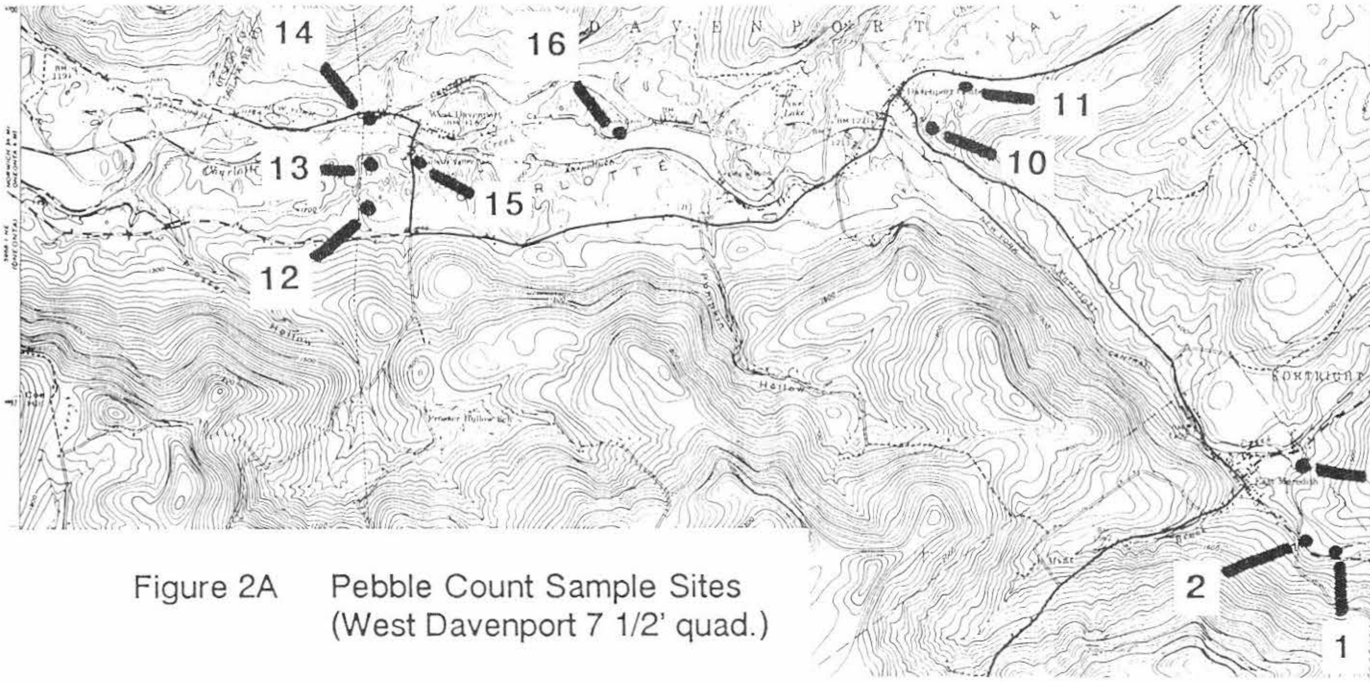

Figure 2A Pebble Count Sample Sites
(West Davenport 7 1/2' quad.)

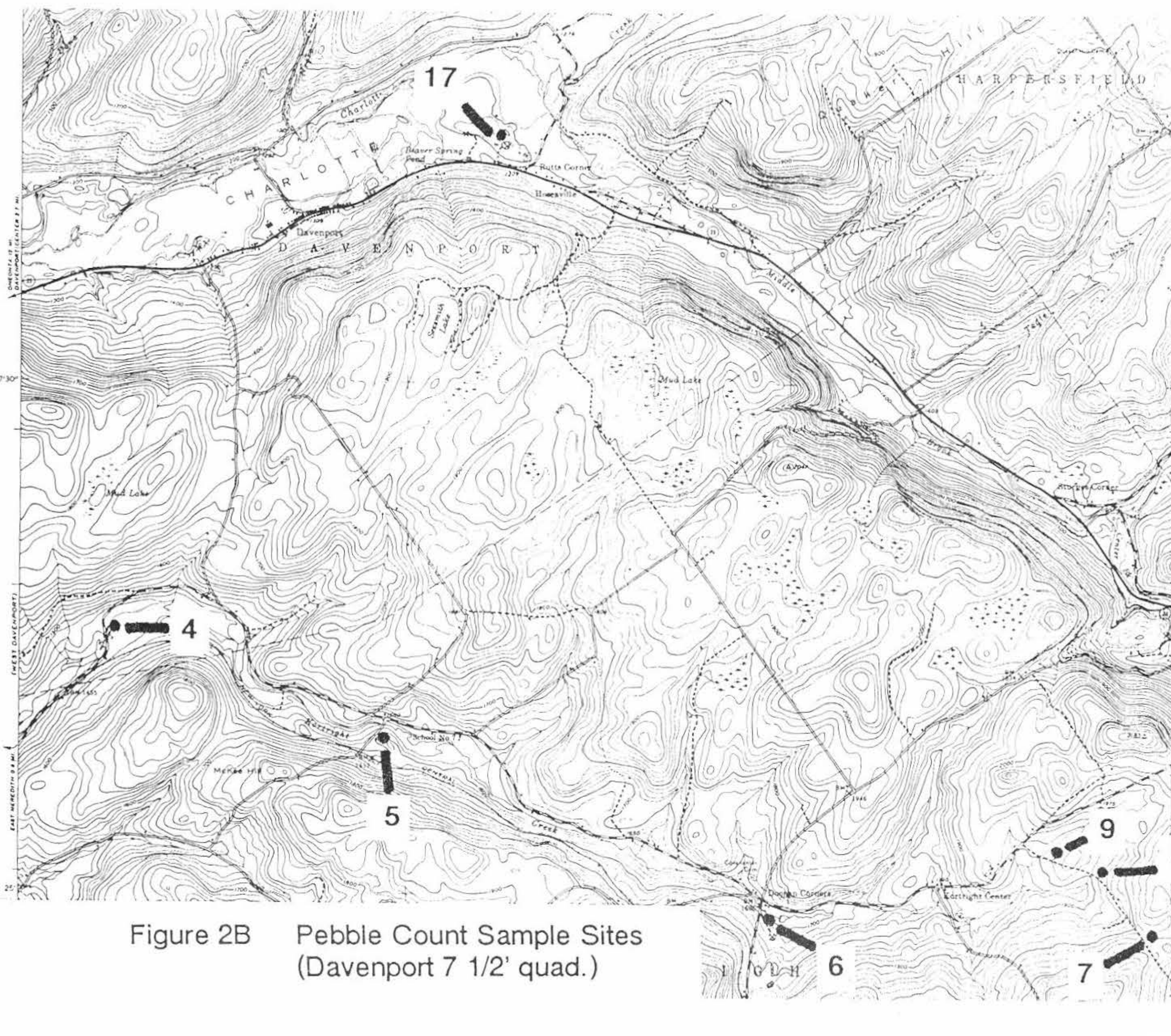

Figure 2B Pebble Count Sample Sites
(Davenport 7 1/2' quad.)

West Davenport Moraine

Samples 12, 13, and 14 were taken along a north-south traverse across the moraine at West Davenport and sample 15 from a road cut at the Charlotte Valley Cemetery in West Davenport.

Dead-ice sink complex

Sample 16 is the only source of data from the dead-ice sink complex. It was collected at McMinn Cemetery west of Pine Lake.

Butts Corner

A large delta/fan complex at the confluence of Middle Creek with Charlotte Valley (sample 17) was sampled 30-40 feet below the top of the landform in an active working face of the borrow pit.

SAMPLING AND ANALYSIS

Samples of approximately 100 pebbles were sought. Care was taken to avoid contamination from human activities. At borrow pit locations, samples were simply scooped from the exposed face into a bag. In all other locations, samples were obtained from holes dug in the surface. Samples were passed through a half inch sieve and washed. Pebbles were split to expose a fresh surface. The color and lithology of each pebble were noted under a binocular microscope. A summary of all data is shown in Tables 1 and 2.

DISCUSSION

Kortright Creek

Data for samples 1-9 (Fig. 4-10) show little change in color or lithology from the head of the valley to the bedrock knick point at East Meredith. No pebble-sized exotics were counted, yet exotic boulders were observed within the drainage basin.

Pitted hanging delta

Data for the delta seem to show two different suites of lithologies but in each a low percentage (2-3%) of exotic pebbles were counted.

Charlotte Valley

Data from two samples taken at Butts Corner represent a single lithologic suite, which differed from data obtained on the moraine and dead-ice sink complex.

Comparison of data

Lithologic suites from the head of Kortright Creek valley compare favorably with those found on the moraine, but differ from those at Butts Corner. Data on color characteristics vary from place to place and do not appear to show meaningful changes.

CONCLUSIONS

The data collected neither proves nor disproves the postulated inwash source for drift in the western part of the Charlotte Creek valley. The dominant lithologies in all samples are local. No specific lithology can be used to indicate provenance. However, similarity of lithologic suites suggest inwash from Kortright Creek provided material to the Dead-ice Sink Complex. A different suite at Butts Corner, may indicate still another source for upvalley deposits.

TABLE 1. PERCENTAGE OF PEBBLES BY COLOR

Color	Sample No.									
	1	2	3	4	5	6	7	8	9	Avg 1-9
Red	39	30	39	37	30	33	53	31	31	36
Gray	33	29	25	16	36	26	31	37	38	30
Dark Gray	1	7	5	5	2	2	1	4	7	4
Tan	24	34	25	41	23	36	17	26	23	44

Color	Sample No.									
	10	11	Avg. 10-1	12	13	14	15	Avg. 12-15	16	17
Red	25	10	18	16	16	23	19	19	29	8
Gray	45	29	37	31	36	22	20	27	31	1
Dark Gray	14	33	24	24	21	23	31	25	18	49
Tan	16	26	21	29	26	32	27	29	45	30
Exotic	2	2	2	0	0	0	2	1	5	2

Sample Locations: #1-9, Kortright Creek; #10-11, Hanging delta; #12-15, Moraine; #16, Dead-ice sink; #17, Butts Corner

TABLE 2

PERCENT OF LITHOLOGIES

Lithology	Sample No.									
	1	2	3	4	5	6	7	8	9	Avg. 1-9
Sandstone	33	37	28	32	34	31	41	23	32	32
Siltstone	45	48	55	47	40	51	45	55	50	48
Mudstone	22	15	17	21	26	18	14	21	18	19
Exotic	0	0	0	0	0	0	0	1	0	0

Lithology	Sample No.									
	10	11	Avg. 10-11	12	13	14	15	Avg. 12-15	16	17
Sandstone	21	27	24	25	9	26	29	22	22	8
Siltstone	44	52	48	43	47	53	25	42	42	28
Mudstone	34	19	27	31	44	21	44	35	34	62
Exotic	1	2	2	0	0	0	2	1	2	2

Sample Locations: #1-9, Kortright Creek; #10-11, Hanging delta; #12-15, Moraine; #16, Dead-ice sink; #17, Butts Corner

Case #4 Implications of pebble count data; confluence of Unadilla River and Tallette Creek, Columbus Quarters, New York

Jim Yuchniewicz and P. Jay Fleisher, Earth Sciences Department, SUNY-Oneonta

INTRODUCTION

Classical debates concerning contrasting lithologic suites within the Appalachian Plateau drift (MacClintock and Apfel, 1944; Meritt and Muller, 1959; and Moss and Ritter, 1962) indicate that upland drift has a lower percentage of erratic pebbles then through valley stratified drift. This led to consideration of different transport mechanisms for each type of drift. Evenson and Clinch (1987) emphasize the significance of glacio-fluvial transportation and the importance of inwash as a sediment source in the glacial environment.

Tallette Creek is an upland tributary to the Unadilla River approximately four miles north of New Berlin along Route 8. It drains an area of approximately ten square miles on the west side of the Unadilla Valley. As with all major through valleys, the Unadilla shows conspicuous valley asymmetry, with larger tributaries on more gentle east-facing slopes.

This study considers inwash as a possible sediment source for drift in the Unadilla Valley. Pebble counts of "bright" and "drab" drift along Tallette Creek Valley are compared to those taken from within a kame field at it's confluence with the Unadilla Valley. By definition, "bright" drift is composed of 30-50 percent erratic lithologies, whereas "drab" contains 0-15 percent (Randall, 1973).

SAMPLING PROCEDURE, PREPARATION, AND PEBBLE COUNTS

Random pebble samples were taken from various locations along Tallette Creek Valley and from within the kame field (Figure 1). Gravel pits offered access to some locations, but in many cases pebbles were obtained from dug holes. Some samples were taken from mounds next to woodchuck holes. Samples were taken approximately every half mile within the valley, alternating between the valley floor and upper slopes, as well as from convenient spots from within the kame field.

Twenty samples were returned to the laboratory, placed in a 12.5 mm (1/2 inch) sieving screen, washed, and individual pebbles broken for binocular microscope identification and determination of roundness.The average number of pebbles counted per site was 118 with a range between 57 and 222. The pebble grain size averaged between 15 and 25 mm, but did include a few cobble size clasts.

LITHOLOGIC ANALYSIS

As anticipated, local lithologies dominated all samples, with an average of 91 percent mudstones and siltstones from the Panther Mountain Formation. Bright

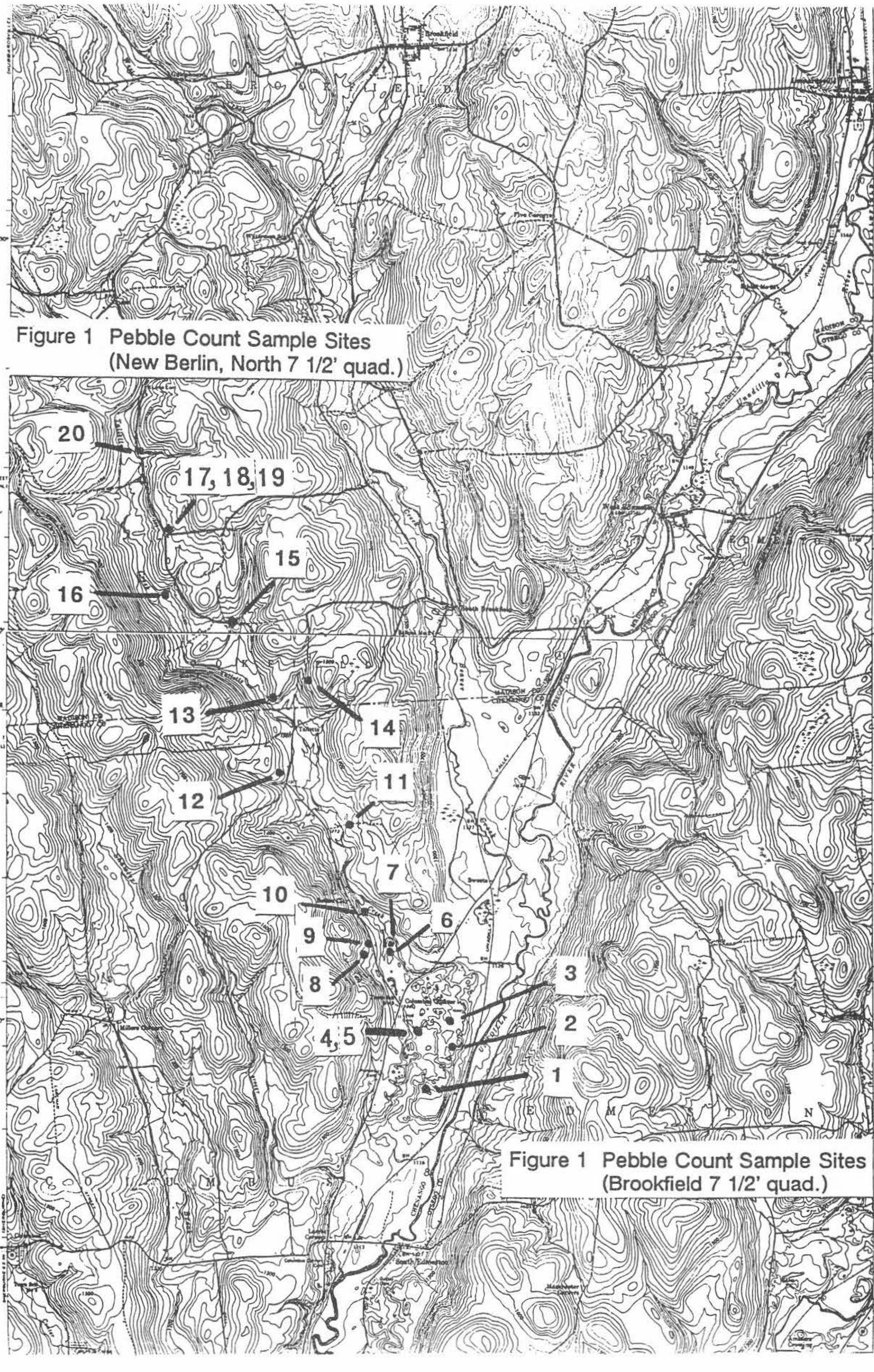
Figure 1 Pebble Count Sample Sites
(New Berlin, North 7 1/2' quad.)
20
17, 18, 19
15
16
13
14
12
11
10
7
6
9
8
3
2
4, 5
1
Figure 1 Pebble Count Sample Sites
(Brookfield 7 1/2' quad.)

lithologies varied between 0 and 27 percent. Kame field samples contained an average of 18 percent bright lithologies and ranged from 5 to 27 percent. Upland drift contains from 1 to 10 percent brights, with an average of 5 percent. The average along Tallette Creek Valley is 4 percent, with 3 percent along the stream channel.

Five types of bright lithologies were noted. Quartz sandstone and chert dominate, with decreasing amounts of limestone, quartzite and gneiss. Additional lithologies observed in the field, but not counted are anorthosite, megagabbro, amphibolite gneiss, and granitic gneiss.

Figure 2 summarizes all pebble count data. Samples from the upper slope at sites 12, 14, and 17-20 yielded relatively higher percentages of bright pebbles, which decreased gradually along the valley floor of Tallette Creek. At sites 6 and 7, an abrupt increase in the bright lithologies occurs. Samples taken from sites 4 and 5 contained even higher percentages bright lithologies, but the percentage decreased again along the eastern margin of the kame field (samples 1-3).

Figure 2. Summary of All Pebble Counts

SAMPLE NUMBER	LOCAL BEDROCK	QUARTZ SANDSTONE	CHERT	GNEISS	QUARTZITE	LIMESTONE
1	85.5	5.6	8.0	0.9	0.0	0.0
2	94.7	4.0	1.3	0.0	0.0	0.0
3	80.3	11.1	6.8	0.0	1.7	0.0
4	73.3	10.3	9.0	0.0	0.7	6.9
5	74.6	8.8	7.0	0.0	0.9	8.8
6	81.7	7.5	6.7	2.5	1.7	0.0
7	85.9	9.4	4.7	0.0	0.0	0.0
8	93.8	4.9	1.4	0.0	0.0	0.0
9	100.0	0.0	0.0	0.0	0.0	0.0
10	94.6	5.4	0.0	0.0	0.0	0.0
11	96.5	0.0	0.0	0.0	0.0	0.0
12	89.5	5.7	4.8	0.0	0.0	0.0
13	97.1	2.4	0.5	0.0	0.0	0.0
14	93.8	4.9	1.4	0.0	0.0	0.0
15	98.6	0.0	0.0	0.0	1.4	0.0
16	99.1	0.0	0.0	0.0	0.9	0.0
17	96.3	2.5	1.3	0.0	0.0	0.0
18	94.3	1.4	1.4	0.0	2.8	0.0
19	95.1	1.8	1.8	0.0	1.2	0.0
20	94.0	2.4	3.6	0.0	0.0	0.0

DEGREE OF ROUNDNESS

By definition, roundness refers to angularity of particle edges and corners (Powers, 1951). Most clasts are assumed to begin angular and become more rounded with increased distance of transport.

The roundness of pebbles from Tallette Creek Valley was compared with samples from the kame field. Along the upper slopes, pebbles were consistently sub-angular to angular, whereas along Tallette Creek's lower gradient and the kame field, they were predominantly sub-rounded to rounded. The change occurs at the location of sample 11, where Tallette Creek flows on a more gentler gradient and across a wider flood plain.

In addition, separate roundness studies were made for each prominent bright lithology (limestone, quartz sandstone, chert). As might be expected, limestone pebbles were sub-rounded to well-rounded (possibly by solution). Quartz sandstone pebbles seemed to become slightly more rounded, while chert remained angular.

CONCLUSIONS

1. The highest concentrations (14 to 27 percent) of bright pebbles are found within the kame field, whereas values along Tallette Creek Valley are significantly lower (1 to 10 percent) (Figure 3). Therefore, drift within the kame field is "bright" relative to the "drab" drift of Tallette Creek Valley.

2. Pebbles transported down Tallette Creek show some degree of increased roundness.

3. The drift within the kame field is from two sources: (1) reworked upland drift transported as alluvium down Tallette Creek after the uplands became ice free (deposited as inwash on the Unadilla Valley ice-tongue), and (2) supraglacial meltwater transportation along the western edge of the valley ice-tongue led to deposition at the mouth of Tallette Creek Valley. Therefore, the kame field contains pitted outwash that has been supplemented by inwash from Tallette Creek

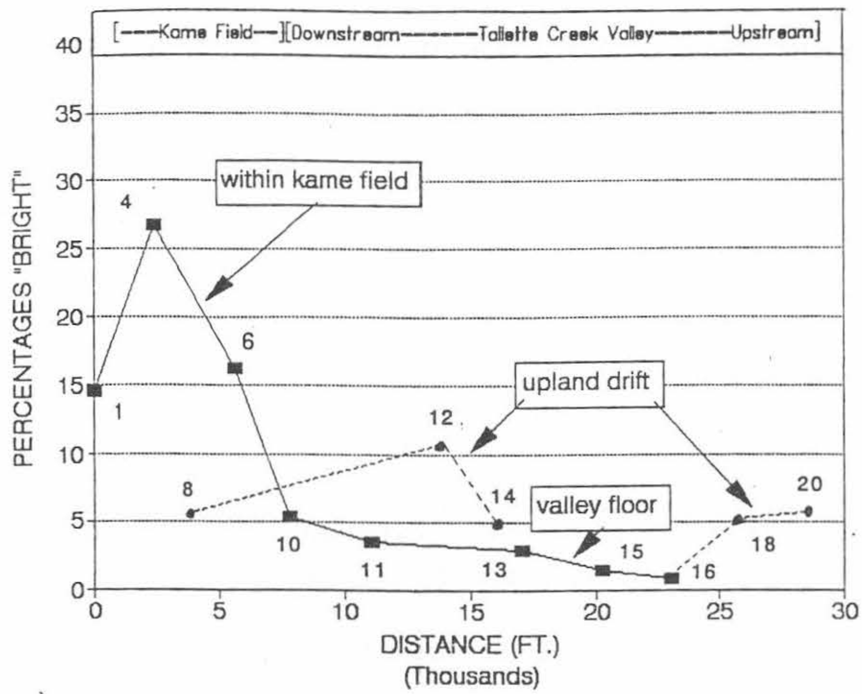

Figure 3. Changes in bright lithologies* along Tallette Creek. Plot of data (Figure 2) indicates the percentage of exotic lithologies is greatest within the kame field (14.6 - 26.7) and least along the valley of Tallette Creek (0.9-5.4). Values from upland drift vary between 4.3 and 10.5%. *All non-local rock type.

Road Log begins at intersecton of State Routes 7 and 205, West End, Oneonta.

Miles from last point	Cumulative Miles	
0	0	Intersection of Rt. 7 and 205, proceed north on Rt. 205.
1.1	1.1	Rt. 23 enters from right; continue north on Rt. 205 and Rt. 23.
.7	1.8	Bear left on Rt. 23.
9.2	12.9	Intersection of Rt. 23 and 51 in Village of Morris; continue straight through intersection on County Rt. 13 to New Berlin
7.2	20.1	Dead-ice sink on right
.8	20.9	Unadilla River
.2	21.1	Intersection of County Rt. 13 and Rt. 8; turn left (south) on Rt. 8
.4	21.5	Pull off to the right in gravel excavation and walk up dirt road (Angell Hill Road) to upper level of quarry

STOP # 1 - New Berlin: Through valley landforms - deltaic valley train and dead-ice sink. Large-scale deltaic foreset beds exposed here and within a currently inactive quarry in the center of the valley contain clast supported, very coarse gravels and pebbly, coarse sands; leached and reprecipitated carbonate (limestone) form partial cement in some units. This valley train is breached by the Unadilla River that flows from a dead-ice sink upvalley to a lacustrine plain downvalley. Landforms indicate detachment of large ice mass during active ice retreat.

		Turn around; proceed north on Rt. 8 through Village of New Berlin
.7	22.2	Intersection of Rt. 8 and 80 (traffic light), proceed north on Rt. 8 and 80.
1.3	23.5	Continue north on Rt. 8
.5	24.0	Road rises on gravel terrace above lacustrine plain
1.4	25.4	High-level terrace (1280') on west side of valley above road (continues for 1.5 miles) is of unknown origin.

2.2	27.6	Road rises on pitted planar gravel and kame field
.9	28.5	Columbus Quarters at intersection of Rt. 8 and Chenango County Rt. 41.

STOP # 2 - Columbus Quarters: kame field and pitted plain in through valley. These landforms, breached by the Unadilla River at its confluence with Tellette Creek, may have served to temporarily dam a local lake upvalley. Pebble counts from upland drab drift and kame field bright drift suggest tributary inwash from the Tellette Creek drainage basin was a partial source of sediment (see Case # 4 for discussion of pebble count data).

		Turn around, proceed south on Rt. 8.
1.9	30.4	At Lambs Corners turn left (east) onto Chenango County Rt. 25 (which changes to Otsego County Rt. 20 at Unadilla River), proceed east across lacustrine plain
4.2	34.6	Road descends onto the pitted and discontinuous valley train of Butternut Creek.
.6	35.2	Intersection of County Rt. 20 and Rt. 80, proceed east on Rt. 80.
3.0	38.2	Rt. 51 enters from the right, continue east on Rt. 80.
.4	39.3	Rt. 51 turns left, continue east on Rt. 80.
6.7	46.0	Rt. 205 enters from the right, continue east on Rt. 80 and 205.
1.7	47.7	Road descends onto the kame and kettle topography of the moraine at Oaksville.
.2	47.9	Intersection with Rt. 28, turn right, proceed south on Rt. 80 and 28.
.5	48.4	Road crosses the crest of the moraine and continues on and off the moraine for next 1.5 miles through the villages of Oaksville and Fly Creek.
1.7	50.1	Turn right (south) in Fly Creek on County Rt. 26, then bear left at fork.

1.2	51.3	Turn left onto gravel road, park and walk 1/4 mile into quarry.

STOP # 3 - Fly Creek: Valley train and dead-ice sinks in through valley. Sand and gravel valley train with classic assemblage of glaciofluvial sedimentary structures. What criteria may be used to distinguish outwash from inwash?

		Return to County Rt. 26, continue south.
2.3	53.6	Pull off on right shoulder, park and walk to crest of moraine.

STOP # 4 - Cassville-Cooperstown moraine at Index: This classic through valley landform (moraine and associated valley train) represents active ice deposition. It consists of stratified sand and gravel that grade upvalley into silt 120+ feet thick (see Case #1, Figure 2). Is it an end moraine, as proposed by Krall, 1977, or a recessional moraine? Although high enough, it did not dam Glacial Lake Cooperstown. Landforms and stratigraphy between here and Otsego Lake suggest detachment of a marginal-ice cleat and development of a dead-ice sink (see Case #1, Figure 4).

.2	53.8	Proceed to the intersection with Rt. 28, turn left (north).
.2	54.0	Dead-ice sink behind Cassville-Cooperstown moraine occupies the valley for the next mile.
1.7	55.7	Junction Rt. 80 and 28, proceed east on Rt. 80.
.3	56.0	Traffic light intersection with Main St., Cooperstown; continue straight through intersection east on Rt. 80.
.1	56.1	Stop sign, junction Lake St. with Rt. 80, turn right onto Lake St.
.3	56.4	Park at Intersection of Lake and River St.; stairway to Council Rock to the left.

STOP # 5 - Council Rock Park on Doubleday Ice Margin: Well data indicate 180 feet of bouldery silt beneath a moraine that plugs the valley and dams Otsego Lake. Subtle hummocky terrain in the village of Cooperstown can be traced westward to the golf course. While ice-cored, this landform dammed Glacial Lake Cooperstown at an elevation of 1250'. The view northward reveals conspicuous cross-sectional valley asymmetry, which is an important aspect of sediment source for Glacial Lake Cooperstown (see Case #1, Figures 5-9).

		Turn around and return to Rt. 80.

.3	56.7	Turn left onto Chestnut St. (Rt. 28 and Rt. 80 West).
.4	57.1	Junction Rt. 80 west, proceed south on Rt. 28.
2.1	59.2	Highway traverses Cassville-Cooperstown moraine.
1.0	60.2	Hamlet of Hyde Park; from here south for several miles, planar gravels appear related to upland tributary inwash sources.
5.3	65.5	Junction Rt. 166 at village of Milford (traffic light), continue south on Rt. 28. Valley floor consists of lacustrine clays from here south to a moraine at Portlandville at 68.9.
.9	71.6	Turn onto entrance road to new development in Milford Center
.2	71.8	Turn around, park.

STOP # 6 - Goodyear Lake overview at Milford Center: Pitted valley train and dead-ice sink. These landforms indicate local stagnation of the valley ice-tongue during active retreat. However, the valley fill consists of 60+ ft. of ice-contact stratified drift over 300+ ft. of silt interpreted to be of lacustrine origin. Does the stratigraphy indicate two stades (readvance) or is there a single-stade environment that accounts for both stratigraphic units? Aggradation over ground ice islands is suggested.

.3	72.1	Return to Rt. 28, turn south (right) onto Rt. 28.
1.8	73.9	Junction with Rt. 7, continue south on Rt. 28 to I-88 interchange.
.8	74.7	Rt. 7 overpass, dead-ice sink right and left.
.5	75.2	Turn right onto I-88 west.
2.5	77.7	Take Exit 16 (Emmons/Davenport Center)
.3	78.0	Stop sign, turn left toward Davenport Center on County Rt. 47.
.3	78.3	Cross Susquehanna River.
I.6	79.9	Delaware County line (Otsego County Rt. 47 becomes Delaware County Rt. 11).

l.0	80.9	Turn left in Davenport Center and continue past the Post Office on left. Road parallels Charlotte Creek through the moraine on the right.
1.9	82.8	Turn right into parking lot for Hartwick College's Pine Lake Camp. Walk downhill to pavilion.

STOP # 7 - Davenport Center: Pine Lake Dead-ice Sink Complex. These kames and sinks formed in an ice-cored terrain at the mouth of Charlotte Creek where the valley was clogged by remnant ice. Case # 3 considers the origin of landforms here. Also see Case # 3, Appendix A for the results of a pebble count study that tested an inwash origin for these materials.

		Return to road, turn right.
.4	83.2	Right hand fork goes downhill and across the lacustrine plain of Glacial Lake Davenport.
.4	83.6	Intersection with Rt. 23, proceed straight on Delaware County Rt. 10.
.2	83.8	Turn left into Town of Davenport Transfer Station, county gravel excavation

STOP # 8 - Davenport Center: pitted hanging delta. Current and past exposures contained well-developed, large-scale deltaic foreset beds in a hanging delta marked by kettles 40+ ft. deep. This indicates progradation onto and across grounded ice in Glacial Lake Davenport. See Case # 3 for discussion.

		Backtrack to Rt. 23.
.2	84.0	Turn left (west) onto Rt. 23.
1.9	85.9	Highway traverses moraine at West Davenport.
2.3	88.2	Terrain to the left is what may be the only lateral moraine in this area.
.7	88.9	Road descends to Susquehanna valley lacustrine plain.
1.1	90.0	Highway enters the breach of the Oneonta moraine (significantly altered by urban development).
1.3	91.3	Traffic light intersection I-88, 28 N/S, 23W; continue straight on 28 south.

.6	91.9	Turn right toward I-88 west.
.3	92.2	Turn left onto I-88 west.
2.0	94.2	Take Exit 13 (Morris and Route 205).
.3	94.5	Turn right (north) onto Route 205.
.2	94.7	This brings you back to the start of this log at Junction of Rts. 7 and 205.

OPTIONAL - continuation up non-through valley of Otego Creek.

0	0	Start at fork of Rt. 23 west (to West Oneonta and Morris) and Rt. 205 (Laurens) north, proceed north on Rt. 205.
4.2	4.2	Winnie Hill Road joins Rt. 205 from the right. Pull over and walk up Winnie Hill Road 0.1 mile for downvalley view south across local kame field.
.7	4.9	Turn left on County Rt. 11A (unmarked) toward Laurens.
.3	5.2	Turn left on County Rt. 11 (Maple Street) through village of Laurens.
.4	5.6	Bear left at fork (sign to Oneonta).
.8	6.0	Turn left into landfill entrance at cemetery, park.

OPTIONAL STOP # 9 - Village of Laurens landfill: ice-contact stratified drift. At various times, excavation has exposed tilted and truncated sequences of interstratified gravel and pebbly sand with fluvial sedimentary structures and local silt layers of lacustrine origin. This is interpreted to indicate semi-continuous aggradation of ice-cored terrain. Tributary inwash from the west is a possible sediment source.

		Return to road, turn right (north), backtrack through village of Laurens.
.8	6.8	Junction 11A and 11 (Maple Street), continue north on Maple Street.
2.1	8.9	Planar gravel, valley train remnant for next half mile.

.4	9.3	Turn right on unnamed road; sign indicates bridge closed.
.5	9.8	Turn left (.1 mile short of bridge) on unpaved gravel entrance road.
.3	10.1	Park and walk to various exposures.

STOP # 10 - Otsego County sand and gravel quarry: planar gravel remnant. This landform, typical of many others within the Otego Creek valley, is a discontinuous, planar gravel deposit interpreted as a remnant of a surface of aggradation that included semi-continuous ice masses as vestiges of a collapsed ice tongue. Ice-contact fluvial and lacustrine facies are common.

.2	10.3	Return to paved road (County Rt. 11).
.4	10.7	Turn right onto County Rt. 11, proceed north.
.9	11.6	Junction Rts. 11 and 11B, continue straight on 11. Road rises onto pitted planar gravel remnant.
1.0	12.6	Junction Rts. 11 and 15; bear right continuing on Rt. 11, road traverses terrain of ice-cored origin including a small esker out of sight to the right.
.3	12.9	Turn right onto Angel Road.
.6	13.5	Junction Angel Road and Rt. 205; turn left (north) onto Rt. 205 toward Hartwick.
.6	14.1	Pull off on left shoulder.

STOP # 11 - Overview of meltwater channel through discontinuous planar gravel remnant and terrain of ice-cored origin.

.5	14.6	Road descends to a local lacustrine plain (dead-ice sink?)
.7	15.3	Gravel excavation to the left removed a landform thought to be a proto-esker.
2.5	17.8	County Rt. 45 enters from the right, pull off on right shoulder.

STOP # 12 - Overview of discontinuous, planar gravel surface and terrain of ice-cored origin bordering a local, upvalley lacustrine plain.

		Proceed north on Rt. 205.
1.2	19.0	Village of Hartwick, Junction Rt. 205 and County Rt. 11, proceed north on Rt. 205.
1.3	20.3	Pull off on right shoulder.

STOP # 13 - Overview of headward area in non-through valley. Valley floor contains a few isolated kames. Unlike through valleys, the valley floor gradually rises to upland slopes across which ice flow diminished during glacial retreat resulting in downvalley ice-tongue starvation and collapse.

END OF ROAD LOG

Continue on Rt. 205 (north) 4 miles to intersection with Rt. 80 or return on Rt. 205 (south) 16.5 miles to the Junction of Rts. 23 and 205 where this optional trip began.

UNDERSTANDING THE EAST CENTRAL ONONDAGA FORMATION (MIDDLE DEVONIAN) - AN EXAMINATION OF THE FACIES AND BRACHIOPOD COMMUNITIES OF THE CHERRY VALLEY SECTION, AND MT. TOM, A SMALL PINNACLE REEF

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INTRODUCTION

The Onondaga Formation of New York and Ontario, Canada has been extensively studied (see Oliver, 1976 for references), and yet is still poorly understood. It is a unit which is, in the western part of New York and in Ontario, clearly transgressive, and yet it lacks any of the classic peritidal facies associated with shallow water carbonates. It is a "reefy" unit, but the major reef building paleocommunity of the Middle Devonian (stromatoporoids and algae) are either extremely rare or absent. Finally, it has been described as an example of carbonate deposition along a gently subsiding ramp, which would seem to imply symmetry on either side of the basinal axis, and yet the pinnacle reefs - which are so highly sought after by explorationists - have been found on the western side of the basinal axis, but to date not on the eastern side. Kissling (1987) suggested that these unusual characteristics were due to deposition in deep water, possibly below the photic zone. As an alternative, Wolosz (1990, 1991) has argued that the Onondaga represents an example of a Devonian temperate water limestone.

On this trip we will attempt to come to our own conclusions by examining a nearly complete section of the Onondaga (Cherry Valley) and a small pinnacle reef (Mt. Tom). In the following text, Lindemann analyzes the stratigraphy and depositional environments, Feldman the significance of the brachiopod communities, and Wolosz and Paquette the depositional history of the Mt. Tom reef.

STRATIGRAPHY AND DEPOSITIONAL HISTORY

The Onondaga Limestone is a 21-50+ meter thick unit of lower Middle Devonian marine limestones deposited during the final major phase of carbonate production prior to the influx of siliciclastic sediments shed from the Acadian mountain buildup. This component of the field trip is intended to provide an overview of Onondaga stratigraphy and depositional environments in Otsego County, New York. It centers on a nearly complete composite section of the formation exposed in road cuts on U.S.Route 20 at Cherry Valley (Sprout Brook, New York 7.5'quadrangle). These exposures mark the easternmost extent of the "typical" Onondaga as defined in central New York by Oliver (1954).

The concept of what is now the Onondaga Formation began to be developed prior to the First Geological Survey of New York (see Eaton, 1832). During the survey Vanuxem (1842) recognized four "formations," the uppermost of which he named the "Seneca limestone," a label which persists today. Hall (1843) recognized a three-fold division and applied the term "Onondaga limestone" to those strata now known as the Edgecliff Member. Through time the "Onondaga Limestone" came to include the entire interval of limestone strata which overlies formations of the Lower Devonian and is itself overlain by the black shales of the Marcellus Formation. Oliver (1954, 1956a) formally subdivided the Onondaga into four members. In the best tradition of the recently canonized Nicolas Steno, the members arranged from oldest to youngest are the Edgecliff, Nedrow, Moorehouse, and Seneca limestones. Descriptions of the members provided herein are specific to the exposures at Cherry Valley, NY. Lithologic terminology corresponds to that of Lindholm (1964).

Biotratigraphy and Correlation

The biostratigraphic basis for correlation of the Onondaga Formation, particularly the Edgecliff Member, to the standard biozones and stages of Europe reads like a "who-done-it?" with the last few pages missing. Brachiopod and coral faunas have long served to place the formation at the base of the Middle

Devonian Series (Rickard, 1975). Dutro (1981) reports that the Edgecliff Member coincides with the base of the *Frimbrispirifer divaricatus* Subzone of the *Amphigenia* Assemblage Zone which marks the base of the Southwoodian (=Upper Onesquethawan) Stage. Similarly, Oliver and Sorauf (1981) state that the Edgecliff base coincides with the base of the *Acinophyllum segreatum* Assemblage Zone and the bottom of the Southwoodian Stage. High in the formation, the Seneca Member is within the *Paraspirifer acuminatus* Assemblage Zone as well as an unnamed coral assemblage zone (8 of Oliver and Sorauf, 1981), both of which place the Seneca in the Cazenovian Stage. These fossils firmly establish the Onondaga Edgecliff Member as the lowermost Middle Devonian unit in New York State relative to the North American stages. However, Onondaga corals and brachiopods are geographically restricted to North America, a condition which precludes direct correlation with the Eifelian Stage of Europe. Cephalopods do little to facilitate this correlation. *Foordites* cf. *buttsi* from the Nedrow Member (Oliver, 1956b; House, 1962) suggests an Eifelian age. However, House (1981) notes that *Foordites* is a long-ranging genus, and that European zonal taxa are not known from the Onondaga. Furthermore, *Foordites* has not been reported from the Edgecliff and cannot lend its support to interpretation of an Eifelian age for the lowermost Onondaga member.

The International Union of Geological Sciences recently ratified the decision of the Subcommission on Devonian Stratigraphy to drive the golden spike marking the base of the Middle Devonian Series and the Eifelian Stage at the first occurrence of the conodont Polygnathus costatus partitus (Ziegler and Klapper, 1985). The subspecies partitus is the second in a lineage of three. The bottom of the P. c. patulus Zone is high in the Emsian Stage and the bottom of the P. c. costatus Zone is well within the Eifelian. Klapper (1981) reports that the upper Nedrow beds at Cherry Valley yield both P. c. costatus and P. c. patulus, placing the member's top well within the partitus Zone. Noting this along with the fact the P. c. partitus is unknown from the Onondaga, Ziegler and Klapper (1985) suggest, with question marks, that the Edgecliff Member is within the patulus Zone and correlative to the Emsian Stage of the Lower Devonian Series. At the very least, this obfuscates the Lower-Middle Devonian boundary in New York State and speaks for a "handle with care" approach in transatlantic correlation of the North American Stages which abut that boundary.

Recent studies of the Onondaga's styliolinid and tentaculitid faunas have done little to improve upon this situation. A previously unknown nowakid fauna has been discovered in the Nedrow and Moorehouse Members at Cherry Valley, but the taxonomic status of the species is currently undetermined. Lindemann and Yochelson (1984) reported that the first occurrence of Styliolina fissurella (Hall) in the Devonian of New York is coincident with the base of the Edgecliff Member. Indeed, at Cherry Valley this enigmatic microfossil is present in the lowermost bed of the Edgecliff and absent from the subjacent Carlisle Center. S. fissurella (Hall) was a zooplankter reputed to have had a nearly worldwide distribution (Boucek, 1964). The potential for correlation is obvious. However, Lindemann and Yochelson (in press) have found that many, possibly all, reports of the species from the Devonian of Europe are incorrect. Thus, without specific confirmation reports of S. fissurella (Hall) from the Lower Devonian must be regarded with suspicion. To date the chronostratigraphic placement of the lowermost Onondaga member remains uncertain.

Descriptions of the Members

Edgecliff Member - The Edgecliff is seven meters thick and is divisible into two components which correspond to the C1 and C2 zones of Oliver (1956a). The lowermost beds contain quartz sand and silt, glauconite sand, and phosphatic nodules. The limestones associated with these particles and which overlie the beds containing them are thin to medium bedded, dark gray, argillaceous, packed biocalcisiltite. The middle and upper Edgecliff consists of thickly to very thickly bedded, medium gray, poorly washed to unsorted biosparites. While corals dominate the macrofauna, pelmatozoan ossicles and fenestrate bryozoans volumetrically dominate the sediment. The uppermost Edgecliff bed is a poorly washed biosparite which contains an abundance of pyrite. This bed is abruptly overlain by the basal Nedrow.

Nedrow Member - The Nedrow is a four meter thick package of what might be described as coarsening upward cycles. More accurately, they are cycles of progressive carbonate enrichment without pronounced textural cyclicity. A cycle begins abruptly with a thickly laminated, argillaceous and pyritic, fossiliferous biocalcisiltite and grades vertically into medium bedded, dark gray, sparse biocalcisiltite. The sediments are extensively bioturbated. While pelmatozoans and trilobites are the most abundant biogenic particles, ramose bryozoans and styliolines reach maximal abundances. Crushed styliolines in the more argillaceous beds indicate an overall thickness loss of approximately 75% due to soft sediment compaction. Thus, the original Nedrow sediment may have been 15-20 meters thick. Laminae within the Nedrow beds are the result of compaction of the sediment. They are not primary sedimentary structures.

Moorehouse Member - The Nedrow/Moorehouse contact coincides with the first occurrence of black chert (Oliver, 1956a). Nodules, anastomosing masses, and thin beds of dark gray to black chert are characteristic of the lower and middle Moorehouse. Limestones associated with the chert are a sequence of medium bedded, dark gray, fossiliferous to sparse biocalcisiltites. Terrigenous mud occurs as thin laminae and pyrite is rare. Bioturbation is abundant to pervasive, though individual burrows are indistinctly defined. Pelmatozoans, trilobites, and brachiopods variously dominate the sediment. Fenestrate bryozoans increase in abundance to become a major component high in the middle Moorehouse. Corals such as Aulopora and Thamnopora also increase in abundance, as do goniatite cephalopods.

The uppermost Moorehouse is distinct from the lower and middle sections. Chert is rare to absent. Terrigenous mud is minimal. The limestone itself consists of thickly bedded, medium gray, packed biocalcisiltites and poorly washed biosparites. Cross stratification is present though not common. While the macrofauna is dominated by the encrusting cyclostome Fistulipora and other bryozoans of ramose form, the sediment matrix is dominated by pelmatozoans and fenestrate bryozoans.

Seneca Member - The lowermost bed of the Seneca Member is the Tioga Bentonite (Oliver, 1954). At Cherry Valley the Tioga is ten centimeters thick. It is extremely weathered, producing a deep reentrant between the more resistant Moorehouse and Seneca limestones. The Seneca proper consists of approximately two meters of thickly bedded, medium gray, packed biocalcisiltites and poorly washed biosparites. High angle cross laminae are present and the majority of disarticulated brachiopod valves are in a convex up orientation. Pyrite is virtually absent and terrigenous mud attains a formational minimum for this locality. While the macrofauna is dominated by atrypid brachiopods the sediment matrix is predominantly pelmatozoan debris and fenestrate bryozoans. The top of the Seneca is approximated though not attained.

Depositional History

The Onondaga has long been interpreted as a sequence of limestones deposited in progressively deepening waters and terminated by the progradation of the Marcellus black shales derived from the rising Acadian orogen. Within this model the Nedrow Member represents an influx of terrigenous mud; a hint of greater things to come. The Tioga Bentonite is a single event horizon. The Moorehouse and Seneca Members become increasingly argillaceous as the sea gradually deepened and the Marcellus muds slowly advanced from east to west across the state. Sir Charles Lyell would have found comfort in this model. It's ponderous unfolding would have appealed to his aesthetic tastes. However, to badly paraphrase Mark Twain - Recent study has cast much darkness upon the subject.

To begin with, the fidelity of the Tioga Bentonite has been called into question. For some years now it has been known that there are three separate bentonites high in the Onondaga of western New York. It has been supposed that they converge to one in the vicinity of Syracuse due to a relatively low rate of sedimentation in that area. However, on an NYSGA field trip in 1986 a second bentonite was discovered at Jamesville, NY. How many more are there? The recent report of multiple bentonites in the Lower Devonian Kalkberg Limestone of eastern New York (Shaw, et.al., 1991) suggests that there may be several.

During the above mentioned field trip (Feldman and Lindemann, 1986) a classic Devonian bone bed was found high in the Seneca Limestone. It was also observed that there is no lithologic gradation between the Seneca and the Marcellus and that the contact between the two units is an erosional truncation surface involving up to three beds of the uppermost Seneca. Lindemann and Feldman (1987) described a comparable disconformity at the top of the Onondaga in the central Hudson Valley of eastern New York. At Cherry Valley the precise top of the Seneca is not exposed, but the beds which can be seen give no indication of gradually giving way to shale. As is the case in the Hudson Valley, a relatively brief time of rapid crustal subsidence and an interruption of sedimentation would seem to be indicated.

Abrupt fluctuations in water depth are also indicated at the lower end of the column. Glauconite sand and phosphatic gravel in the lowermost beds of the Edgecliff Member at Cherry Valley indicate an interruption in sedimentation. It was during this unrecorded interval that deposition of the Carlisle Center ended as the depositional environment shifted to one favoring carbonate production. Since there is no definitive interpretation for the depositional history of the Carlisle Center it is difficult to ascertain what might have transpired during the unrecorded interval. Quartz silt and well rounded grains of quartz sand at the base of the Edgecliff could suggest relatively high levels of water energy, but the abundances of terrigenous mud and calcisilt with which they occur suggest otherwise. Furthermore, the phosphatic gravels at the base of the Edgecliff appear to have been involved in multiple generations of exhumation and reburial. There would seem to be more involved here than was previously supposed. Considering the absence of a biostratigraphic basis for the correlation of the Edgecliff to either the upper Emsian or the lower Eifelian, this phosphatic diastem is intriguing. Hopefully an ongoing study of this interval will soon yield results.

The remainder of the Edgecliff at Cherry Valley is equally intriguing though less cryptic from a paleoenvironmental point of view. The Edgecliff consists of dark gray packed biocalcisiltites (= C1 zone of Oliver, 1956a) overlain by medium gray, coraliferous, biosparites (= C2 zone of Oliver, 1956a). Obviously deposition did not begin in a high energy environment. Wolosz (1985) reported that Edgecliff reefs of the Hudson Valley exhibited evidence of a brief lowering of relative sea level. This was followed by a sea level rise and a resumption of reef growth. Wolosz and Lindemann (1986) correlated the shallowing event to the abrupt onset of biosparite deposition in the Edgecliff throughout eastern New York. This interpretation remains appropriate for the Edgecliff at Cherry Valley.

The top of the Edgecliff is anomalously pyritic. It is immediately overlain by the argillaceous biocalcisiltites of the Nedrow Member. The contact between the two is interpreted to be a diastem resulting from a pulse of crustal subsidence. Pyrite in the Nedrow and the Lower Moorehouse indicate relatively low concentrations of oxygen, possibly due to stratification of the water column. The Nedrow sediments do not suggest an influx of terrigenous mud but rather a shift to an offshore position coupled with a drastic reduction in carbonate production. The sediments' fine grained nature suggests a flocculent or soupy sediment-water interface, a condition not particularly conducive to colonization by the larvae of sessile organisms. Thus, the Edgecliff reefs were drowned in deep water rather than suffocated in mud.

Moorehouse deposition marks a return to enhanced carbonate production by benthic organisms living at depths well in excess of wave base. This is quite different from the top of the Moorehouse where a carbonate bank environment near wave base is indicated. Unfortunately the Moorehouse is not fully exposed and the transitional beds are not available for study. However, detailed study of polished slabs and thin sections through the lower and middle Moorehouse reveals a symmetry in the sequence of lithologic changes which centers around beds about ten meters from the base of the member. The beds below indicate a progressive increase in water depth and soupiness of the substrate. The beds above show the exact opposite trend. Unlike the remainder of the Onondaga at Cherry Valley, it appears likely that the shallowing upward trend was gradual and not a punctuational event. Lyell would have preferred it this way.

BRACHIOPOD COMMUNITIES

What Is A Community?

Communities are often defined as recurrent associations of taxa which were presumably controlled by a set of environmental factors such as: substrate, salinity, temperature, pressure, current action, wave action, light penetration, nutrients, dissolved oxygen, and water chemistry. Ecologists are not necessarily in agreement as to what the definition of a community is, nor how to recognize one. Boucot (1981) notes major subdivisions of current conceptualizations of community definition, including those who define "community" as a superorganism that has a virtual life of its own - a living and breathing community. At the other extreme are those who hold that communities are no more than chance aggregations of organisms conducting their affairs quite independently of one another - ships that pass in the night; apartment dwellers who have never been introduced to their neighbors.

Paleoecologists are at a distinct disadvantage in attempting to reconstruct ancient communities, since it is extremely difficult to determine the various relationships of taxa in terms of parasitism, commensalism, mutualism, and other dependent and interdependent variables that are not readily apparent in the fossil record. As Boucot (1981) notes, the paleoecologist is reduced to examining statistical data on relative abundance and presence or absence of taxa in an attempt to infer ecological interaction. There is much biological information important in community reconstruction which cannot be retrieved from the rock record, and this must be kept in mind when coming to conclusions about community make-up. Ecologists who study Recent communities have a distinct advantage in this regard over paleoecologists, and are able avoid dependence solely on hard part data.

Brachiopod Communities of the Onondaga Limestone

When studying the brachiopod communities of the Onondaga Limestone in New York State, other faunal constituents and their fragments, such as trilobites, corals, and gastropods were tabulated (Feldman, 1980; Feldman and Lindemann, 1986; Lindemann and Feldman, 1987). Numbers of brachiopods were determined by counting the most abundant valve. Relative abundance was variable, depending on geographic area and member sampled. For example, collecting in the shaly Nedrow Member in central New York was much more productive than in the dense Moorehouse Member. However, in eastern New York, the silicified Moorehouse yielded many more well-preserved taxa than did the nonsilicified Nedrow. Therefore, relative abundance seems to be a function of: (1) lithology, (2) rate of weathering, and (3) silicification. The Onondaga Limestone is most productive, in terms of brachiopods, when well silicified. Unfortunately, this occurs rarely, notable localities being in the mid-Hudson and Genesee valleys. In the mid-Hudson Valley heavy jointing is associated with silicification. There are many outcrops which do show evidence of weak silicification and collecting from these areas can range from excellent to poor, depending on the degree of silicification (whether surficial or deep). Beekite rings on shells observed in outcrop are usually indicative of weak silicification. The Onondaga brachiopod communities recognized in New York State are briefly described below.

Atrypa-_Coelospira_-_Nucleospira_ Community. The ACN (= _Atrypa_-_Coelospira_-_Nucleospira_) Community ranges from Leeds to just south of Kingston, New York and occurs predominantly in the Moorehouse Member. Diversity here is great (29 brachiopod genera), but only 13 genera comprise the bulk of the community (Feldman, 1980). Of those, three genera (_Atrypa_, _Coelospira_, _Nucleospira_) represent a trophic nucleus of low-level epifaunal suspension feeders. A similar fauna is found in Lenz's (1976) Lower Lochkovian _Howellella_-_Protathyris_ Community in an offshore position. Taxa in common include: _Atrypa_, _Schizophoria_, _Ambocoelia_, _Coelospira_, _Nucleospira_ and "_Schuchertella_." Lenz's fauna is characterized by similar morphotypes (Table 1).

Atrypa-_Megakozlowskiella_ Community. The AM (=_Atrypa_-_Megakozlowskiella_ Community recognized from Clarksville to Cherry Valley, New York, is lower in diversity than the ACN Community (22 compared to 29 genera). This may be indicative of a position closer to shore and consequently nearer to wave base. A major faunal element that appears here is the robust spiriferid _Megakozlowskiella raricosta_, which had a large, triangular delthyrium in the ephebic stage, with lateral bordering ridges indicative of a deltidial plate. If the pedicle had no way of protruding the brachiopod would therefore have lived free on the sea floor. The pedicle valve had deeper ribs and was more convex than the flatter brachial valve, which would have provided a more hydrodynamically stable position for the animal if the pedicle

Table 1

A comparison of Lenz's (1976) Howellella-Protathyris Community with the ACN Community of the Onondaga Limestone.

Morphotype	Howellella-Protathyris Community	ACN Community
Broad, flat	"Schuchertella"	Schuchertella
Relatively smooth	Protathyris, Cryptatrypa	Nucleospira, Athyris
Broad, unequally bi-convex	Schizophoria	Schizophoria
Frilly	Atrypa	Atrypa

valve was in an "up" position. Some gerontic shells had secondary shell material deposited in the umbonal region as a counterweight, serving to keep the anterior commissure above the sediment-water interface.

Atrypa Community. The Atrypa Community occurs from the mid-Hudson Valley to Cherry Valley and is dominated by Hudson Valley "reticularis" (52.4%), with a relatively high diversity of 18 brachiopod genera (compare with the diversity of the AM Community of 22 genera). This community is very similar to Copper's (1966) European Eifel magnafacies which is composed of calcareous shales, muddy limestones and rare dolomites. Although there are no dolomites within the Onondaga, the Nedrow and Moorehouse members certainly contain a fair amount of mud. Other similarities, in addition to lithology, include the presence of varied brachiopod genera in both environments (such as spiriferids, rhynchonellids, athyrids, meristellids and gypidulids) and the occurrence of rugose and tabulate corals, stromatoporoids and crinoids.

Leptaena-Megakozlowskiella Community. The AM (= Leptaena-Megakozlowskiella) Community is recognized in the Syracuse area of central New York and is dominated by Megakozlowskiella raricosta and the ubiquitous Leptaena "rhomboidalis." Within the Onondaga Leptaena occurs more frequently in "muddier" limestone units and is relatively rare in the Edgecliff Member. A distinct association between the two genera is very evident on bedding plane surfaces in the shaly Nedrow Member, where they comprise a trophic nucleus of low-level suspension feeders. The diversity is fairly high, with 17 brachiopod genera represented. Crinoidal fragments and Platyceratid gastropods which are common in the ACN Community are absent here; in their place are other gastropod genera such as Straparollus, Liospira and Ecculiomphalus.

"Pacificocoelia" Community. This community has been found at only one outcrop in the Nedrow Member near Syracuse, New York, and is similar to the LM Community in two respects: [1] There is a close association between Leptaena (9.7%) and Megakozlowskiella (8.6%), and [2] Both communities are typically found in the shaly rather than the "cleaner" lime units. They differ in that in the "Pacificocoelia" Community the brachiopod diversity is low (10 genera) and no corals were recovered, whereas in the LM Community 17 brachiopod genera were found as well as 3 rugose and 3 tabulate coral genera.

Levenea Community I. This community occurs in the Edgecliff Member from Cherry Valley southeast to Kingston, New York and is dominated by Levenea sp. A (67.4%), with minor occurrences of Atrypa, Levenea sp. B, Leptaena, Pentamerella and Elytha. In general, the brachiopods are poorly represented in the Edgecliff Member. This may somehow be related to the large amount of chert present in the east, which seems to correlate with a reduced coral fauna. In central New York there is a large coral fauna, relatively little chert and more brachiopods.

Levenea Community II. Found only in the Moorehouse Member of southeastern New York, at an abandoned quarry in Wawarsing, the Levenea Community II consists exclusively of Levenea sp. A (100%). It differs from Levenea Community I in two respects: [1] the diversity is extremely low, and [2] the lithology is very different, consisting of "muddy" rocks interpreted to represent deposition in a more offshore position. This is consistent with the interpretation of a deepening structural basin in Onondaga times southwest towards Port Jervis (Lindemann and Feldman, 1987).

Amphigenia? Community. The Amphigenia? Community is found in the basal Edgecliff near Syracuse, New York and is based on the recovery of only 12 specimens. There is a possibility that these fragmental shells were reworked and transported, since the occurrence of Amphigenia in the sandy facies of the Edgecliff is not compatible with Boucot's (1975) placement of the genus in a Benthic Assemblage 3-5 position.

Hallinetes Community. The Hallinetes Community (Racheboeuf and Feldman, 1990), formerly recognized as a Chonetes Community (Feldman, 1980), occurs only in the Seneca Member of the Onondaga Limestone. Three taxa comprise the chonetacean brachiopods in this community: Hallinetes lineatus (92%), Longispina mucronata (5.4%) and "Eodevonaria" hemispherica (2%). Other brachiopod taxa are present but represent minor faunal constituents (see

Feldman, 1980, p. 40). Based on new observations, it is apparent that the Hallinetes Community is most accurately represented by the shells within the dark mudstone matrix rather than by those distributed on bedding plane surfaces. The community is a low-diversity, "highly dominated" (although not monospecific) community within a quiet water environment (Racheboeuf and Feldman, 1990).

Communities of Western New York

Based on preliminary analysis of material collected from the Moorehouse Member of the Onondaga Limestone in the Genesee Valley of western New York, an almost identical ACN Community to the one found in the mid-Hudson Valley is recognized. Similarities include dominance by the low-level epifaunal suspension feeders Atrypa, Coelospira and Nucleospira as well as very high diversity (39 brachiopod genera, including two new athyrids). The two communities differ in that the ACN Community of the Genesee Valley has a significantly larger proportion of strophomenids, including some genera absent in southeastern New York: "Brachiprion" aff. mirabilis, Protoleptostrophia perplana, Plicostropheodonta? sp. and Costistrophonella ampla. Also, there are other taxa in the west not recovered from southeastern New York: Camarospira? sp., Alatiformia? sp., Mediospirifer sp.A and B, Paraspirifer sp., Cranaena sp. and Cryptonella sp.

Community Paleogeography

Work is currently in progress which will clarify the relationships of these various communities to one another across New York State. However, a general pattern can be observed (Table 2). Most data have been collected from the Nedrow and Moorehouse members, therefore, by omitting those communities found only in the Edgecliff and Seneca members (Amphigenia, Levenea Community I, Hallinetes), it appears that during Nedrow-Moorehouse time [1] there was a trend towards increasing diversity away from the basinal axis and, [2] diversity decreased towards a subsiding structural basin.

Lindemann and Feldman (1987) note that in central New York, a transgression submerged the region initiating Edgecliff deposition in a shallow shelf environment. Soon thereafter subsidence in central New York, resulting from a northward extension of the Appalachian Basin, brought deeper water and an offshore environment to the area. The initial pulses of subsidence are recorded in the Nedrow Member, while continued subsidence is evidenced in the Moorehouse and Seneca members of the central region. The eastern and western areas, that is, those areas away from the basinal axis, remained in shallow shelf conditions resulting in a symmetric shelf-basin-shelf pattern as seen in east-west outcrop.

A subsiding structural basin in the Tristates area, (not directly related to the topographic basin of central New York), which had existed since the Middle Silurian was noted by

Table 2

Brachiopod communities in the Onondaga Limestone of New York.

Brachiopod Community	Number of Genera*	Member	Geographic Location
ACN	39	Moorehouse	Genesee Valley
ACN	29	Moorehouse	Hudson Valley
AM	22	Moorehouse	Cherry Valley
Atrypa	18	Nedrow-Moorehouse	Hudson Valley, Cherry Valley
LM	17	Nedrow-Moorehouse	Syracuse
Hallinetes	10	Seneca	Syracuse
"*Pacificocoelia*"	10	Nedrow	Syracuse
Levenea Community I	6	Edgecliff	Hudson Valley
Levenea Community II	1	Moorehouse	Wawarsing
Amphigenia	1	Edgecliff	Syracuse

* Refers to brachiopod genera; ACN = *Atrypa*-*Coelospira*-*Nucleospira* Community; AM = *Atrypa*-*Megakozlowskiella* Community; LM = *Leptaena*-*Megakozlowskiella* Community.

Lindemann and Feldman (1987). This basin greatly influenced Onondaga deposition in southeastern New York by creating a carbonate slope, or ramp, dipping into the Port Jervis area. It is from this ramp that the Levenea Community II was recovered, indicating a trend toward lower diversity in the direction of the deep waters of the structural basin. Further collecting and analysis of brachiopod communities along the ramp will help support or reject this proposed paleogeographic distribution and correlation with basin depth.

MT. TOM - A SMALL EDGECLIFF PINNACLE REEF

Oliver (1956c) described the location and size of Mount Tom, labeled it Mt. Tom #1, and included it among the seven reef exposures comprising the Mt. Tom Reef Group which are scattered over an approximately 9 square mile area at the boundary of the East Springfield, Richfield Springs, Jordanville, and Van Hornesville 7.5 minute quadrangles. Mt. Tom is the largest reef exposure in the group, forming a prominent hill in the northwest corner of the East Sprinfield 7.5 minute Quadrangle (it is, in fact, the thickest known surface exposure of an Edgecliff reef (Oliver, 1956c, p.21)).

While Oliver considered all seven Mt. Tom Group exposures to represent separate reefs, Paquette and Wolosz (1987) noted that the two exposures closest to Mt. Tom #1 (see Fig. 1) - Mt. Tom #2 reef (approximately due west of Mt. Tom) and Mt. Tom #6

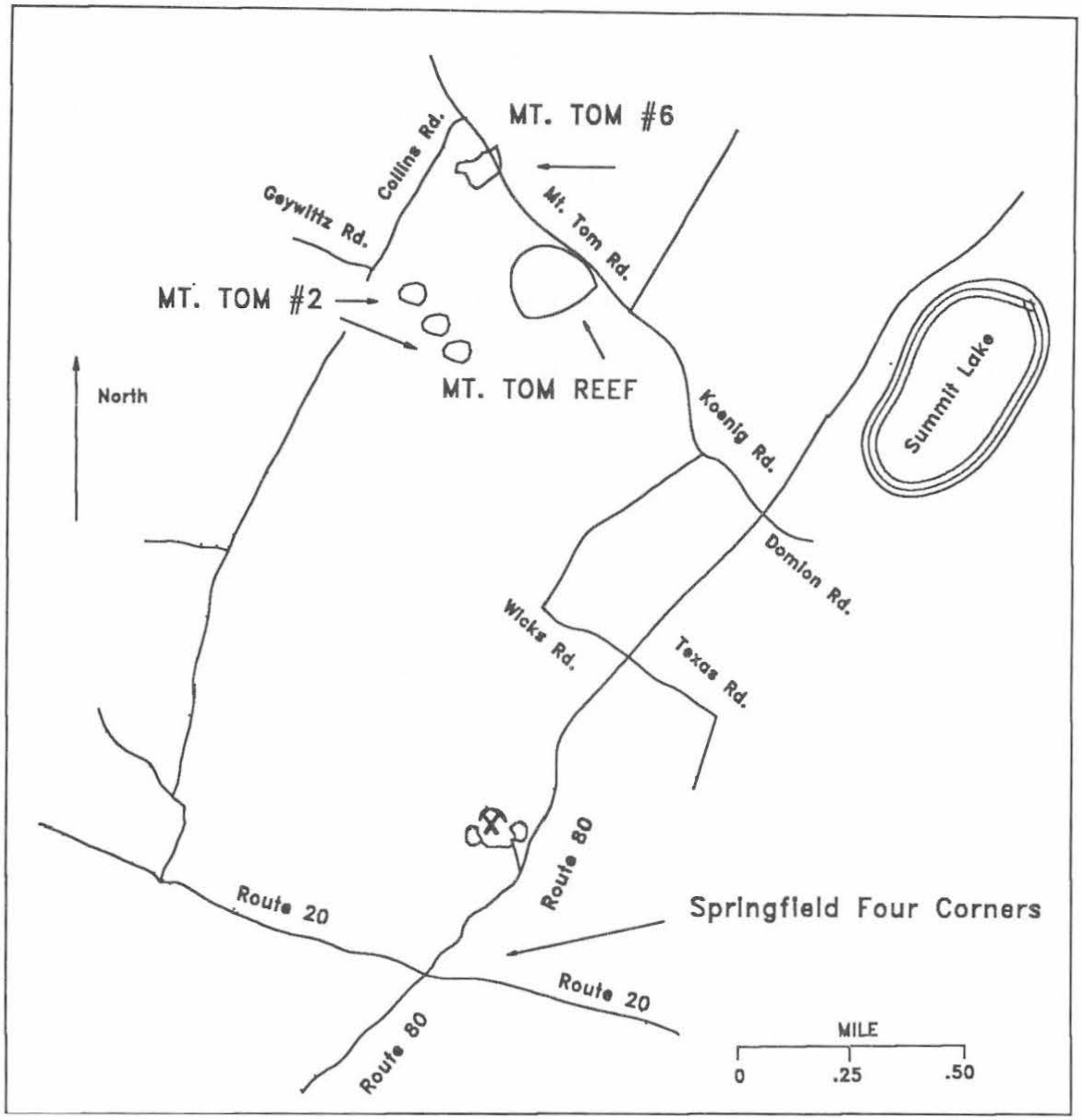

FIGURE 1. Location map of Mount Tom reef #'1, 2 and 6. Note relative positions of reefs.

(northwest of Mt. Tom) - were comprised mainly of crinoidal grainstone/packstone which dipped away from Mt. Tom #1. They argued that these three exposures represent the erosionally dissected remains of a small pinnacle reef, approximated at about 150 acres.

Reef Communities

Typical of Edgecliff reefs, Mt. Tom is made up of two distinct paleocommunities - the phaceloid colonial rugosan paleocommunity and the favositid/crinoidal sand paleocommunity.

The phaceloid colonial rugosan paleocommunity consists almost exclusively of colonial rugosans. Common genera include *Acinophyllum*, *Cylindrophyllum*, and *Cyathocylindrium*; with *Eridophyllum*, *Synaptophyllum*, and rare phaceloid colonies of *Heliophyllum* as accessories. The dense growth of these rugosan colonies appears to have restricted most other organisms to only minor roles, with favositids (both domal and branching) being small and rare, brachiopods uncommon, and bryozoans mainly fragmentary encrusters.

The favositid/crinoidal sand paleocommunity displays a much higher diversity than the rugosan paleocommunity. This paleocommunity is more biostromal than biohermal. Large sheet-like to domal favositids are abundant, but never form a constructional mass. Solitary rugose corals are also extremely abundant as are fenestrate bryozoan colonies. Single colonies of the mound building phaceloid rugosans are occasionally found. Brachiopods and other reef dwellers are also common although never extremely abundant. Stromatoporoids and massive colonial rugosans, while extremely rare in the Edgecliff reefs, when found are part of this paleocommunity. The crinoids were the greatest contributor to this paleocommunity - ossicles making up the bulk of the rock and indicating abundant growth of these organisms - but complete calyces are never found.

Mt. Tom Reefs #1, 2 and 6

Wolosz (1990a, in press) presented a classification of Edgecliff reef types based on the relative importance of the two reef paleocommunities to the development of the reef structure. Mt. Tom reef is an example of a Mound/Bank Composite Structure. Mounds are distinct high relief buildups of the phaceloid colonial rugosan paleocommunity which occur as either small (generally not more than 1 - 3 meters thick) monogeneric to mixed faunal buildups; or as Successional Mounds up to roughly 15 meters thick which display an internal succession of mound building colonial rugosan genera. The term "bank" follows the definition of Nelson, et al. (1962, p.242): "a skeletal limestone deposit formed by organisms which do not have the ecologic potential to erect a rigid, wave resistant structure." Hence, Mound/Bank Reefs are large structures resulting from the repetitive intergrowth of rugosan mounds and the favositid/crinoidal sand facies. Pinnacle reefs found in the subsurface in New York and Pennsylvania also represent this type of structure and reach thicknesses of up to 60 meters.

The mound\bank nature of Mt. Tom #1 is displayed in the cliff face along the southeast side of the hill (Figure 2). The reef is underlain by the basal Edgecliff calcisiltite (C1 unit of Oliver, 1956a), with the base of the reef marked by thickets of

Acinophyllum. Small phaceloid colonial rugosan mounds (again, mainly Acinophyllum) can be observed along the cliff near the base of the reef. These small mounds and thickets coalesced to begin the formation of the larger structure. Dominance of the inital large mound shifted between Acinophyllum and Cylindrophyllum prior to onlapping by the crinoidal sands of the favositid/crinoidal sand paleocommunity. A second mound stage made up of Cylindrophyllum thickets overlies these grainstones and packstones. In turn, the second mound stage is itself onlapped and eventually swamped by the favositid/crinoidal sand paleocommunity (exposed further back on the top of the hill, not shown in Figure 2). Overall, Mt. Tom #1 is roughly 18m thick as preserved.

Wolosz and Paquette (1988) have interpreted this mound\bank\mound\bank pattern as catch-up\fall back cycles controlled by fluctuations in water depth above the top of the reef. It is important to note that the second mound building stage at Mt. Tom #1 (Figure 2) does not drape the entire pre-existing structure, but is instead restricted to the top of that structure. In effect, during bank stage, the reef was a high relief platform on the sea-floor with its top within the ecologic mound building zone of the colonial rugosans. Upward growth of the reef is mainly due to the repetitive establishment of new mounds on the top of the platform. As sea-level was approached, the mound building colonial rugosans were overwhelmed by increased turbulence conditions and the mounds onlapped by encroaching crinoidal sands producing a bank stage; but with sea-level rise the mounds became re-established. This shifting

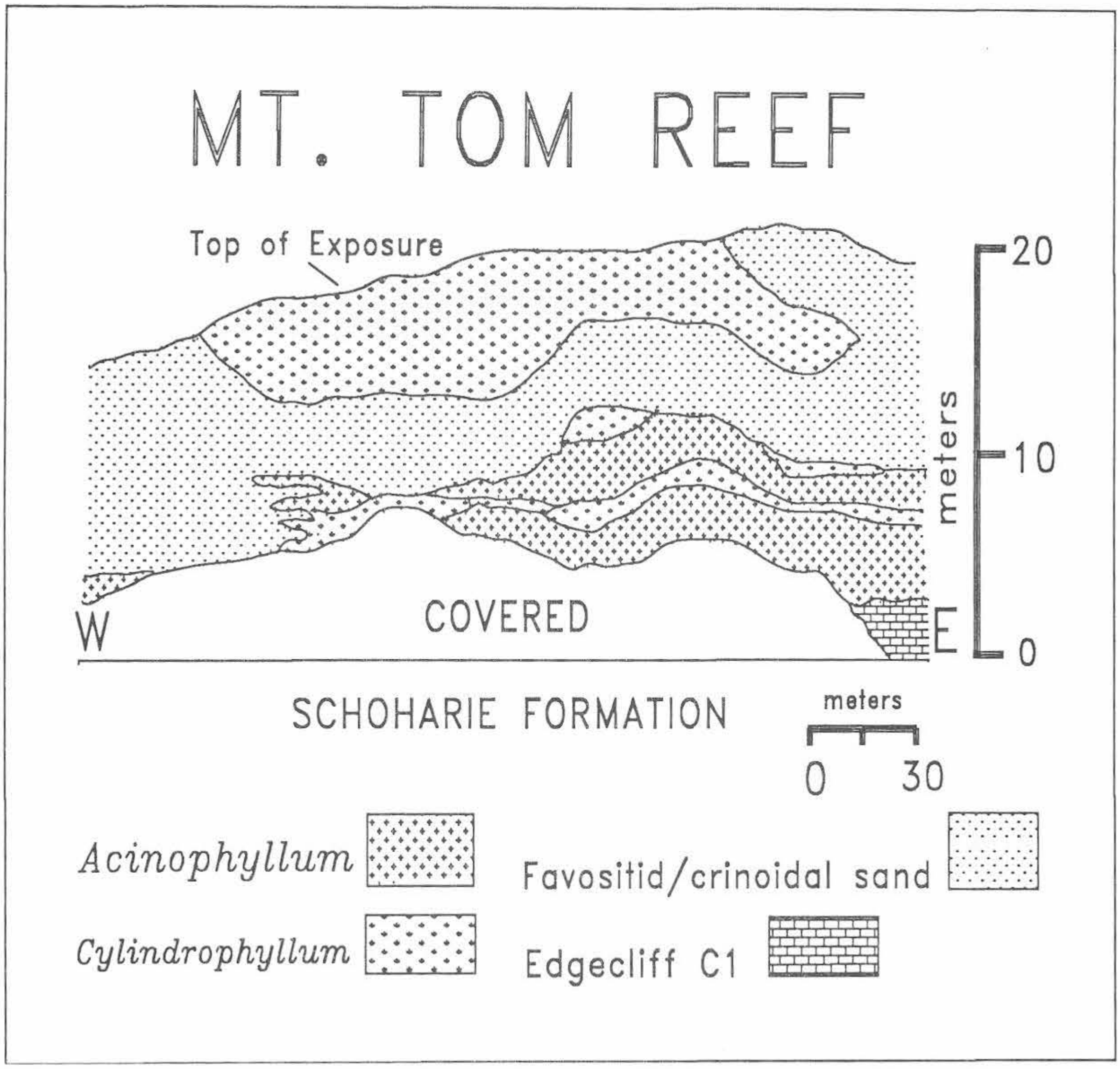

FIGURE 2. Cross-section of cliff face at Mt. Tom #1 reef illustrating mound/bank structure. Two rugosan mound stages are separated by favositid/crinoidal sand facies (bank stage). Note that second rugosan mound stage does not drape entire reef structure (after Wolosz, in press a).

between rugosan mound/thicket construction and the favositid/crinoidal sand paleocommunity has been attributed to a water turbulence controlled community succession (Wolosz, 1989a, 1989b, in press b).

Following the initial mound building stage, lateral growth of Mt. Tom appears to have been due mainly to deposition of crinoidal debris flanks with occasional small mound structures (satellite mounds) growing in those flanks (see discussion of Mt. Tom #2). A similar, but less well developed mound\bank\mound sequence has been described at Roberts Hill Reef south of Albany (Wolosz, 1985).

To the northwest, Mt. Tom #6 is a small ridge which consists mainly of crinoidal grainstone/packstone but with more abundant fossils. Small overturned favositids are common as are both solitary and phaceloid rugosans, but no evidence of mound formation is present. However, when one observes Mt. Tom #6 from Collins Road (see map, Figure 1), the questa-like nature of this small ridge is evident, with the dip slope pointing to the north-northwest, directly away from the main mass of Mt. Tom.

Topographically, Mt. Tom #6 is at the same elevation as the present top of Mt. Tom. Since the regional southwest dip of about 18 meters/kilometer (Rickard and Zenger, 1964, p.5) would not greatly alter this topographic relationship, the elevations of the Mt. Tom #6 exposure and the top of Mt. Tom were probably also equivalent at the time of deposition. Paquette and Wolosz (1987) cited this as evidence that the two exposures are parts of one reef, with Mt. Tom #6 consisting of distal flank beds. Mt.

Tom reef would then be at least 0.8km. long on an northwest axis from Mt. Tom #1 to Mt. Tom #6.

In contrast, Mt. Tom #2 lies to the west of Mt. Tom #1 and is topographically roughly 18 meters below #6. Stratigraphically older beds can be examined here, with the Edgecliff/Carlisle Center contact marked by the appearance of a spring just east of the intersection of Collins and Geywittz Roads. A small quarry visible from the road exposes bedded Edgecliff with overturned colonial coral. To the southeast of this quarry is an exposure of a small colonial rugosan mound roughly 17 meters across and of indeterminate thickness. East from the quarry, along the south side of the creek, there are numerous outcrops of bedded crinoidal grainstone/packstone with abundant favositids. Small patches or lenses of colonial rugosans within the bedded packestones are common, and represent small satellite thickets or mounds which appear to range stratigraphically from near the C1/C2 contact (roughly the point at which growth of Mt. Tom #1 began), upwards to about 6 meters above that contact. The packstones surrounding these upper mounds dip away from Mt. Tom #1 at roughly 15 degrees.

Tying The Exposures Together - Development of the Mt. Tom Pinnacle Reef

Figure 3 illustrates an interpreted developmental history for the Mt. Tom (small) pinnacle reef. As sea-level dropped from possible deep water conditions of Carlisle Center deposition through the early Edgecliff (C1), abundant small rugosan thickets and mounds began to form in the late C1 calcisilts. By the beginning of C2 deposition these thickets and small mounds had begun to coalesce to form the initial large mound at Mt. Tom #1 (Mound Stage I), while an abundance of other small mounds dotted the crinoidal sand sea-floor as satellites to the growing reef. Crinoidal debris of the favositid/crinoidal sand paleocommunity lapped up onto the large mound, eventually forming flank beds which spread outward from the main mass of the reef. Small satellite mounds continued to develop along distal flank beds (Mt. Tom #2), contributing to the overall volume of the reef structure, but never coalescing into a large central structure similar to Mt. Tom #1. Continued sea-level drop resulted in the cessation of rugosan mound growth and the eventual swamping of the mound by the crinoidal sand beds, resulting in Bank Stage I. A second cycle of sea-level rise resulted in the establishment of new rugosan thickets and mounds on the top of the bank (Mound Stage II), but later shallowing over the crest of the reef again caused the demise of the colonial rugosans and the re-establishment of the favositid/crinoidal sand paleocommunity in Bank Stage II.

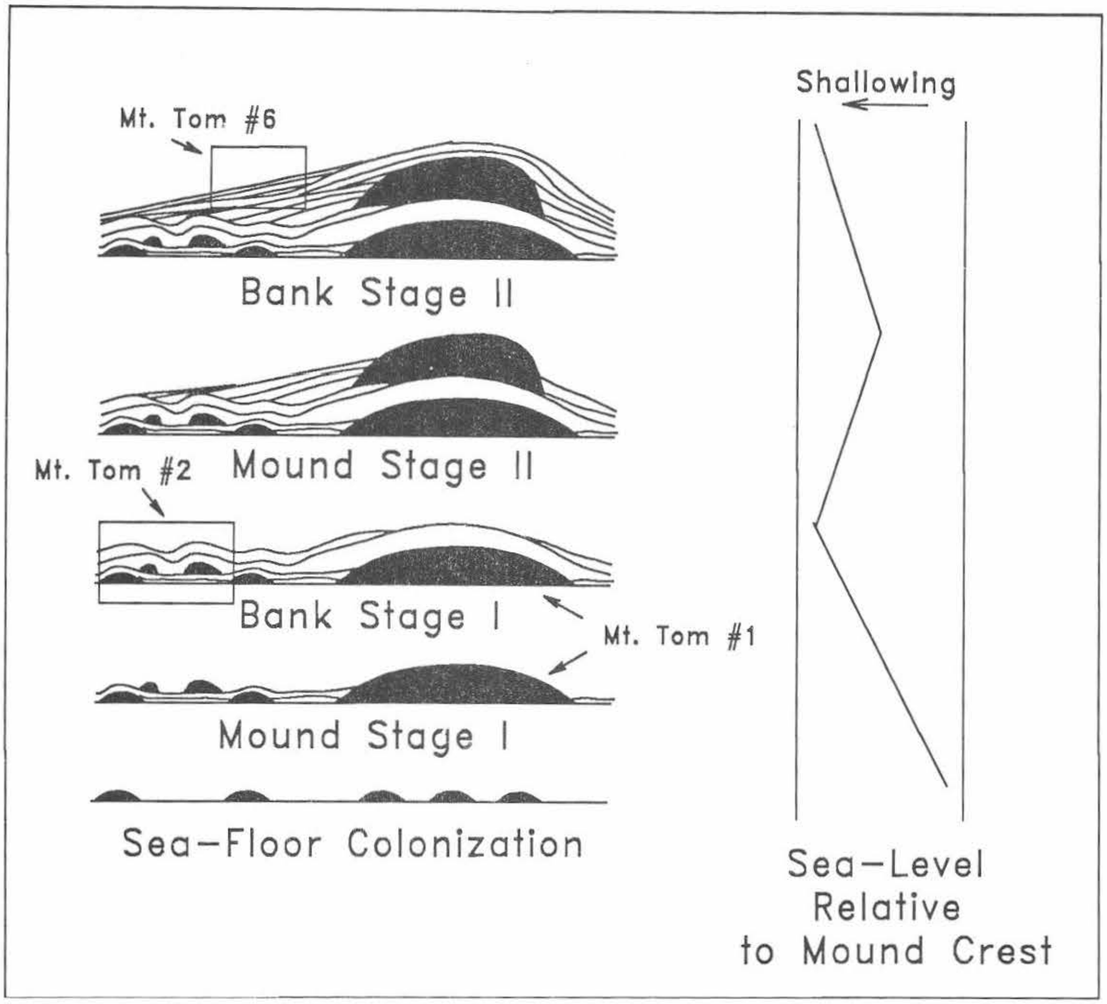

FIGURE 3. Sequential model for growth of the Mt. Tom pinnacle reef. Interpreted sea-level changes shown at right. Boxes indicate interpreted position of Mt. Tom #'s 2 and 6 exposures. Main mound is Mt. Tom #1 exposure. See text for further details.

In Figure 3, Mt. Tom #6 is illustrated as distal flank beds. While this is correct, the illustration is not to scale and somewhat misleading. When the distance from #6 to Mt. Tom #1 is considered (roughly 0.8km.), along with the 15° to 25° dip of the beds at #6, and the already noted paleo-topographic equivalence of the present top of #1 and #6; the conclusion that, if totally preserved, Mt. Tom reef would be much thicker than the present erosional remnant is easily arrived at. Unfortunately, there appears to be no way to achieve a valid estimate of that thickness.

Any attempt to directly correlate the reef growth cycles preserved at Mt. Tom with the non-reefal Onondaga (for instance at Cherry Valley) would require a detailed micro-stratigraphy which is, unfortunately, not available. However, the following statements can serve as a basis for discussion and further research.

The first mound/bank cycle at Mt. Tom follows the shallowing trend from C1 to C2 deposition in the Edgecliff. The initial pattern here is similar to that described at Roberts Hill (Wolosz, 1985; Wolosz and Lindemann, 1986). However, as sea-level begins to rise, leading to the second mound/bank cycle, the environment at the crest of the reef (or the top of the bank) becomes decoupled from that of the surrounding sea-floor. In order to produce the large pinnacle structure, the top of the bank must be maintained within a fairly narrow environmental range suitable for the two reef building paleocommunities (see Wolosz, in press b, for discussion). If we assume that by the second bank stage (current top of Mt. Tom #1) the reef was roughly 18m thick with bank top at least 10m above the surrounding ocean floor (given that the C2 at East Springfield is roughly 4m thick (Oliver, 1956a) and allowing for a 50%

compaction of the carbonate sediments); and also assume that the Edgecliff/Nedrow contact marks a starvation boundary (see discussion of stratigraphy), then at this point much lateral growth of the bank would occur since large amounts of crinoidal debris from the favositid/crinoidal sand community would be washed off the bank onto the flanks while upward growth would be limited by sea level. Such a senario would leave a well developed bank with the potential for continued upward growth once renewed subsidence led to the onset of Nedrow deposition. In effect, environmental conditions characteristic of the Edgecliff would continue on the bank top while Nedrow sediments were being deposited on the surrounding sea-floor.

The Edgecliff Reefs - Cool Water Structures?

As mentioned in the introduction, Kissling and his students have pointed to the lack of stromatoporoids and calcareous algae, in conjunction with the absence of clear peritidal deposits to suggest that the Edgecliff reefs may have been deposited in deep water. An alternative hypothesis to the deep water model is for the Edgecliff to have been deposited under cool water conditions. Wolosz and Paquette (1988) suggested a cool water environment for the Edgecliff, as have Koch and Boucot (1982) based on the Edgecliff brachiopod fauna; Blodgett, et al., (1988) based on gastropod faunas; and Wolosz (1990b, 1991) based on stromatoporoid abundance trends.

The cool water model for Edgecliff deposition supplies answers to many of the questions listed in the Introduction. The C2 facies is a shallow water facies, but one more akin to modern FORAMOL deposition (Lees, 1975) then to tropical carbonate deposition. The reefs are then analogous to modern ahermatypic coral banks, built by relatively slow-growing colonial rugosans poorly adapted to high energy conditions - hence their replacement by the favositid/crinoidal sand community under high energy conditions. The cool waters would also explain the rarity of stromatoporoids and the absence of algae - both groups being restricted to warm waters.

In conclusion, the paleo-biological evidence appears to support a model of the Edgecliff as a temperate water carbonate.

ACKNOWLEDGEMENTS

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FIELD TRIP STOPS

CUMULATIVE MILEAGE	MILES FROM LAST POINT	ROUTE DESCRIPTION
0.0	0.0	Intersection of Routes 20 and 10, Sharon Springs. Proceed west along Route 20.
7.0	7.0	STOP 1. Cherry Valley Section. Park along road at top of west end of road cut. An almost complete section of the Onondaga Limestone is exposed along this cut. (See discussion of

		Stratigraphy and Brachiopod Communities). Return to cars, proceed west along Route 20.
13.2	6.2	Right turn on Route 80.(see Figure 1)
14.7	1.5	Left turn onto Koenig Road.
15.3	0.6	Bare left onto Mt. Tom Road
15.4	0.1	STOP 2. Mt. Tom Reef. makes up the large hill to the south of the road (See discussion of Mt. Tom). Return to cars and continue northwest on Mt. Tom Road.
15.9	0.5	STOP 3. Mt. Tom #6 forms the low, wooded ridge to the southwest of the road (See discussion in test). Return to cars and continue northwest on Mt. Tom Road.
16.05	0.15	Left turn onto Collins Road.
16.55	0.5	STOP 4. Intersection of Collins and Geywittz Roads. Leave cars and proceed east from the intersection. Mt. Tom #2 forms the low hill to the south of the small creek, and numerous small outcrops may be examined along the south side of the creek valley or on the hill itself. A small quarry on the northwest edge of the hillside exposes bedded Edgecliff facies, while a small rugosan mound is located just to the southeast of the quarry among the trees. (See text for discussion). Return to cars follow Collins Road back to Route 20.

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STORM-DOMINATED SHELF AND TIDALLY-INFLUENCED FORESHORE SEDIMENTATION, UPPER DEVONIAN SONYEA GROUP, BAINBRIDGE TO SIDNEY CENTER, NEW YORK

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INTRODUCTION

The Upper Devonian paleoshoreline of the Catskill clastic wedge in New York State has been interpreted for nearly a century as a complex deltaic sequence (Barrel, 1913, 1914; Chadwick, 1933; Cooper, 1930; Sutton, Bowen and McAlester, 1970; and many others). Friedman and Johnson (1966) envisioned this deltaic complex as a series of coalescing deltaic lobes that progressively filled the Catskill epeiric sea and existed as an uninterrupted deltaic plain from New York to West Virginia. In addition, some geologists believe that such epeiric seas were tideless owing to rapid tidal wave attenuation (Shaw, 1964; and Mazzullo and Friedman, 1975). Others presume that storm (wave) processes were dominant with little or no tidal influence (Dennison, 1985).

This study offers significant departures from these interpretations, by documenting nondeltaic environments with significant tidal influence along the Catskill paleo-shoreline. The purpose of this study (Fig. 1) is to delineate sedimentary environments spanning the nonmarine to marine transition in the Upper Devonian Sonyea Group and to test and challenge previous deltaic models of the Sonyea Group (Sutton, et al., 1970). Recent publications have introduced evidence for non-deltaic shoreline environments within the Catskill clastic wedge (Walker and Harms, 1971; 1975; Bridge and Droser, 1985; VanTassel, 1986; also see Sevon, 1985, Table 1, p. 83). Others have shown that tidal processes were significant in the Catskill sea (Slingerland, 1986; and Bridge, et al. 1985).

Figure 2 shows the paleogeography of eastern North America during Frasnian-Fammenian time. Paleogeographic reconstructions indicate that the Acadian Mountain range supplied large quantities of sediment westward, allowing a vast alluvial plain to develop. The alluvial plain was fronted by deltaic lobes which prograded westward into the Catskill epeiric sea.

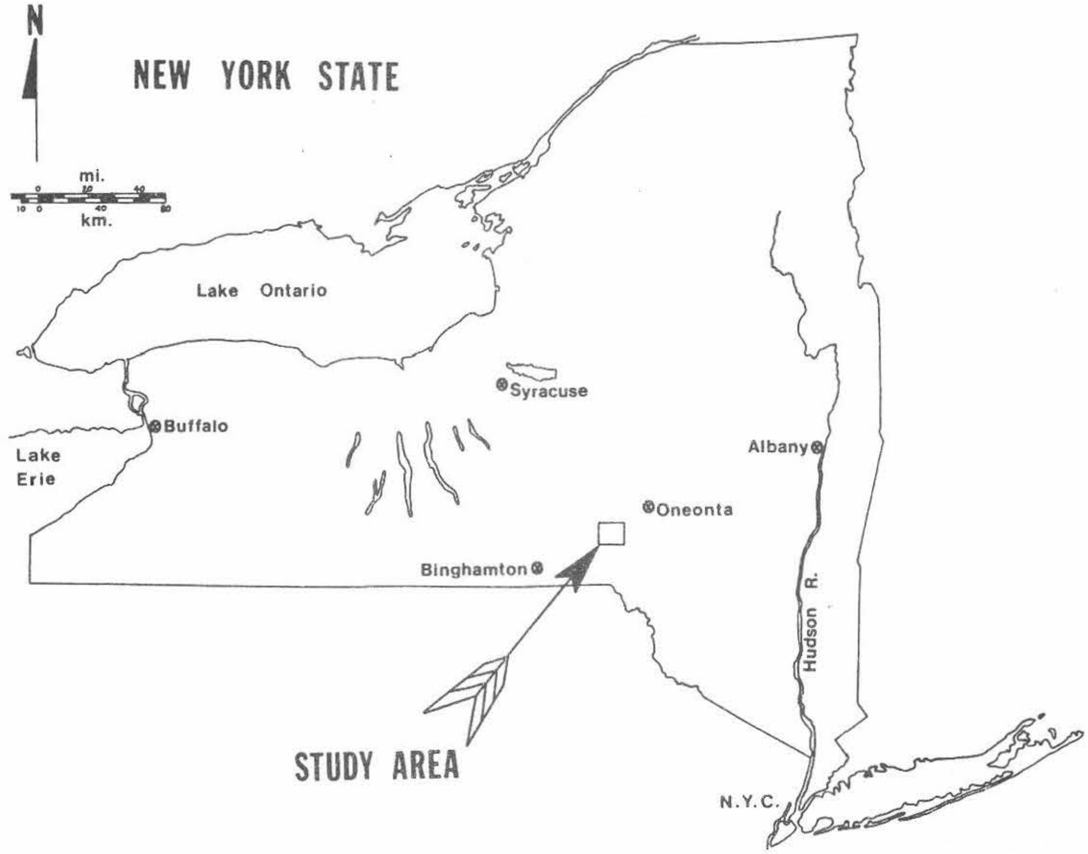


Figure 1: Index map of New York state showing the location of the study area.

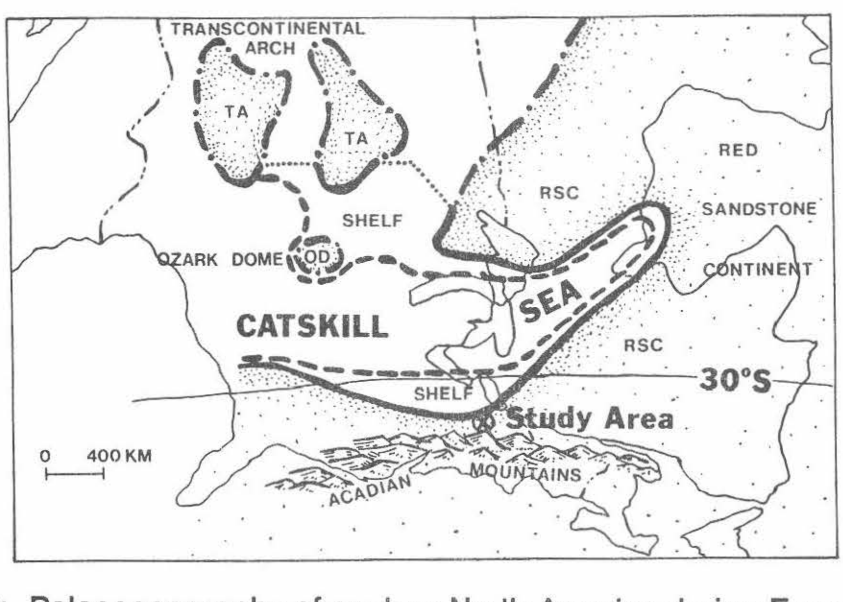


Figure 2: Palaeogeography of eastern North America during Frasnian-Fammenian time adapted from Haeckel and Witzke (1979), as cited by Slingerland (1986). The study area is shown in relation to the Acadian mountains and the Catskill sea. Notice the study area is positioned approximately 30 degrees south of the equator. OD= Ozark Dome; TA = Transcontinental arch; and RSC = Red sandstone continent.

Despite the compilation of a tremendous library of geologic information, very few papers specifically investigated the depositional environments along the paleoshoreline of the Catskill clastic wedge. Early workers may have biased the scientific community into thinking that deltaic lobes extended along all portions of the paleoshoreline (Barrel, 1913, 1914; Chadwick, 1933; Cooper, 1930; Friedman and Johnson, 1966). Such simplification is regrettable and is probably in error. The sequence of facies in this study suggests nondeltaic progradation of the Catskill clastic wedge during Sonyea time.

Location of Study Area and Methods

The study area is located in south-central New York in the extreme northwest corner of Delaware county and eastern-most Chenango county (Fig. 1). Most outcrops are located in the area of Bainbridge, Sidney, and Sidney Center, New York (Fig. 3).

Thirty-six outcrops were measured during the summer and fall months of 1987. Sequences of sedimentary structures and lithology, faunal content, biogenic structures, and soft sediment deformation were used as the basis for interpretation. Refer to Bishuk (1989) for detailed measured sections. Additional sections are being described by Applebaum (in prep.).

A Brunton compass and metric tape measure survey was conducted at the Sidney Center outcrops located at the intersection of Dunshee Road and Delaware County Route 35 (Fig. 4) to establish the stratigraphic succession of key outcrops in this area of limited exposure.

Stratigraphy

The Sonyea Group is the second oldest of seven groups in the Upper Devonian in New York State (Fig. 5). Present Sonyea Group stratigraphy was redefined by Sutton, Bowen, and McAlester (1970). The lower and upper group boundaries and several formational contacts are defined by thin, laterally persistent, black shale tongues. The Sonyea Group includes the rocks lying between the base of the Middlesex-Montour black shales and the base of the overlying Rhinestreet-Moreland black shales. The lithology and thickness of the Sonyea rocks between these black shale varies greatly with a general coarsening and thickening from west to east. The Sawmill Creek Shale divides the eastern portion of the Sonyea Group into two strikingly similar units, the Triangle Formation (lower), and the Glen Aubrey Formation (upper). Both formations consist of small-scale repetitive fining-upward sequences of marine shelf sandstone, siltstone, and shale. These formations differ only in their distribution, abundance, and type of invertebrate fauna. Below the Sawmill Creek Shale, the dominant taxa of the Triangle Formation include the brachiopods Productella, Mucrospirifer, and Leirorhynchus, the gastropod Bellerophon, and the bivalve Palaeoneilo (Sutton, et al., 1970). Above the Sawmill Creek Shale, the Glen Aubrey Formation displays a fauna that includes the bivalve Cypricardella, the brachiopods Cupularostrum, Tylothyris, Chonetes, Rhipidomella, Platyrachella, and Ambocoelia, and crinoid debris (Sutton, et al., 1970).

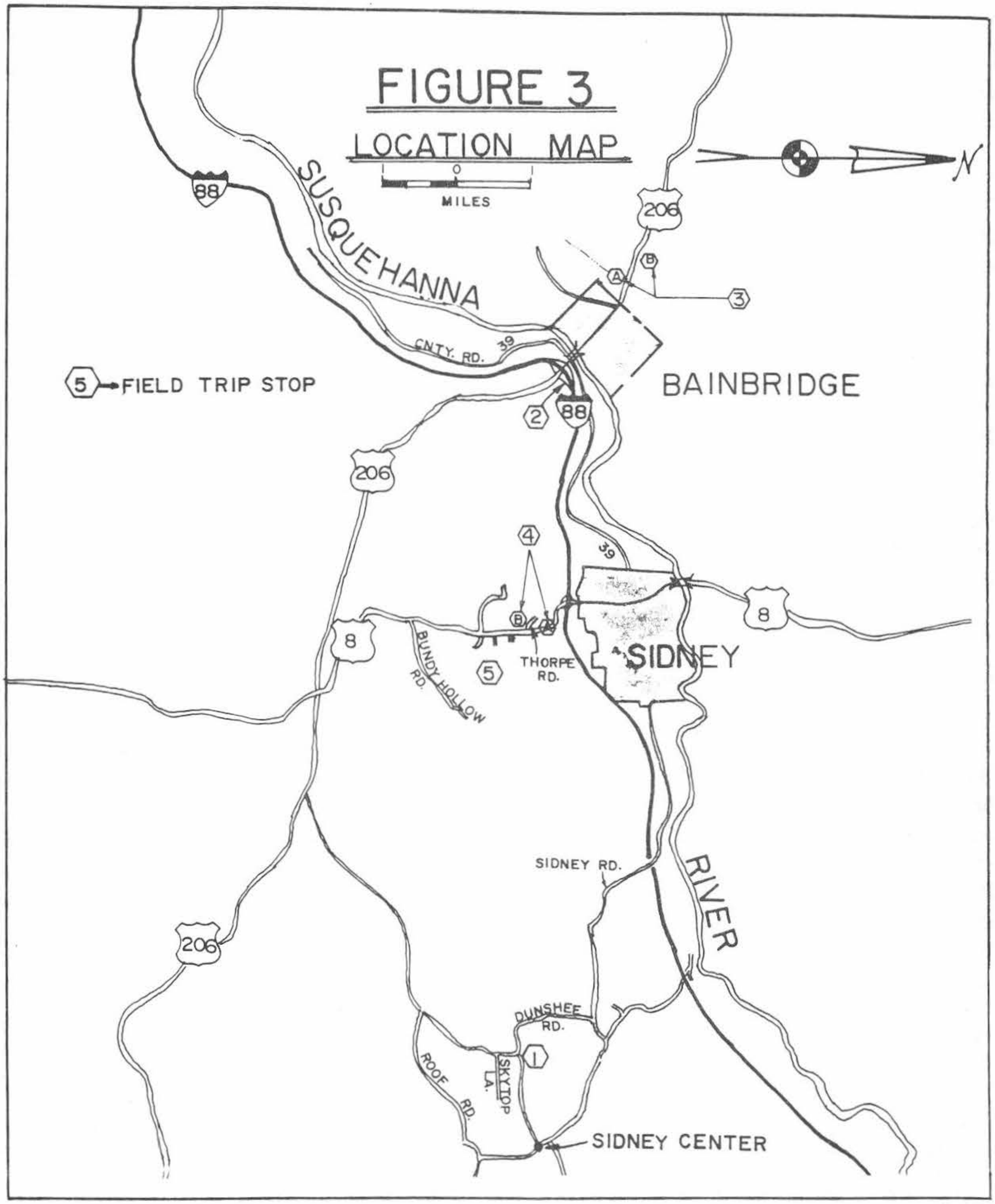
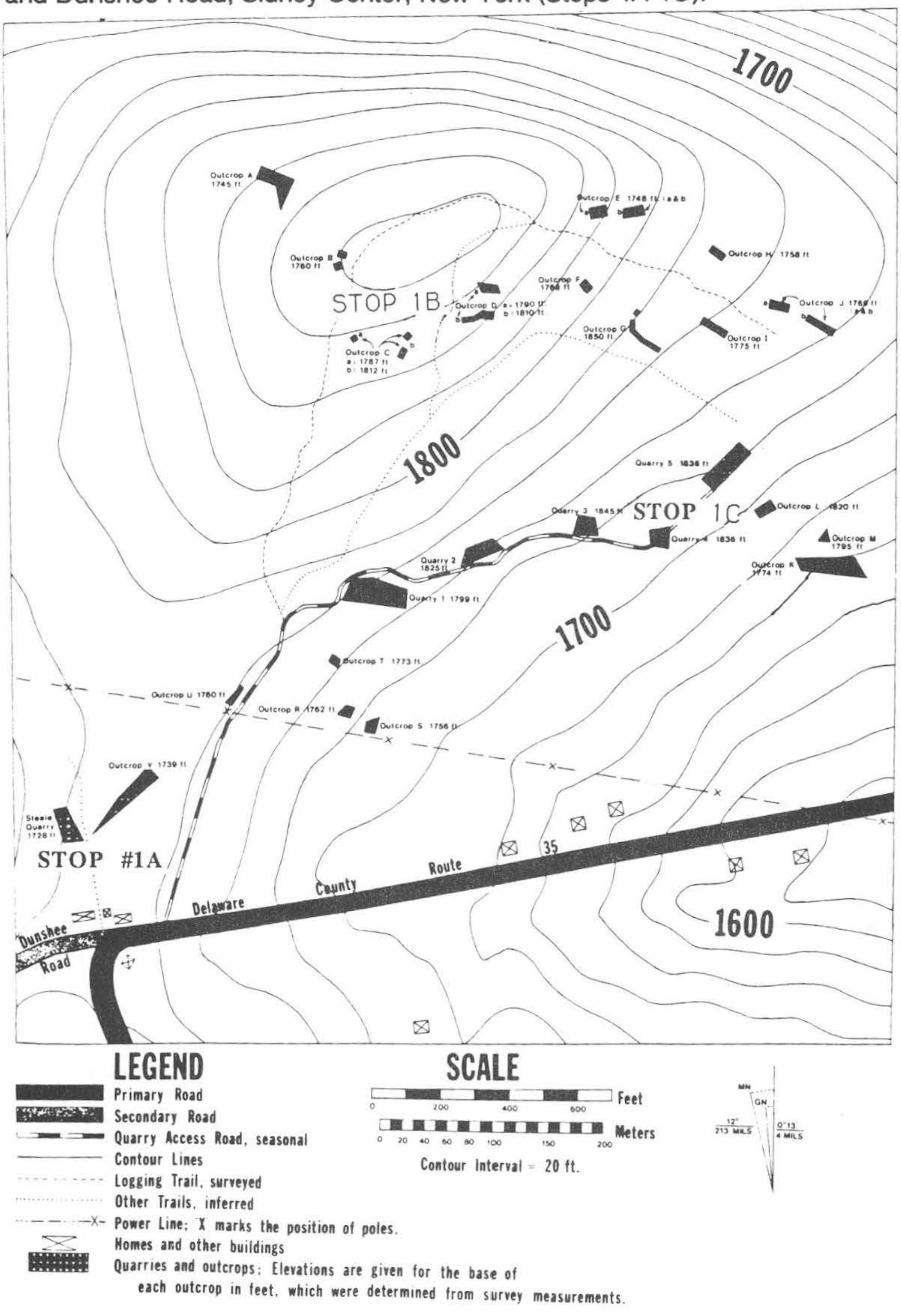


Figure 3: Roadmap of the Bainbridge, Sidney, and Sidney Center areas showing the location of the field trip stops.

Figure 4: Location of outcrops at the intersection of Delaware County Route 35 and Dunshee Road, Sidney Center, New York (Stops 1A-1C).



Black shales of the Sonyea Group are interpreted by Sutton and others (1970) as long transgressive intervals accompanied by low siliciclastic input to the shelf. Alternative interpretations of these units are included in this study.

Strata within the study area are probably assignable to the Glen Aubrey Formation (marine) based on faunal content, and the Walton formation (nonmarine) (Fig. 5). However, stratigraphic placement is difficult owing to limited exposure. Nonmarine to marine transition rocks are found within an unnamed unit assignable to the Cattaraugus facies (Fig. 5). Sections farther west at Bainbridge are likely in the Triangle Formation.

MARINE ROCKS (Chemung Facies)

Six facies that record fully marine to nonmarine transitional environments have been delineated from detailed measured sections of outcrops in the study area. These facies are shown in a composite section to show stratigraphic context and relative thicknesses (Fig. 6). See Bishuk (1989) for detailed measured sections of individual outcrops within each facies.

Two facies record deposition within the marine shelf of the Catskill Sea. These are: 1) the hummocky cross-stratified facies; and, 2) the dark gray shale facies. Exposures of these facies are more abundant than facies higher in the section, but all facies are still difficult to trace laterally. Limited exposure has made facies reconstruction difficult. Recent road construction has exposed several new outcrops since the study of Sutton, Bowen, and McAlester (1970) alleviating some of this difficulty.

Description of the hummocky cross-stratified facies

The hummocky cross-stratified facies consists of very fine sublitharenite, with common interbeds of siltstone and shale. Fine sand sublitharenite and conglomerate composed of shale and siltstone clasts are rare in the stratigraphically lower sections, but are dominant stratigraphically higher in this facies. Sandstones are moderately well- to well-sorted. The hummocky cross-stratified facies is the lowest unit in the stratigraphic sequence. It is present along Route 8, just south of Sidney, and at outcrops along Interstate 88 at Sidney and Bainbridge and points westward.

Hummocky cross-stratification is the dominant sedimentary structure. Hummocky cross-stratification is primarily found interbedded with siltstone and shale (Fig. 7), but also occurs as amalgamated beds of very fine sublitharenite (Fig. 8). Hummocky cross-stratified beds have sharp bases, with moderately rare directional sole marks such as flute and tool marks. Mean paleocurrent direction for flute and tool marks is 275 degrees.

Coquinite layers, coquinite-filled scours, coquinitic hummocks or ripple forms, graded bedding and conglomerate are commonly found at the base of hummocky cross-stratified sandstone. Coquinite layers are the most common of these features

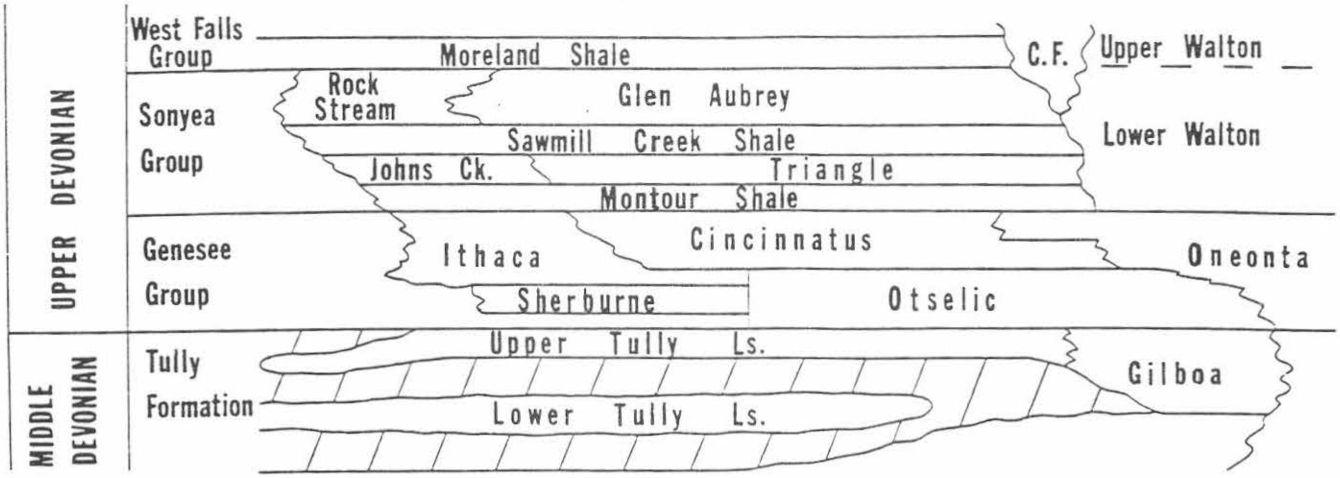


Figure 5: Stratigraphic correlation of part of the Middle and Upper Devonian in New York. Diagram simplified from Rickard (1975). The study was conducted within the Glen Aubrey and Triangle Formations (marine) and the Lower Walton Formation (non-marine). Nonmarine to marine transition rocks are best assigned to an unnamed portion of the stratigraphy designated as the Cattaraugus Magnafacies (C.F.). Diagonal ruling represents hiatus.

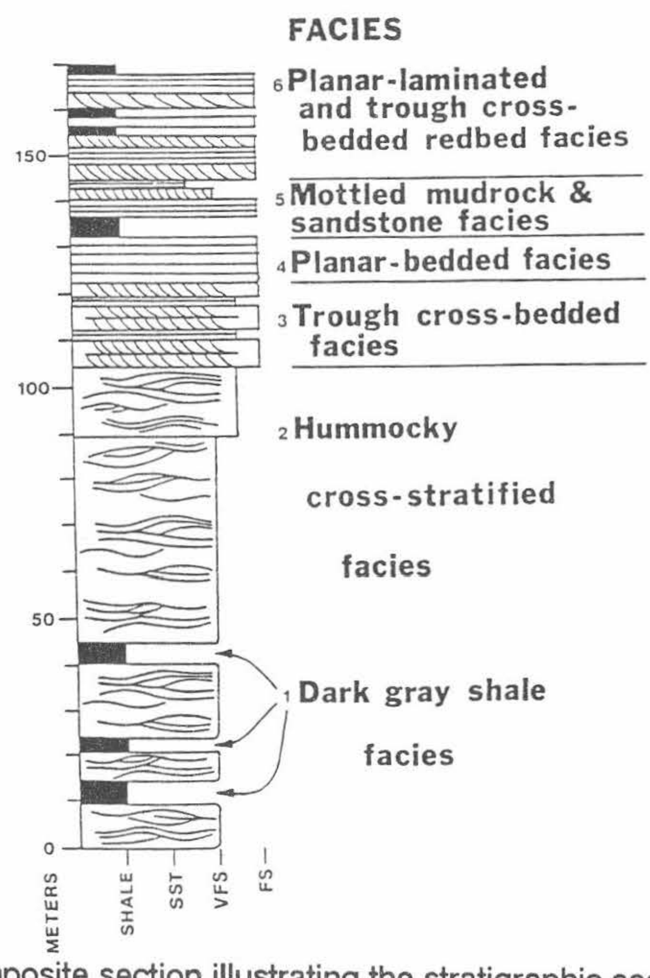


Figure 6: A 170 meter composite section illustrating the stratigraphic sequence of the six facies described in the study area.

Figure 7: Interbeds of hummocky cross-stratified sandstone and shale. The top-most sandstone bed (next to hammer) illustrates an excellent hummock and concave-up and convex-up laminations. The bottom-most shale marks the position of the first reoccurrence of marine fauna at the top of Sidney Mountain quarry (see interpretation of mottled mudrock and sandstone facies for details). The hummocky cross-stratified facies disconformably overlies the planar-laminated and trough cross-bedded facies here, after a brief half meter covered interval (Stop 5).

Figure 8: A large, amalgamated, hummocky bedform with swale found in the amalgamated portion of the hummocky cross-stratified facies higher in the section. The crest to trough distance is 2.3 meters. Hammer and meter stick for scale (Stop 1A).

Figure 9: Close-up of Figure 8 showing internal structure of hummock. Climbing ripple cross-lamination in left-center of the photo are inclined at a steep angle of propagation, which suggests rapid deposition by a storm event with no post-storm reworking. Hammer for scale (Stop 1A).

Figure 10: A solitary wedge-shaped form is cut into shale and subsequently filled with shale. Basal surface of the wedge is listric in nature. This is a characteristic feature of the dark gray shale facies. Hammer for scale (Stop 4A).

Figure 11: Interbedded trough cross-bedding and planar-bedding found within the trough cross-bedded facies. Individual laminae on cross-beds are lenticular and occasionally rippled. Cross-bed sets are consistently inclined in the same direction (toward left side of page), which probably records the ebb flow tidal direction. Ripple cross-lamination resting on troughs are inclined in the opposite direction of the troughs. This strongly suggests current reversal induced by tides. Planar laminated interbeds are interpreted as swash bars within tidal inlets (Stop 1C).

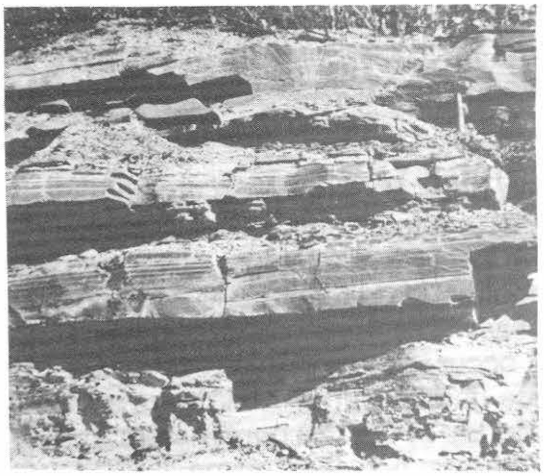


FIGURE 7

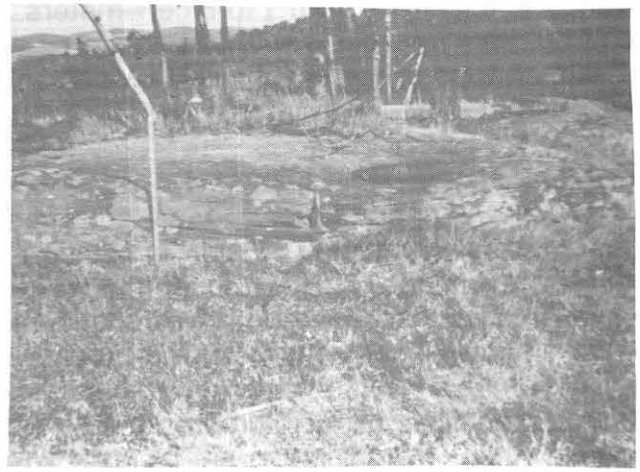


FIGURE 8


FIGURE 10

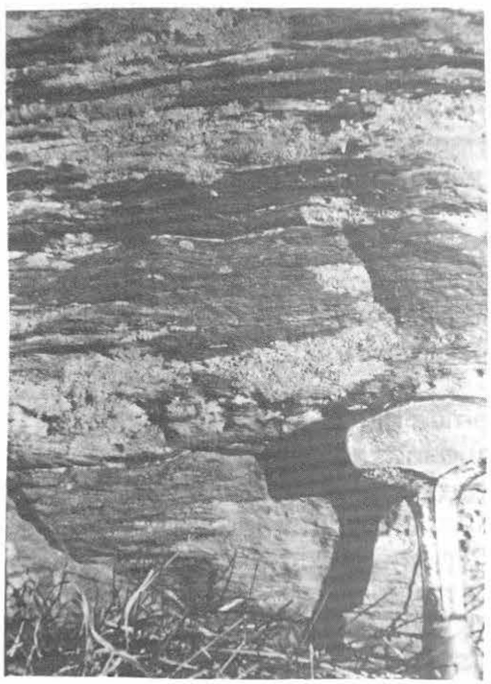


FIGURE 9

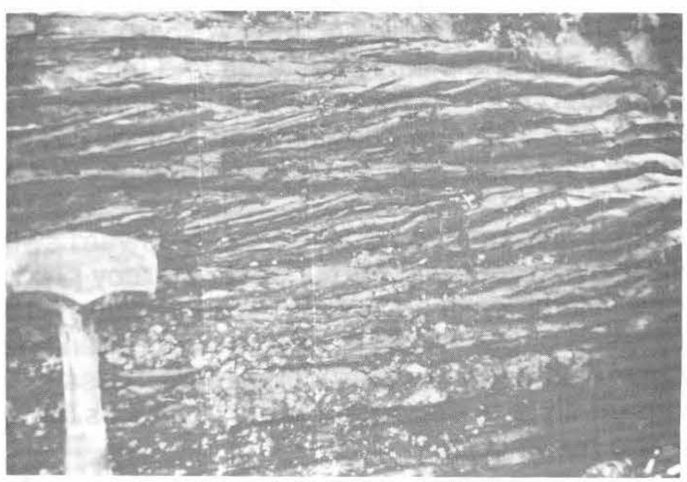


FIGURE 11

and range in thickness from 1 to 8 centimeters. These layers are not restricted to the base of hummocky cross-stratified beds. They occur less commonly as thin drapes or scour fills on tops of hummocks and swales and along truncations within amalgamated hummocky cross-stratified beds.

Laminations within hummocky cross-stratified beds show several variations, each with a regular sequence. Above coquinite layers, a few centimeters of planar lamination typically occurs. Planar lamination, with gentle undulations, often overlies low angle truncations. This is succeeded by low-angle curved laminae illustrating both concave- and convex-upward laminations, commonly with a form concordant style of deposition. Low-angle truncations may or may not occur. Hummocky cross-strata commonly dip less than 10 degrees. This sequence occurs within hummocks and swales of sandstone beds. Hummocks have amplitudes of 5 to 20 centimeters above adjacent swales. Wavelengths are typically 2 to 5 meters; however, some wavelengths of tens of meters have been observed. Hummocks and swales are commonly capped with slightly asymmetrical ripples and symmetrical wave ripples, which commonly contain transverse ribs or exhibit chaotic patterns. Paleocurrent readings on asymmetrical ripples average 260 degrees. Cross-lamination within ripples is usually not well preserved.

Lateral transitions are common within a single hummocky cross-stratified bed. The most common lateral transition is from hummocky cross-stratification as described above to planar laminae capped by wave ripples. Hummocky cross-stratification also changes laterally to wave-ripple laminae. Soft sediment slumps with internal hummocky cross-stratification and troughs with ball and pillow structures generally thin laterally to wave ripple-laminated beds or planar laminae capped by wave ripples. Craft and Bridge (1987) cite similar lateral transitions within hummocky cross-stratified beds.

Siltstone and shale interbeds are planar-laminated, blocky, or structureless, and commonly heavily bioturbated. However, siltstone interbeds may exhibit small-scale, current ripple cross-lamination, which is commonly discontinuous and cryptic.

The style of hummocky cross-stratification becomes dominantly amalgamated higher in the stratigraphic section. A sharp, scoured and loaded contact with numerous ball and pillow structures and intraformational conglomerate occurs where amalgamated hummocky cross-stratification commences. Very fine and fine sublitharenite with rare conglomerate composed of shale and siltstone clasts are dominant lithologies. Siltstone and shale interbeds are rare.

In amalgamated beds, concave-upward swaley surfaces pass laterally into convex-upward hummocky surfaces, thereby producing adjacent hummocks and troughs. Hummocks and swales are truncated laterally by hummocky cross-stratification or planar strata. Pebble lags often line these erosional surfaces. Most amalgamated beds are form-concordant. Less commonly, amalgamated beds show a discordant style of deposition with hummocks filling in underlying swales and swales overlying hummocks.

Heights of the hummocks are up to 20 centimeters above adjacent swales. Wavelengths range from 2 to 5 meters. Crests of hummocks are straight to slightly sinuous with a mean orientation of 128 degrees. Structures, internal to amalgamated hummocky bedforms, include climbing ripples with steeply inclined axes of propagation (70 to 80 degrees) along the flanks of hummocks (Fig. 9). The ripples generally show migration toward the crest of hummocks. However, there is evidence for rare current reversal recorded by ripple migration toward swales. Hummocks and swales commonly contain several horizons of Rhizocorallium (?) burrows and sinuous epichnial trails of Scalarituba in convex hyporelief. Amalgamated hummocky beds rarely have distinct sole marks along erosional bases. Where present, sole marks include flute casts, load casts, and a variety of tool marks.

Ball and pillow structures are abundant at or near the base of hummocky cross-stratified beds and display minimal penetration into underlying lithologies. Penetration and detachment of ball and pillow structures into underlying siltstone and shale interbeds and development of flame structures are less common, but occur in abundance along some horizons. Hummocky cross-stratification is preserved rarely within ball and pillow structures and soft sediment slumps.

Other structures of minor occurrence are generally found capping hummocks and swales. These minor structures include wrinkle marks (also known as "runzel marks"), asymmetrical ripples with or without transverse ribs (commonly found as ridges and furrows), and scour depressions with or without symmetrical wave ripples within depressions. Mean paleoflow from asymmetrical ripples is 359 degrees (Fig.13, Steele quarry locality). Paleoflow direction is indicated for scour depressions by their elliptical shape, with a mean directions of 160 or 320 degrees. Rare trochoidal ripples are only present at the Steele quarry in Sidney Center.

The hummocky cross-stratified facies has a diverse invertebrate fauna. All fossils have been transported and are never found in life position. However, articulated crinoid stems are common within coquinite lenses and intervening shale at the Bainbridge exit on Interstate 88 (Stop 2). Locally, Tentaculites are oriented with a mean paleocurrent direction of 245 degrees. Bivalves (Palaeoneilo, Nuculoidea, and Cypricardella), brachiopods (the productid, Productacea; the orthid, Tropidoleptus; the rhynchenellid, Cupularostrum; and the spirifers, Platyachella and Mucrospirifer) and carbonized plant fragments are also locally common in siltstone and shale interbeds.

The abundance of invertebrate fauna varies throughout the facies. Fauna are locally abundant, sparse or absent. This fluctuation in fossil content occurs in cyclic patterns within the facies, Spiriferids, rhychonellids, orthids, atrypids, bivalves, crinoids, gastropods, and Tentaculites dominate where fauna are abundant. The rhychonellid, Cupularostrum, and the spiriferid, Platyrachella, far outnumber all other taxa. Productids and diminutive individuals such as the bivalves Paleoneilo and Nuculoidea persist in the sparsely fossiliferous intervals.

Fossils decrease markedly in abundance and diversity where amalgamated hummocky cross-stratification dominates the section. However, carbonized plant fragments are abundant. The most common fossil is the rhynchennelid, Cupularostrum. Other fossils, in order of decreasing abundance, include the bivalve, Sphenotus, the spiriferid, Platyrachella, other unidentifiable brachiopod fragments, fish fragments, crinoid ossicles, and the bivalve, Cypricardella. The fossils are most often found associated either with loaded scour and fill bases as coquinite shelly lags or within intraformational conglomerates consisting mostly of flat shale intraclasts.

The cumulative thickness, in which amalgamated hummocky cross-stratification occurs, is approximately 7 to 15 meters. Similar thicknesses are reported by McCrory and Walker (1986) and Walker (1984).

Interpretation of the hummocky cross-stratified facies

Hummocky cross-stratification in the study area is interpreted as a storm-produced structure occurring below fair weather wave base and above storm wave base (Harms, 1975, and many others). The lower portion of the hummocky cross-stratified facies is therefore interpreted as representing deposition on a storm-dominated shelf.

Harms, Southard, and Walker (1982) state that hummocky cross-stratification forms under high-velocity oscillatory flow conditions as a continuum from 2-D wave ripples to hummocks of increasing wavelength to plane beds. Observed spatial variations in sedimentary structures are consistent with this interpretation. Climbing ripples, found within the amalgamated hummocky cross-stratified portion of this facies, augment an interpretation of rapid deposition of sediment by storms (Fig. 9). Steep inclination of climbing ripple propagation and the form-concordant style reflects vertical growth of hummocks with minimal migration, suggesting a dominant oscillatory flow and rapid deposition rate. Similar interpretations have been made by Craft and Bridge (1987) for rocks of the type Chemung facies.

The rare occurrence of current reversal in climbing ripples (Fig. 9) suggests disequilibrium in hummock growth with occasional migration down-flank toward swales. Rhyzocorallium (?) burrows are concentrated at various horizons within amalgamated hummocky beds (Fig. 9), which demonstrates the amalgamated nature of these intervals.

Sole marks are most easily observed in the lower portions of this facies where shaly interbeds are most common. Tool marks and groove casts trend east-west to east northeast-west southwest which is normal to the paleoshoreline in the paleogrogrpahic reconstructions of Barrell (1913, 1914) and Chadwick (1933). Rare flutes and asymmetrical tool marks record paleoflows which were directed offshore (west to west southwest). Following Duke (1990), we interpret these structures as recording high bed shear stresses produced by oscillatory flow in the inner boundary layer during storms (see also Duke, Arnott and Cheel, 1991). Offshore-directed structures record augmentation of bed shear stress on the offshore stroke of waves by geostrophically balanced coastal downwelling (Duke, 1990, Duke, et al., 1991).

Ball and pillow structures result from an inverse density gradient and low shear strength associated with high pore-water pressure (Allen, 1982, v. 2, p. 363). It is unclear from field evidence whether the high pore-water pressure was induced by rapid deposition and/or storm wave- or seismic-induced pressure pulses (Craft, et. al., 1987). The fact that some major ball and pillow horizons show minimal penetration into underlying lithologies is significant This lack of penetration suggests that the underlying siltstones and shales may have been at least semi-cohesive, which hampered soft sediment deformation. Detached balls and pillows in other parts of the measured section reflect rapid deposition causing liquefaction of underlying siltstones and shales and subsequent penetration of sand.

Most coquinite lenses are interpreted as postmortem storm-transport of shells to areas below fair-weather wave base (Sutton, et al., 1970). Articulated crinoid stems within coquinite lenses capping hummocks and swales and within intervening shale between hummocky cross-stratified beds suggests some in situ burial of crinoids. In addition, the presence of coquinite layers on both top and bottom of a single hummocky cross-stratified bed implies that the bed is amalgamated and that crests of hummocks may have been periodically modified by subsequent storms.

The cyclic patterns of fluctuation of invertebrate abundance, faunal change, and levels of biogenic activity is probably produced by fluctuations of the pycnocline (Byers, 1977), and may provide clues to relative rates of progradation and subsidence (Thayer, 1974). High diversity of taxa and moderate to low levels of biogenic activity indicate aerobic conditions and implies a relatively deep oxygen mixing depth. Low numbers of taxa and higher levels of biogenic activity indicate dysaerobic conditions and a shallower oxygen mixing depth (Byers, 1977). Marine transgressions are not likely to account for this cyclicity, because the cycles are too frequent. The cyclicity may have been induced by: 1) an event controlled cyclicity, in which storms dump fauna and oxygenated sediment into deeper dysaerobic zones; or 2) differential subsidence controlled by changing rates of progradation may also account for pycnoclinal fluctuations. Other less likely alternative explanations include sea-level rises that drove the shoreline eastward (Dennison, 1985) or sea-level rise in response to epeirogenic lithospheric downflexure causing basin-wide subsidence, along with interbasinal arches and domes, which existed in fluctuating submergent and emergent conditions (Quinlan and Beaumont, 1984).

The stratigraphic position of this facies, the predominance of fine sand sublitharenite, and a marked upward decrease in fauna suggest that the amalgamated hummocky cross-stratified portion of this facies occupies the lower shoreface. Siltstone and shale interbeds are rarely preserved, indicating that waves or currents (e.g., longshore-, tidal-, and storm-driven) had effectively winnowed the fine fraction (Swift, 1984). The abundance of amalgamated hummocky cross-stratification implies that storms were more frequent, thereby providing additional winnowing. The sharp, scoured and loaded contact at Thorpe road (Stop 4B) is interpreted as the contact between lower shoreface and shelf deposits. Paleoflow direction is consistently to the north, which may represent the longshore or geostrophic current direction. The marked decrease in faunal abundance indicates environmental stress associated with the nearshore zone (Thayer, 1974).

The hummocky cross-stratified facies is integrally involved in the overall progradation of the Catskill clastic wedge. Vertical upbuilding of hummocky cross-stratified beds at rates greater than the average rate of subsidence causes shallowing in the nearshore zone (Hamblin and Walker, 1979). This provides conditions conducive for rapid progradation of shoreface and foreshore deposits over the hummocky and swaley cross-stratified facies. The dominance of hummocky cross-stratified beds offshore warrants against an interpretation of deltaic deposits in the nearshore. Frequent storms would inhibit outbuilding of deltaic lobes and argues for a straighter shoreline.

Progradation ceased when interrupted by marine transgression and when tectonic conditions and/or the weight of nearshore deposits were sufficient to cause rapid subsidence. This is substantiated by hummocky cross-stratified beds overlying nonmarine, fluvial rocks at Sidney Mountain quarry (Stop 5, Fig. 7). Since hummocky cross-stratification is chiefly deposited below fair weather wave base, the nonmarine deposits subsided to an approximate minimum of 10 meters below sea level. Subsidence alone may not account for this, so it was probably coupled by a slight marine transgression. A major river avulsion and/or a directional change in dispersal of sediment from the source area may have contributed to abandoned or diminished nearshore deposition, which would allow subsidence to outpace accumulation.

The abrupt deepening apparent at Sidney Mountain quarry can not be attributed to Milankovitch cycles of the PAC hypothesis (Punctuated Aggradational Cycles, Goodwin and Anderson, 1985), because the study area lacks good stratigraphic control and exposure, which are essential criteria to establish PAC boundaries. Although Van Tassell (1987) established PAC boundaries in similar Frasnian-aged deposits of the Brallier, Scherr, and Foreknobs Formations in the Catskill Clastic wedge in Virginia and West Virginia, evidence is lacking in the Sonyea Group of New York.

Description of the dark gray shale facies

The dark gray shale facies occurs as three distinct intervals within the hummocky cross-stratified facies at Route 8 in Sidney (Stop 4A, Fig. 6). Lithologies are generally restricted to uniform dark gray shale and siltstone, with rare lenses of very fine sand sublitharenite. Biogenic structures and invertebrates are absent. Planar bedding is most common in this facies. Discontinuous, ripple cross-lamination in siltstone is cryptic, but occurs in all three shale units of this facies. Similar ripple cross-laminae are also reported by Hamblin and Walker (1979).

Trough and wedge-shaped forms characterize this facies. Basal surfaces truncate underlying planar-laminated shale, and are overlain by a wedge of shale (Fig. 10). Most of the truncation surfaces occur as solitary wedges, which are characterized by a sharp concave-up discontinuity surface. Truncation surfaces have a smooth, listric ("spoon-shaped") geometry. The shale is inclined along the discontinuity surface and is in angular discordance with underlying planar beds. Inclination of shale decreases to subhorizontal to horizontal planar laminations progressively upward within the wedges. Truncation surfaces also occur as facing pairs of intersecting, U-shaped troughs filled with shale that truncate each other (see route 8 measured section, 10.6-11.6 meters [Bishuk, 1989]). The shale that fills the U-shaped troughs conforms to its U-shape, and progressively flattens upward within troughs. Most wedges and troughs measure 5 to 20 meters in width, and truncate 1 to 2 meters of underlying shale. Applebaum (in prep.) has observed similar bedding geometries in mudstones at Bainbridge (Stop 2).

Similar truncation surfaces possessing a listric geometry have been reported by Davies (1977). However, there are two differences between the surfaces found in this study and those found by Davies (1977). The features in this study are much smaller in scale, and Davies (1977) does not recognize any facing pairs of U-shaped troughs.

Interpretation of the dark gray shale facies

The dark gray shale facies records three brief transgressive periods, which represent desposition on deeper portions of the shelf. This is substantiated by the complete absence of invertebrates and biogenic structures, and the predominance of silt and clay. The aerobic to dysaerobic conditions of the hummocky cross-stratified facies repeatedly alternates with anoxic conditions of the dark gray shale facies. Alternations are explained by fluctuations of the pycnocline to a more landward position, causing basinal anoxic conditions to briefly develop in areas which are normally oxygenated. Repeated shifts produced the alternation of euxinic environments of the dark gray shale facies and fossiliferous aerobic to dysaerobic sediments of the hummocky cross-stratified facies. Byers (1977) has reported similar pycnoclinal shifts within the Middlesex Shale in the distal portions of the Sonyea Group in western New York.

The factors that cause the pycnoclinal fluctuations are still problematic, so only some of the possible causes will be discussed here. Initial epeirogenic downflexure of the crust may have formed a deeper-water trough of the dark gray shale facies during the Acadian collision (Quinlan and Beaumont, 1984). The pycnocline adjusts simultaneously to this lithospheric downflexure by migrating to a more landward position. This was intermittently counteracted by sedimentation of the hummocky cross-stratified facies despite continued isostatic subsidence and sediment loading. Vertical upbuilding of hummocky beds and progradation rates were probably so rapid as to only record brief, euxinic conditions in the nearshore zone. During periods of frequent storms and rapid progradation rates, the pycnocline readjusted to a more seaward position. Similar interpretations of initial tectonic downflexure have been made in the Antler foreland basin of Nevada (Harbaugh and Dickinson, 1981) and in other groups of the Paleozoic Appalachian basin of the eastern interior of North America (Quinlan and Beaumont, 1984).

Other factors involved in oxygen mixing and density stratification associated with pycnoclines include control by wave base, tides, and climatic variability. Larger waves would allow deeper penetration of oxygenated water. The relation between oxygen decrease and depth is also influenced by any lateral influx of oxygenated water across a deep sill and by the input of organic material from surface production or sediment gravity flows from the basin margin which consumes oxygen at depth (Byers, 1977).

Isostatic and/or eustatic rise and fall of sea level may have also contributed to pycnoclinal fluctuations (Johnson, Klapper, and Sandberg, 1985). However, further evidence is needed to establish the magnitude and frequency of sea level change needed to account for frequent facies alternations. Eustatic marine transgression(s) operating alone is a less attractive explanation to account for the cyclic nature of the dark gray shale units. It would imply 3 transgressive-regressive events over a short interval of time. Frequent sea level change is not known to have occurred during Fammenian time (Haeckel and Witzke, 1979). Alternatively, the hummocky cross-stratified facies and the dark gray shale, facies represent laterally migrating, subjacent environments stacked vertically according to Walther's Law (Fig. 22).

The truncation surfaces that form shallow wedges and troughs are probably rotational gravity-slide failure scarps (Fig. 22). Large quantities of unlithified sediment were disaggregated during liquefaction and removed by gravity sliding, owing to the fact that large-scale breccia, rotated blocks, or crumpled or other disturbed bedding structures are absent in the sediment wedge above truncation surfaces. Trough-shaped truncation surfaces found in this facies are typical of submarine gravity slides. Overall, a gravity-slide mechanism is favored over an alternative interpretation of erosional channel forms that form a network of subaqueous distributaries for reasons defined by Davies (1977). A summarization of the criteria used by Davies (1977) that support a gravity-slide mechanism for this study include the following:

1) The listric geometry of the truncation surfaces is typical of landslide and other gravity-slide structures.
2) The sharp and regular truncations without any obvious local channels or erosional irregularities is best explained by shear rather than by current scour.
3) No radical change in depositional processes are detected from overlying fill to the truncated rocks, which both have a shale lithology. Therefore, there is no evidence for a period of increased current scour.
4) The absence of a coarse, basal lag above truncation surfaces discounts the formation of channels by a strong traction current and favors gravity-slide failure. Alternatively, coarse material may not have been present to form basal lags.

The rotational slumps were probably triggered by storm events, overloading, or seismic activity causing liquefaction and foundering of the sediment.

The occurrence of discontinuous ripple cross-lamination implies an environment above storm wave base. The cross lamination may also suggest reworking by storms which leave cryptic signatures in deep portions of the shelf during waning stages of storm events. Hamblin and Walker (1979) find similar ripple cross-lamination between hummocky cross-stratified beds of the Fernie Formation in the Rocky Mountains of Alberta.

The lowest shale unit depicted on Figure 6 is located at the Route 8 locality in Sidney and contains the darkest shale of all of the black shale units and lacks evidence of bioturbation. This unit bears closest resemblance to the characteristics of the Sawmill Creek Shale, a laminated and barren dark gray shale (Sutton, Bowen, and McAlester, 1970). The Sawmill Creek Shale is interpreted by Sutton and others (1970) as recording a marine transgressive event. We question the liklihood of unambiguous recognition of the Sawmill Creek in the study area because many shale intervals meet the general description of Sutton et al. (1970). A much more detailed definition is needed for this key stratigraphic marker.

Sutton and others (1970) place the Sawmill Creek Shale at the Sidney Mountain quarry (Stop 5) in Sidney. The only unit even closely reminiscent of the Sawmill Creek Shale at Sidney Mountain quarry is the siltstone-shale interval nearest the base of the quarry. We argue that correlation of black shale tongues of the Sonyea Group such as the Montour, Sawmill Creek, and Moreland Shales into the nearshore environments is difficult and probably erroneous. Many units of this study resemble the Sawmill Creek Shale. Therefore, it is more likely that the shale in question is separated by a higher frequency of storm units in the nearshore or proximal tempestite regime (see Brett, C. E., et al., 1986).

ROCKS AT THE NONMARINE TO MARINE TRANSITION (CATTARAUGUS MAGNAFACIES)

Rocks at the nonmarine to marine transition are assigned to an unnamed portion of the Cattaraugus Magnafacies (Fig. 5). This portion of the Sonyea Group stratigraphy is problematic and revision should be considered. We recognize three facies within the Cattaragus Magnafacies. They are: the trough cross-bedded facies, the planar-bedded facies, and the mottled mudrock and sandstone facies (Fig. 6).

Description of trough cross-bedded facies

The trough cross-bedded facies is a 7 to 15 meters thick unit that consists of moderately well sorted, fine sand sublitharenite. The trough cross-bedded facies is laterally traceable from the north-east at Delaware County Route 27 in Sidney Center to the south-west at Pine Hill Road on Sidney Mountain (Figures 3 and 4). Outcrops of this facies occur at similar elevations (approximately 1720-1780) throughout most of the study area. The trough cross-bedded facies is absent or covered near Sidney Mountain quarry.

The base of this facies is erosional and locally marked by a coarse sandstone-supported conglomerate and/or pebble lag. Carbonized stems, branches and occasional logs are common at the erosional base and are abundant throughout the facies. Locally, crinoid ossicles and Cupularostrum line the bottoms of troughs at and just above the erosional base.

Small- to medium-scale trough cross-bedding is the dominant sedimentary structure in the facies. Dip angles on trough cosets range from 14 to 25 degrees. Cross-beds commonly climb at low angles of propagation. Some cross-beds contain normally graded, lenticular laminae with asymmetrical ripples which oppose the dominant cross-bed paleoflow direction (Fig. 11). The trough cross-beds are largely unidirectional, but locally are multidirectional. Symmetrical and asymmetrical ripples are uncommon. Rare occurrences of reformed ripples with small-scale herringbone cross-lamination are found at Roof Road (Fig. 12). The herringbone cross-lamination is confined to scours at the crests of ripples marked by reactivation surfaces. These ripples are thought to be reformed, based on their asymmetry opposite the dip direction of the primary cross-lamination within the ripple. The geographic distribution of paleocurrent directions obtained from asymmetrical ripples and trough cross-beds are shown as rose diagrams in Figure 13.

In general, cross-bed set thickness thins upward within the facies. In addition, dip angles on troughs decrease upward in the facies. Dune bedforms are uncommon, occurring sporadically in the study area.

Trough cross-beds are interbedded with gently inclined and horizontal planar lamination. In general, planar-laminated interbeds become more common upward in the facies.

The trough cross-bedded facies closely resembles the planar-laminated and trough cross-bedded redbed facies positioned stratigraphically higher. It is differentiated from it by its sparse marine fauna, including the bivalve, Sphenotus, the rhynchenellid, Cupularostrum, the spiriferid, Platyrachella, and crinoid ossicles.

Interpretation of the trough cross-bedded facies

The trough cross-bedded is best explained as a laterally accreting, tidal inlet sequence of a barrier island and/or strandplain/tidal creek system (Fig. 22). This interpretation is based on the predominance of trough cross-beds, evidence of current reversals with differing flow magnitude, a sparse marine fauna, and the stratigraphic position of this facies. The trough cross-bedded facies occurs stratigraphically above the hummocky cross-stratified facies (storm-influenced lower shoreface and offshore deposits) and stratigraphically below the planar-bedded facies (foreshore beach/barrier deposits).

The tidal inlet study of Ossabaw Sound, Georgia by Greer (1975) describes a plausible modern analog to the trough cross-bedded facies. Similar interpretations of ancient analogs include Carter (1978) and Leckie (1985).

The small to medium scale trough cross-beds record deposition by migrating dunes within the inlet or tidal channel. Davidson-Arnott and Greenwood (1976) describe similar lunate dunes.

There is subtle, but unmistakable, evidence for tides in the trough cross-bedded facies. Asymmetrical ripples that climb larger scale foresets (Fig. 11) and reformed ripples with herringbone cross-lamination record current reversal induced by tides (Fig. 12). Some trough cross-beds, dune bedforms, and asymmetricai ripples on troughs, exhibit a rare multidirectional component, particularly at the Roof Road locality

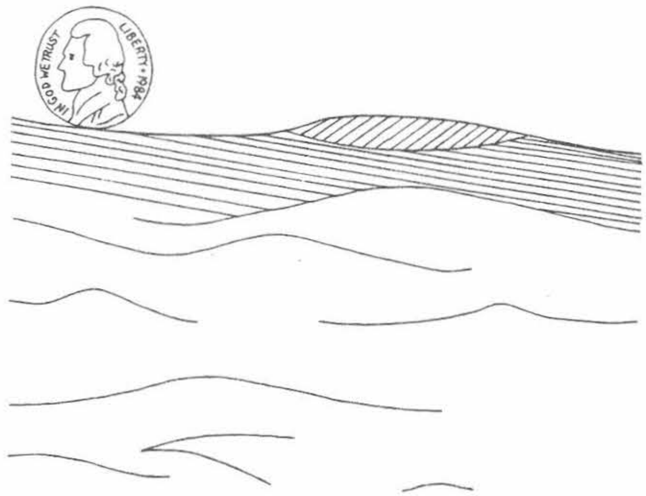


Figure 12: Reformed ripple with small-scale herringbone cross-lamination marked by a reactivation surace found within the trough cross-bedded facies. This reformed ripple suggests current reversal induced by tides. Other ripple forms shown are symmetrical to slightly asymmetrical. The diagram is a traced enlargement of a field photograph. Nickel for scale. Marvin residence, Roof Road, Sidney Center.

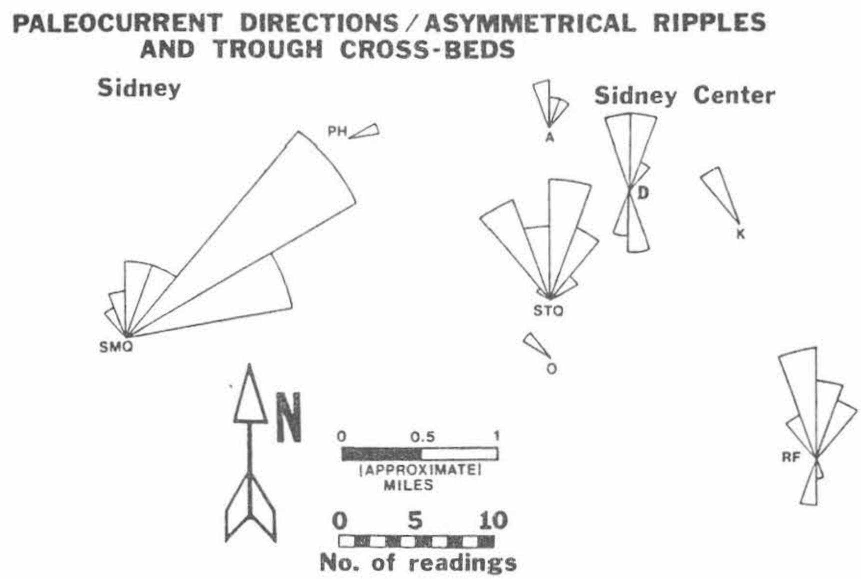


Figure 13: Rose diagrams depicting the geographic distribution of paleocurrent directions obtained from asymmetrical ripples and trough cross-beds (dune bedforms). The abbreviations represent outcrops in their approximate geographic location (see figures 3 and 4). A= Outcrop A, Figure 4 (trough cross-bedded facies); D= Outcrop D, Stop 1C, (trough cross-bedded facies) Figure 4; K= Outcrop K, Figure 4 (trough cross-bedded facies); O= Quarry #4, Figure 4 (Mottled mudrock and sandstone facies); PH= Pine Hill Road, Sidney Mountain (trough cross-bedded facies); SMQ= Sidney Mountain quarry, Stop 5, Figure 3 (planar-bedded facies); STQ= Steele quarry, Stop 1A, Figure 4 (Hummocky cross-stratified facies).

and Stop 1C of this field trip (Fig. 11). Asymmetrical ripples found on flanks of unidirectional trough cross-beds oppose the dominant northerly paleoflow, which indicates a subordinate southerly tidal current. This suggests an ebb-tidal delta deposit rather than a flood-tidal delta deposit within the tidal inlets (Reinson, 1984).

These structures lend credence to Slingerland's (1986) computer model, which proposed mesotidal to low macrotidal paleotide conditions in the Catskill sea during the Upper Devonian. Similar herringbone structures have been documented in the Becraft limestone of the Lower Devonian Helderberg Group (Ebert, 1987).

Figure 13 shows that paleoflow was dominant toward the north, with minor variations toward the northeast and northwest. Tidal inlets are therefore thought to have migrated in a northerly direction. This northerly paleoflow records either the dominant storm track, the prevailing wind direction and longshore current, or a flood tidal flow directed into headward portions of the Catskill sea.

Planar-laminated interbeds are interpreted as swash bars within tidal inlets, possibly representing spit development. Linear bars and swash bars exhibiting upper stage plane beds (areas of intense wave and current interaction) are common along the channel-margins of inlets (Reinson, 1984). Interbeds of subhorizontal planar-laminations showing opposing directions of inclination are strikingly similar to the seaward slope facies described by Davidson-Arnott, et al.,1976). These interbeds may represent complex interbedding of landward and seaward dipping, subhorizontal planar-lamination of an inlet bar. They are interpreted as fluctuations in the relative strength of waves and seaward flowing currents, forming complex interbeds where these currents interact.

Description of the planar-bedded facies

The planar-bedded facies is present at the Sidney Mountain quarry (Stop 5) in Sidney Center bluestone quarries 1, 2, 4, and 5 (Fig. 4), and at both the Skytop Lane and Sheetz bluestone quarries in Sidney Center. Rocks of this facies are also assignable to the Cattaraugus Magnafacies.

The planar-bedded, fine-grained sublitharenites of the Sidney Mountain and Sidney Center bluestone quarries are moderately well to well sorted. Planar beds are horizontal or subhorizontal. Subhorizontal planar laminations are gently inclined at angles less than 5 degrees. Bedding plane partings average 2 to 6 centimeters in thickness. Internal lamination is subtle and detectable only in thin section. Upper stage planar laminations are locally abundant. Structures such as parting lineation and aligned plant fragments show a dominantly NW-SE paleocurrent direction at all localities (Fig. 14). Inversely graded laminae (Fig. 15) are rare.

Bedding planes are commonly strewn with carbonized and pyritized plant fragments, small branches, bark, and rare logs. Discrete conglomerate beds consisting of flat shale intraclasts occur on some bedding surfaces as well (Fig. 16).

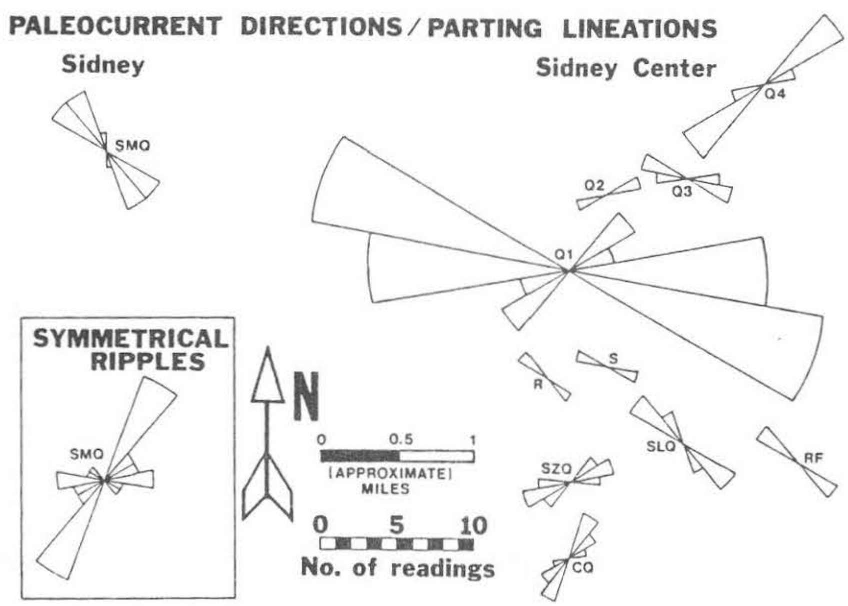


Figure 14: Rose diagrams depicting the geographic distribution of paleocurrent directions obtained from parting lineations. The inset at the bottom left corner is a rose diagram of paleocurrent directions obtained from symmetrical ripples at Sidney Mountain quarry. The abbreviations represent outcrops in their approximate geographic location (Figs. 3 and 4). CQ= Cardi quarry, Sidney Center (Planar-laminated and trough cross-bedded redbed facies; Q1= Quarry #1, Figure 4 (Trough cross-bedded facies); Q2= Quarry #2, Figure 4 (Planar-bedded facies); Q3= Quarry #3, Figure 4 (Mottled mudrock and sandstone facies); Q4= Quarry #4, Figure 4 (Planar-bedded facies); R= Outcrop R, Figure 4 (Trough cross-bedded facies); RF= Marvin Roof Barn, Roof Road, Sidney Center (Trough cross-bedded facies); S= Outcrop S, Figure 4 (Trough cross-bedded facies); SLQ= Skytop Lane quarry, Sidney Center (Planar-bedded facies); SMQ= Sidney Mountain quarry, Sidney (Planar-bedded facies); SZQ= Sheetz quarry, Cummings Rd., Sidney Center (Planar-bedded facies).

Figure 15: Subhorizontal and horizontal planar laminations are the predominant structure in the planar-bedded facies. Individual laminae are inversely graded as shown by alternating dark (very fine sand) and light (fine sand) laminae. Laminations are truncated by a reactivation surface at the top right of the photograph. Photograph of polished slab. Scale in centimeters (Stop 5).

Figure 16: Bedding plane strewn with flat shale intraclasts and the spirifer Platyrachella within the planar-bedded facies at Sidney Mountain quarry. Dime for scale (Stop 5).

Figure 17: Close-up of two molds of Barroisella campbelli. Specimen was found near the contact between the planar-bedded facies and the mottled mudrock and sandstone facies in Sidney Center. Scale in centimeters (Stop 1C).

Figure 18: The planar-bedded facies (lighter horizons at bottom and middle of photo) and the mottled mudrock and sandstone facies (darker horizons at shed and ladder and top of photo) are repeated twice at Sidney Mountain quarry. An unconformable erosional surface is located just above the top of the ladder. The second occurrence of the mottled mudrock and sandstone facies is present as a broad channel incised into the planar-bedded facies, which is filled with a fining-upward sequence of fine sand sublitharenite displaying low angle lateral accretion bedding grading to red-green mottled mudstone (Stop 5).

Figure 19: Examples of flaser and wavy bedding from near the top of the lower occurrence of the planar-bedded facies at Sidney Mountain quarry (See Fig. 18). Scale in centimeters (Stop 5).

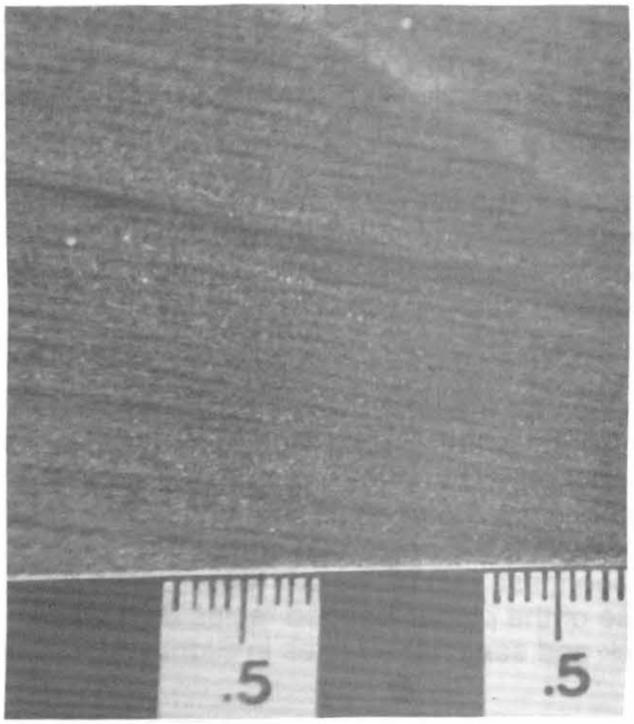


FIGURE 15

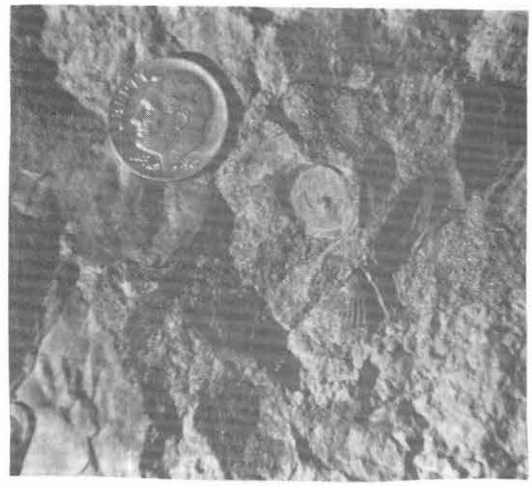


FIGURE 16

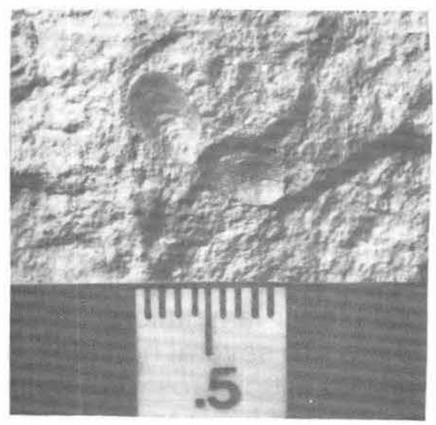


FIGURE 17

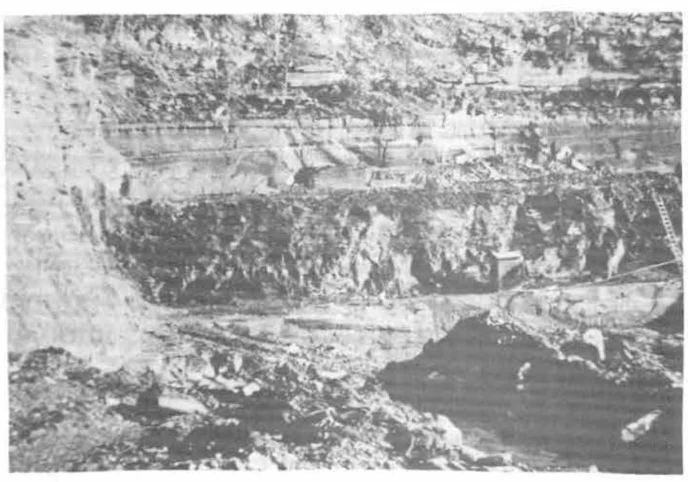


FIGURE 18

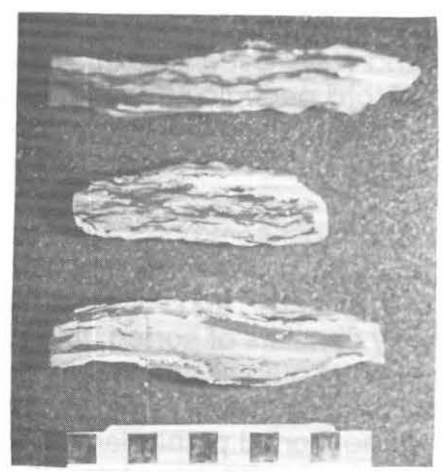


FIGURE 19

At the Sidney Center bluestone quarries (Stop 1), the uppermost half meter of the planar-bedded unit contains vertical, hematitic burrows of Skolithos and numerous shells of the inarticulate brachiopod Barroisella campbelli (?) (Fig. 17) associated with delicate plant leaves and stems.

Large, singular endichnial burrows occur in the float and are rarely found in the section. Similar burrows have been interpreted by Bridge, Gordon, and Titus (1985) as upward escape burrows of the nonmarine bivalve Archanodon occurring in fluvial and deltaic deposits. Gordon (personal communication, 1989) now thinks that they may be an indicator of brackish conditions.

The contact of the planar-bedded facies with the underlying trough cross-bedded facies is gradational in the Sidney Center area. Subhorizontal and horizontal planar laminations increase in abundance upward within the trough cross-bedded facies. In addition, subhorizontal planar laminations increase in abundance are much more common than horizontal planar laminations in the lower portions of the planar bedded facies. Contact relationships between these two facies are not exposed at Sidney Mountain quarry. However, the base of the planar-bedded facies is in sharp (erosional) contact with the mottled mudrock and sandstone facies in Sidney Mountain quarry (Stop 5).

At the Sidney Mountain quarry, the planar-bedded facies consists of two individual, fining-upward sequences with gently scoured and loaded bases. Each unit is interbedded with the mottled mudrock and sandstone facies (Fig. 18).

The first fining-upward unit consists of fine sand sublitharenite. Subhorizontal and horizontal planar laminations are the dominant sedimentary structures. Within the strata, there are several horizons of flat shale intraclasts, which rarely contain the rhynchenellid, Cupularostrum, the spiriferid, Platyrachella, and fish fragments (Fig. 16). Inversely graded laminae and numerous, highly abraded, arthropod fragments are common in this interval (Fig.15). Flaser bedding occurs commonly in this interval, which is characterized by discontinuous shale lenses which drape crests and troughs of ripples (Fig. 19). Wavy bedding with continuous shale lenses intervening between ripple cross-lamination occurs with flaser beds. Some of the shale surfaces of the flaser and wavy bedding exhibit sand-filled mudcracks with crude polygonal patterns, which record periodic subaerial exposure of these surfaces. A discontinuous linguoid rippled surface caps the fining-upward unit. Bioturbation levels are low in this interval. Burrow types are restricted to epichnial sinuous trails.

The most common sedimentary structure in the upper occurrence of this facies is large scale, subhorizontal planar laminations. This unit comprises the largest sand body in the quarry, and will be referred to as the main sandbody (fig. 18). The main sandbody consists of sorted, fine sand sublitharenite to quartz arenite. The subhorizontal planar laminations are inclined at angles of less than 5 degrees. These inclined beds alternate vertically and are truncated laterally by more mature (quartz arenite), horizontal planar lamination containing inversely graded laminae (Fig. 15). The cross beds also pass laterally into rare small channel forms with internal trough cross-lamination. The base of this sand body interval is gently loaded and scoured with occasional higher angle cross bedding. A variety of tool marks, flute casts, and load casts occur on the sole of the main sandbody. The base also contains numerous types of hypichnial burrows along with the rhynchenellid Cupularostrum and broken, unidentifiable brachiopod shells. Some of the hypichnial burrows include abundant vertical tubes, Cruziana arthropod trails, crescent-shaped burrows in convex hypo-relief Rhizocorallium (?), and "figure-8" burrows in convex hypo-relief.

Near the top of the main sandbody is a large, shallow scour 2 meters deep and approximately 50 meters wide. Sedimentary structures above this scour include numerous imbricated ball and pillow structures, climbing asymmetrical ripples, delicate shale interbeds (wavy bedding?), and lingoid ripples.

Numerous structures are found on the top bedding surface of the main sandbody. Lingoid ripples are common. These lingoid ripples are sometimes overlain by a secondary set of straight crested, asymmetrical ripples which have rounded to flat, planed-off crests with pointed and grooved troughs (Fig. 20). Other structures present at this horizon include symmetrical ripples and "ladder-back" ripples (Fig. 21). The ladder-back ripples consist of one set of symmetrical, oscillatory ripples and a second set of asymmetrical ripples within troughs formed at nearly right angles to the crest of the oscillatory ripples. In addition, a large block of silty sandstone was found in the float that contains large polygonal mudcracks which indicates periodic subaerial exposure of this surface.

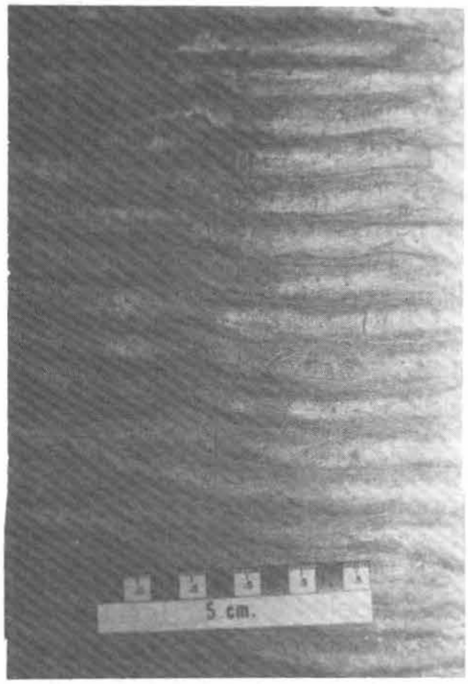
The sequence of facies repeats throughout Sidney Mountain quarry. Also, two unconformable contacts exist: 1) between the mottled mudrock and sandstone facies and the planar-bedded facies in the middle of quarry (Fig. 18); and 2) between the planar-laminated and trough cross-bedded facies and the hummocky cross-stratified facies at the top of the quarry (Fig. 7).

Large scale planar cross-strata inclined at low angles at the top of the main sandbody are probably sand-filled channels (shallow and broad) exhibiting lateral accretion surfaces within deposits of the mottled mudrock and sandstone facies. Channel forms filled with lithologies of the mottled mudrock and sandstone facies are commonly incised into the top of the planar-bedded facies. The contact between these two facies is therefore sharp and erosional.

Figure 14 shows rose diagrams of paleocurrent directions measured from parting lineations at their respective geographic locations in the field.

Interpretation of the planar-bedded facies

The planar-bedded facies is interpreted as a prograding strandplain and barrier beach deposit (Fig. 22). There is subtle evidence for protective barrier islands, which are often associated with strandplain deposits. The absence of washover fans and backdunes makes it difficult to confirm a barrier setting. However, portions of the planar-bedded facies were probably consumed by lateral migration of tidal inlets of the trough cross-bedded facies, which gives barriers a low preservation potential.



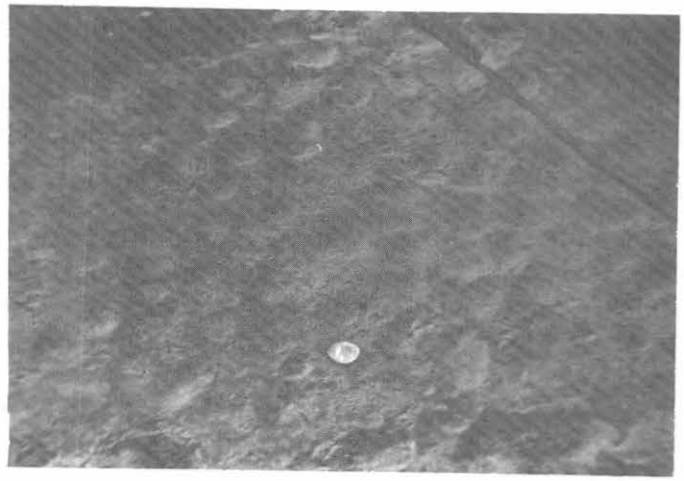


FIGURE 21

FIGURE 20

Figure 20: Flat-crested ripples. Asymmetrical ripples with rounded to flat crests and pointed troughs. These ripples are most likely the product of tide or swash reworking of symmetrical oscillatory ripples. Current direction is toward the bottom of the photograph (field azimuth= 50 degrees). Scale in centimeters (Stop 5).

Figure 21: Ladder-back ripples on the top bedding surface of the planar-bedded facies at Sidney Mountain quarry is clearly evidence of tidal influence (Stop 5).

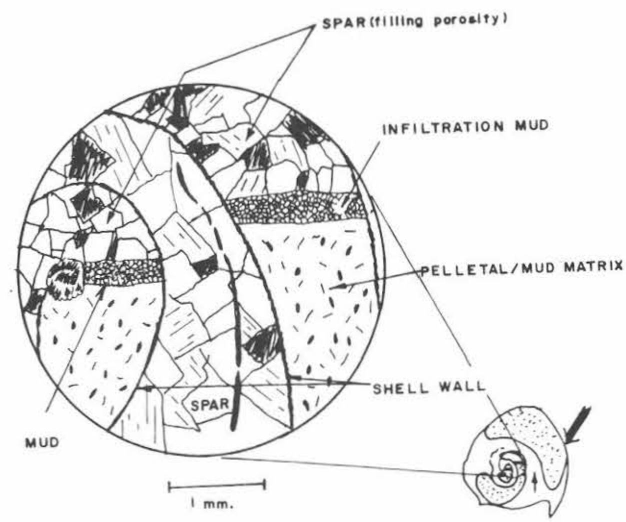


Figure 23: Sketch of photomicrograph showing infiltration mud filling shelter porosity inside the chambers of Bellerophon, float specimen from After, N. Y. Note the smaller depiction of the entire shell with the last whorl and body chamber completely filled by pelletal silty micrite. Shell has been reworked as evidenced by two layers of infiltration mud at the same orientation within smaller central chambers of the shell. Thicker arrow indicates original resting up position before reworking. Smaller arrow shows reoriented position following reworking.

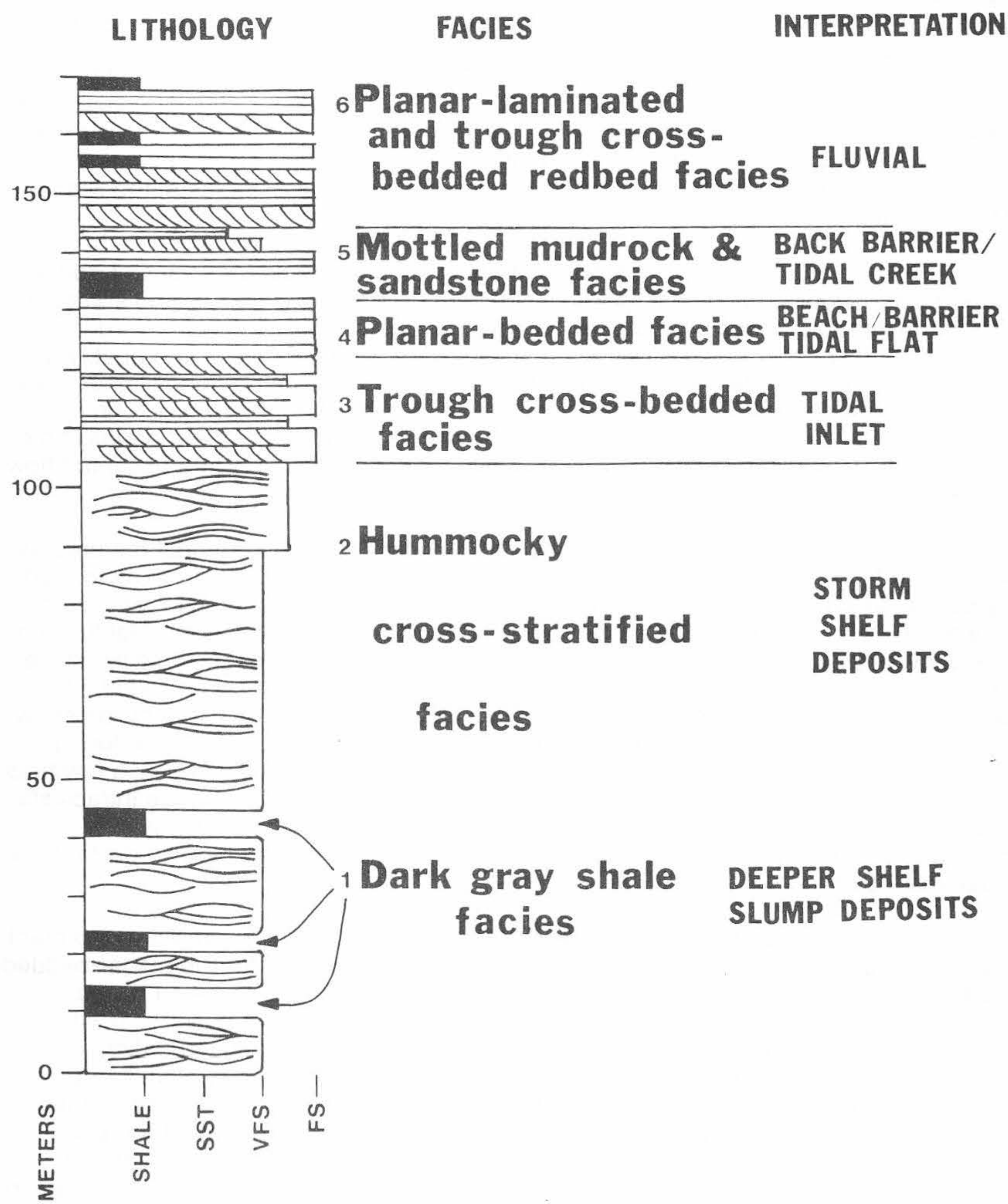


Figure 22: A 170 meter composite section illustrating the stratigraphic sequence and interpretation of the six facies units described in the study area. The overall sequence shows a coarsening-upward trend. Paleoenvironments of deposition are indicated in the right-hand column.

Close examination of the rose diagrams in Figure 14 has provided clues to reconstruct the morphology of the paleoshoreline. A dominant trend is present in a southeast to northwest direction. The paleoshoreline is oriented normal to this, in a northeast-southwest direction. This orientation is in agreement with past paleogeographic reconstructions of the Catskill sea (Haeckel and Witzke, 1984; Barrel, 1913, 1914; and Friedman, et. al., 1966). There is a slightly lesser component which parallels the shoreline orientation. This is interpreted as either an ebb directed tidal flow behind barriers or a flood-dominated flow seaward or barriers directed northeast into the Catskill basin. The orientation of symmetrical ripples at the Sidney Mountain quarry also supports this interpretation.

There is ample evidence for tidal activity in the planar-bedded facies. Ladder-back ripples suggest flood tidal flow with one set of symmetrical ripples and ebb tidal flow normal to this, which is preserved as asymmetrical ripples confined to troughs of the symmetrical ripple set. The secondary set of asymmetrical ripples is probably a late stage emergence feature formed as tidal currents drained off a depositional slope of a beach or tidal flat. Flat-crested ripples also record late stage tidal runoff on intertidal flats. Tidal currents caused planation of crests and subsequent deposition of reworked sediment into adjacent troughs. Field measurements indicate dominant flow to the north, which helps to confirm ebb flow behind barriers.

Flaser and wavy bedding (Reineck and Wunderlich, 1968) are tidal features as well, since they are found in close association with ladder-back ripples. They formed by settling of silt and clay from suspension at slack tide when current velocity is zero. The most likely location for flaser and wavy bedding to have formed is on tidal flats on the landward side of a barrier and adjacent to the back-barrier lagoon. The flat shale intraclast associated with horizons strewn with intraclasts, brachiopods, and plant debris may represent intraclasts derived from a local source, because the clasts show poorly rounded edges and clay clasts are readily weathered and destroyed during transport. Since the shale intraclasts are mostly concentrated in layers only a few tens of decimeters below flaser and wavy bedded units, it is likely that the shale intraclasts are derived from tidal flat sediment via spring tidal currents and/or strong storms. Similar suites of sedimentary structures in tidal flat settings are reported by deRaaf and Boersma (1971).

The association of Skolithos linearis, Barroisella campbelli?, and delicate plant leaves and stems strengthens a beach (foreshore) interpretation for the planar-bedded facies. Also, inverse graded bedding, highly abraded and highly rounded quartz grains, and abraded arthropod fragments are features common on beaches (Clifton, 1969). Inverse graded bedding is rare perhaps owing to infaunal bioturbation causing nearly complete sediment homogenization. An alternative explanation to account for homogenized beds is that the bluestones quarries in the Sidney Center area may represent an upper beach area landward of the swash zone where plant growth and soil development destroyed internal structure (Clifton, 1981). Another alternative is that much of the wave energy is dissipated just offshore on swash bars in the trough cross-bedded facies. This may prevent upper stage plane beds from forming intricate laminations owing to reduced swash agitation.

Thayer (1974) indicates that Cupularostrum and Platyrachella are abundant in the nearshore zone, because they are unusually euryhaline. The globose valves of Cupularostrum and the thick valves and compact shape of Platyrachella inhibited shell breakage in turbulent nearshore waters (Thayer, 1974). These brachiopods probably lived in subtidal environments of Sonyea tidal flats and/or just offshore, and were washed up onshore. Thayer (1974) also notes substantial numbers of bivalves, including Barroisella, in green silts and shales of estuarine, lagoonal, and tidal flat origin of the distal portions of the Oneonta Formation and within proximal portions of other marine formations of the Genesee Group. Barroisella was an infaunal burrower with a long, siphon-like pedicle and lived with its long axis vertical (Boucot, 1981). It probably lived in intertidal creeks and flats in the Sonyea Group.

In summary, the sedimentologic and paleontologic data of the planar-bedded facies best fits a depositional model of a prograding barrier island/strandplain and tidal creek system. Tidal features documented in this study are suggestive of a mesotidal range. These barrier islands probably were drumstick-shaped, which is the characteristic shape of barrier islands along mesotidal shorelines. The progradational tidal creek/drumstick barrier island system of Ossabaw Sound, Georgia (Greer, 1975) is suggested as a modern analog.

Some ancient analogs of regressive barrier island/strandplain and estuarine systems that are reminiscent of this investigation include: the Lower Cretaceous Notikewin Member (Fort St. John Group), northeastern British Columbia (Leckie, 1985); and the Upper Cretaceous Cardium Formation of the Kakwa Field and adjacent areas, Northwestern Alberta (Plint and Walker, 1987).

Heward (1981) contends that longshore and shelf sediment sources provide insufficient quantities of sediment to be maintained for extensive progradation of barrier island systems, despite a constant sea level over a long span of time. However, Leckie (1985) finds that the barrier island system of the Notikewin Member of the Fort St. John Group had prograded approximately 150 kilometers. A delicate balance between sediment supply and rate of subsidence would have been required for the back barrier lagoon to have migrated behind the barrier island complex and not have been filled in (Leckie, p.49, 1985). Also, a large sediment supply by rivers and a moderate wave energy are essential (Carter, 1978). Environmental constraints and processes similar to those described by these authors are thought to have operated during the Upper Devonian in the study area.

Tidal range apparently increased in the Catskill sea progressively toward the northeast. In central Pennsylvania, prograding muddy shorelines of weaker tidal influence (microtidal) have been described (Walker and Harms, 1975). In northern-most Pennsylvania, more abundant evidence for tides has been presented, which probably reflects increasing tidal range (Bridge and Droser, 1985). Sedimentary structures indicate significant tidal influence (mesotidal) within the study area. This trend supports the existence of a narrow, elongated embayment present at the north end of the Catskill sea (Heckel, et al., 1979; see Fig. 1). Tidal augmentation may have continued toward the headward portions of the Catskill sea, similar to modern tidal processes of the Bay of Fundy, Nova Scotia, Canada.

Description of the mottled mudrock and sandstone facies

The mottled mudrock and sandstone facies is characterized by red-green mottled mudstone, and less commonly, green-gray mottled shale. It is also assignable to the Cattaragus Magnafacies of the nonmarine to marine transition zone. This facies can be found at Sidney Center and Sidney Mountain quarry.

Exposure of this facies at Sidney Center consists of 1 to 3 meter fining-upward sequences of fine sublitharenite, sandy and silty mudstone, and mudstone (Fig. 18). Contact relations with the underlying planar-bedded facies is sharp and planar. Sedimentary structures within sandstone interbeds include alternating trough cross-bedding, planar-bedding, and rare asymmetrical ripples. Trough cross-beds are commonly inclined at low angles (5 to 17 degrees).

At Sidney Center, the lower half of the mudstone interval is green and featureless. Sparse, diminutive vertical burrows are present, which are possibly either Skolithos-like burrows or rootlets. The occurrence of this facies marks the first appearance of green and red coloration throughout the Sonyea stratigraphic section. This is significant, because sediments of the fluvial Catskill Magnafacies are dominantly green and red whereas open marine siltstone and shale are characteristically gray in color.

A distinct horizon of red-green mottling overlies the green featureless mudstone interval. The red-green mottled horizon is, overlain in turn, by dominantly red- to green-mottled, sandy- to silty-mudstone. This interval contains abundant fish plates, bone fragments and teeth of Bothriolepis and possibly Eusthenopteron(?). Also present are hematitic Skolithos burrows, and rare Barroisella campbelli(?).

At the Sidney Mountain quarry, the mottled mudrock and sandstone facies contains interbeds of siltstone and very fine sand sublitharenite. The mottled mudrock and sandstone facies is repeated twice at Sidney Mountain quarry (Fig. 18). It is interbedded with the planar-bedded facies, which also occurs twice within the quarry.

The first occurrence of this facies is bounded on top and bottom by sharp and loaded contacts. The siltstone and shale are planar-laminated or blocky, and dominantly green, gray or mottled green-gray. The sandstone is generally tan in color and exhibits planar lamination with rare parting lineations. Hummocky cross stratification is rare. Sandstone and siltstone interbeds are capped sporadically by symmetrical wave ripples and ripple cross lamination. Ball and pillow structures are present at and near the base of the interval. Fossils are restricted to carbonized and chalcopyritized plant fragments. Bioturbation is moderate to low.

The second occurrence of the mottled mudrock and sandstone facies at Sidney Mountain quarry, consists of one fining-upward sequence of fine sublitharenite and moderate to heavily bioturbated, mottled, red siltstone and mudstone. Channels commonly incise approximately one meter into the planar-bedded facies. Some of these channels are filled with sorted, fine sand sublitharenite exhibiting gently inclined planar-lamination interpreted as lateral accretion surfaces. Other channels are filled with moderate to poorly sorted, very fine sandstone, siltstone and shale. Siltstone and shale are characteristically red to red-green mottled.

Sedimentary structures include alternating planar-beds and trough cross-beds. Ladder-back ripples, symmetrical ripples, asymmetrical ripples and ripple cross-laminations are otherwise rare. Mudcracks occur throughout the interval but are rare. Asymmetrical ripples with superimposed mudcracks and ladder-back ripples occur rarely within siltstone lenses. Other dessicated surfaces are covered with various non-descript burrows with a red-green mottling. Wrinkle marks ("runzel marks") are found in close proximity to mudcracks. The bases of the siltstone lenses commonly have flute casts and structures that resemble small gutter casts.

Both the siltstone and mudstone contain abundant mud-filled root casts with macerated plant fragments. The casts are interpreted as roots rather than burrows, because the caliber of the opening shows a slight tapering on some specimens.

Interpretation of the mottled mudrock and sandstone facies

The mottled mudrock and sandstone facies probably represents a tidal creek and back barrier lagoon sequence, which partially dissects through a barrier island/strandplain system. Modern analogs to this sequence are described by Greer (1975) of the estuarine-marine transition of the Ogeechee river, Ossabaw Sound, Georgia. Ancient counterparts are suggested from work done by McCrory and Walker (1986).

The association of fish debris with inarticulate brachiopods and Skolithos, occurring directly above foreshore beach sediment of the planar-bedded facies, argues in favor of brackish water conditions. The mottled mudrock and sandstone facies may represent small, shallow, sluggish tidal creeks. Similar brackish water deposits, containing similar faunal associations, have been reported in the Devonian Catskill Magnafacies by Gordon and Knox (1989) and Thomson (1976). Gordon (1990, personal communication) is re-examining several genera of fish for possible habitation in brackish settings. It is suggested that the abundance of fish in the mottled mudrock and sandstone facies may suggest proximity to marine conditions and may reflect brackish conditions.

The fish debris at Sidney Center could be concentrated by one of the following mechanisms: a) by strong storms and/or floods, thereby producing a tempestite; or b) a channel floor lag deposit; or c) by processes related to tubidity maxima within estuaries, which causes concentrations of debris to be deposited at the nodal zone, where fresh and salt water prisms meet. Lateral migration of the nodal zone is known to accumulate concentrations of materials in modern estuaries and tidal creeks such as the upper reaches of Chesapeake Bay (Schubel, 1971a) and in the Demerara estuary of the coast of British Guiana (Schubel, 1971b). Similar processes could have operated in tidal creeks along the Catskill sea shoreline.

Several horizons at Sidney Mountain quarry suggest brackish to marine conditions. Ladderback ripples and asymmetrical ripples with superimposed mudcracks are found in close proximity along the same horizon. Ladderback ripples are clearly of tidal origin. Therefore, superimposed mudcracks on some ripples surfaces record intermittent emergences of the depositing surface during falling tides and/or discharge fluctuations. Wrinkle marks ("runzel marks") are also found in close proximity to these structures. The wrinkle marks can be interpreted here as another indicator of subaerial exposure when found in association with other dessication features. Reineck (1969) showed experimentally that such wrinkle marks develop when a strong wind blows over a partially cohesive sediment surface covered by a thin film of water, which deforms the sediment into wrinkles. These structures collectively record intertidal flat deposition along tidal creek margins.

The thickness of a fining-upward sequence approximates the depth of the tidal creek assuming that scouring into the mudstone by the planar-laminated and trough cross-bedded redbed facies was minimal. The tidal creeks in Sidney Center were probably not major streams but rather small and shallow, because the mudstone unit is only 3.36 meters thick. Some of the mudstone was stripped, because sandstones at the base of fining-upward sequences are commonly loaded and have scour and fill bases. However, at the Sidney Mountain quarry, the mottled mudrock and sandstone facies is 6.65 meters thick, which probably represents a larger tidal creek. The second occurrence of this facies at Sidney Center represents a smaller tidal creek, because it is only 2.2 meters thick.

Both the planar-bedded facies and the mottled mudrock and sandstone facies are repeated twice at Sidney Mountain quarry (Fig. 18). Initially, barrier island and barrier beach deposits of the lower interval of the planar-bedded facies prograded seaward. Tidal creek and back-barrier lagoon deposits prograded seaward as barriers/beaches prograded. A localized, brief marine trangression occurred, which produced a loaded, sharp erosional base followed by continued deposition of the planar-bedded facies. Soft sediment deformation structures at this locality are not of sufficient size, frequency, or penetration to account for the repetition of facies by subsidence alone. A combination of these processes probably occurred. Marine transgression may have been prompted by an episode of epeirogenic downflexure (Quinlan, et al., 1984). Again, tidal creeks and distal fluvial deposits prograded seaward as barriers/beaches prograded. A second marine transgression of greater magnitude is implied by rocks of the hummocky cross-stratified facies deposited unconformably on fluvial deposits of the planar-laminated and trough cross-bedded redbed facies located at the uppermost 4 meters of Sidney Mountain Quarry.

Red-green mottling in the mottled mudrock and sandstone facies is the product of one of the following mechanisms: a) biogenic activity by infauna or roots; b) salinity fluctuations in a brackish setting; or c) groundwater fluctuations in a terrestrial or coastal setting.

The mottled mudrock and sandstone facies is sharply in contact with, but lithologically similar to the planar-laminated and trough cross-bedded redbed facies. Placement of the contact between these two facies was difficult. It involved painstaking observations to locate evidence for tidal influence in the mottled mudrock and sandstone facies. This poses some problems. Tidal creek and fluvial sedimentary processes, lithologies, and resultant fining-upward sequenced are strikingly similar in many respects. Some intervals in this study remain problematic. It is not clear whether some intervals in the study area are estuarine or fluvial in origin. Within each of the problematic intervals, there is a lack of tidal features. However, root casts, vertical burrows, fish fragments, and plant remains are much more abundant in these intervals than in fluvial deposits of the planar-laminated and trough cross-bedded redbed facies higher in the section. Evidence for tidal influence may have been overlooked here.

Description and interpretation of the planar-laminated and trough cross-bedded redbed facies

The rocks of the planar-laminated and trough cross-bedded redbed facies represent nonmarine, fluvial Catskill Magnafacies of the Lower Walton Formation. These rocks bear close resemblance to Bridge and Gordon's (1985) study of fluvial fining-upward sequences of the Oneonta Formation of the Genesee Group, New York. Bridge and Gordon (1985) conclude that the Oneonta Formation was deposited by single, sinuous streams and rivers flowing over a lowland alluvial plain. The planar-laminated and trough cross-bedded redbed facies is thought to represent a similar environment of deposition. Other noteworthy investigations of fluvial Catskill Magnafacies include Bridge (1988) and Sevon (1985).

SUMMARY OF CONCLUSIONS

1. Figure 22 is a 170 meter composite section illustrating the stratigraphic sequence and interpretation of each of the six facies described in the study area.

2. The sequence of facies and the distribution of sedimentary structures suggest non-deltaic progradation of the Catskill clastic wedge during Sonyea (Frasnian) time.

3. Hummocky cross-stratification is a storm-generated structure formed under high velocity, oscillatory flow conditions. It is found between fair-weather and storm wave base. The abundance of hummocky cross-stratification along with the lack of small scale coarsening-upward sequences throughout the entire section warrants against an interpretation of deltaic deposits in the nearshore. Frequent storms would inhibit outbuilding of deltaic lobes and argues for a straighter shoreline.

4. Vertical stacking of hummocky cross-stratified beds produced nearshore shallowing. This provided conditions conducive for progradation of coastal deposits over the hummocky cross-stratified facies.

5. The dark gray shale facies represents deeper portions of the shelf. Anoxic conditions were produced by either pycnoclinal fluctuations or sea level rise or both. Pycnoclinal adjustment and/or sea level rise may be the result of epeirogenic lithospheric downflexure (Quinlan, et al., 1984) or eustacy. Listric truncation surfaces represent submarine slope-failure surfaces of soft sediment slumps.

6. The trough cross-bedded facies represents migrating tidal inlets of a barrier island or estuarine complex. Dunes show migration in a north to northeasterly direction, which probably indicates inlet migration in that direction. Northerly paleoflow records either dominant direction of longshore transport, tidal current direction, or storm track path. Planar-laminated interbeds may represent either swash bars or spit complexes within tidal inlets.

7. The planar-bedded facies represents either prograding strandplains or barrier beaches with associated intertidal environments (i.e., tidal flats, back-barrier salt marshes) in an overall barrier island complex. Paleocurrent data suggest a shoreline trend of southwest to northeast.

8. The mottled mudrock and sandstone facies represents estuarine and tidal creek deposits that partially dissect into the planar-bedded facies. Estuaries prograded seaward as beaches prograded.

9. Fluvial fining-upward sequences of the planar-laminated and trough cross-bedded redbed facies erosionally overlie coastal deposits of the mottled mudrock and sandstone facies. This facies is interpreted as fluvial Catskill Magnafacies of the Lower Walton Formation. Streams and rivers deposited sediment in a vast alluvial plain, which prograded seaward as transitional facies prograded.

10. Some sedimentary structures (i.e., flaser/wavy bedding, ladder-back and flat-crested ripples, and reformed ripples with herringbone cross-lamination) strongly suggest that tides were active in the Catskill sea.

11. Paleocurrent data and tidally-produced sedimentary structures suggest tidal augmentation in the Catskill sea progressively toward the northeast. This tends to support the existence of a narrow, elongate embayment present at the north end of the Catskill Sea.

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ROAD LOG FOR PROXIMAL STORM-DOMINATED SHELF TO TIDALLY-INFLUENCED FORESHORE SEDIMENTATION, UPPER DEVONIAN SONYEA GROUP, BAINBRIDGE TO SIDNEY CENTER, NEW YORK

CO-LEADERS: BISHUK, D., APPLEBAUM, R., AND EBERT, J.

Cumulative Mileage	Miles From Last Point	Route Description
0.0	0.0	Mileage begins at intersection of Main Street in Oneonta and Exit 14 onramp to Interstate 88. Take westbound onramp onto I-88.
7.3 and 10.6	7.3 and 10.6	To your left are exposures of fluvial deposits of the Oneonta Formation (Genesee Group). This formation is similar in many aspects to the Lower Walton Fm. (Sonyea Group) included in this study.
11.6-12.9	4.3	The Susquehanna river has incised through the Wells Bridge moraine (right). This moraine served as a dam, which impounded glacial Lake Otego. Upgradient of the moraine, 300 to 400 feet of lacustrine silts and clays underlie the valley floor. Downgradient, an outwash terrace can be seen at the rest area (13 mi.).
15.4-16.2	2.5	Immediately after exit 11, a sinuous, beaded esker can be seen at your right (extends for approx. 0.5 miles).
17.3	1.1	Take exit 10 off I-88 at Unadilla. Outcrops of Oneonta Fm. at left.
17.9	0.6	As you descend off of long off-ramp, make your first left onto River Road (located just before bridge over Susquehanna River). Glaciofluvial sand & gravel quarry to the left.
18.9	1.0	Make your 2nd (hard) left onto Delaware County Route 23 (east).

19.3	0.4	Reference marker: Quarry Rd.; Helderberg Bluestone Quarry at end of road; Excellent exposure of the Oneonta Fm. (Genesee Group). From Quarry Rd., count to the second road on your right (Dunshee Rd.).
20.5	1.2	Passing Road #13.
21.7	1.2	Turn right onto Dunshee Road.
22.2	0.5	Hummocky cross-stratified sandstones interbedded with shale of the Glen Aubrey Fm. (Sonyea Gr.).
22.5	0.3	Glen Aubrey Fm.(right); ball and pillow structures predominate the exposure.
23.2	0.7	Small quarry to the right consists of the amalgamated portion of the hummocky cross-stratified facies of this study (Glen Aubrey Fm.).
23.3	0.1	At intersection of Dunshee Rd. and Delaware County Route 35, bear left. Proceed less than a tenth of a mile past trailer home and make first left onto cobble driveway into an open field. Driveway leads to several bluestone quarries.
23.45	0.15	Park cars just before powerline.

STOP 1: SIDNEY CENTER OUTCROPS (See Figures 3 and 4)

STOP 1A: STEELE QUARRY AND PASTURE OUTCROPS OF THE HUMMOCKY CROSS-STRATIFIED FACIES (See Figure 4)

Location: Scattered outcrops and a small, inactive quarry are located in a cow pasture behind the Steele residence. All stop 1 outcrops can be found in the south-central portion of the Unadilla, New York 7.5 minute quadrangle, located at the intersection of Dunshee Road and Delaware County Route 35. Before visiting the Steele quarry, please ask Mr. and Mrs. Steele for permission to walk on their property. Since stop 1A is on private property, the authors ask that you do not use rock hammers on the outcrops or collect any samples.

Description: The Steele quarry and pasture outcrops (outcrop V) (Fig. 4) are stratigraphically positioned in the Glen Aubrey Formation, and are designated as the amalgamated portion of the hummocky cross-stratified facies. These exposures offer an excellent opportunity to examine hummocks and swales of HCS beds in plan view. The amalgamated portion of the hummocky cross-stratified facies is interpreted as dominantly sandy, storm deposits at or above fair weather wave base (consult text for details). Fossils are sparse and are restricted to coquinite lenses consisting of Sphenotus, Cupularostrum, Platyrachella, fish plates, and plant fragments. Sedimentary structures which cap hummocky cross-stratified beds include wrinkle marks crescentic scour depressions partially filled with very fine sand. These scours commonly exibit asymmetrical and symmetrical ripples within the depression.

An enigmatic shale unit is found in the center of the quarry. At approximately halfway up in the shale, there is a light colored and differentially weathered horizon which has rather high concentrations of dolomite. Since this shale can not be traced laterally, it is difficult to decipher its stratigraphic context and facies relationships.

At outcrop V in the cow pasture (Fig. 4), excellent hummocky bedforms are present. Within the hummocky cross-stratified beds, climbing ripple cross-lamination are inclined at a steep angle of propagation (Fig. 9). The ripples are slightly asymmetric in the direction of propagation toward the crest of the hummock (from left to right) suggesting that they migrated slightly up the flank. Steep inclination of propagation and the form-concordant style reflects vertical growth of the hummocks with minimal migration. This supports that hummocky cross-stratification formed under high velocity oscillatory flow conditions and rapid deposition rates during storm events.

STOP 1B: OPEN PIT AND FIELD OUTCROPS OF THE TROUGH CROSS-BEDDED FACIES (See Figure 4, Outcrops R and S)

Outcrops R and S represent the lowest portion of the trough cross-bedded facies, which sharply overlies the hummocky cross-stratified facies. The trough cross-bedded facies is assignable to the lowest facies unit of the Cattaraugus Magnafacies. Locally, an erosional base defines its base and near base, which is characterized by a coarse lithoclast, shelly, and plant debris lag. Contact relationships are not observable at this stop. The outcrops are positioned within a meter or two above the contact. However, a coarse lithoclast, shelly lag is present at outcrop R. The rhynchenellid, Cupularostrum, and the bivalve, Sphenotus, occur sparsely at this horizon. The sparse marine fauna is the best way to differentiate between the trough cross-bedded facies and the strikingly similar planar-laminated and trough cross-bedded facies (Lower Walton Formation) positioned stratigraphically higher. Trough cross-beds commonly climb at low angles of propagation and interpreted as dunes within tidal inlets within a barrier island/strandplain and tidal creek system. Planar laminated interbeds are interpreted as swash bars within the tidal inlets.

STOP 1C: LOGGING TRAIL OUTCROP OF THE TROUGH CROSS-BEDDED FACIES (Figure 4, Outcrop D-b)

Outcrop D is located approximately 40 feet higher within the trough cross-bedded facies than stop 1B. Trough cross-beds are spectacularly illustrated here owing to well developed differential weathering of the outcrop surface. Trough cross-beds exhibit normally graded, lenticular laminae, which probably reflect sediment supply pulsations induced by tides (Fig. 11). In addition, asymmetrical ripples on troughs commonly oppose the dominant, northerly paleoflow direction of trough cross-beds (Fig. 11). These asymmetrical ripples record a subordinate current reversal induced by tides within tidal inlets. Evidence of tidal activity in the Catskill Sea is only sparsely documented in the literature (Ebert, 1987; Walker, 1975; and Bridge and Droser, 1985) and has never been documented from the Sonyea Group.

STOP 1D: WOODLAND OUTCROP OF THE TROUGH CROSS-BEDDED FACIES (See Figure 4, Outcrop K)

Outcrop K is also assignable to the trough cross-bedded facies. Trough cross-beds are likewise spectacularly illustrated at this outcrop owing to well developed differential weathering of the outcrop surface. Trough cross-beds represent dunes migrating within tidal inlets. These alternate with planar-laminated interbeds, which represent swash bars deposited along the margins of tidal inlets. The trough cross-beds are unidirectional to the north and climb at low angles of propagation. Rare ripple cross-lamination on weathered surfaces is suggestive of current reversals, which offers additional supporting evidence for tidal influence.

STOP 1E: BLUESTONE QUARRIES #4 & #5 OF THE PLANAR-BEDDED FACIES AND THE MOTTLED MUDROCK AND SANDSTONE FACIES

The planar-bedded facies is sharply overlain at these quarries by the mottled mudrock and sandstone facies. The contact is planar and lacks incision by channels. The planar-bedded facies is extensively quarried for "bluestone" slabs used for sidewalk and construction purposes. Internal lamination within planar-beds is rare, but where it is present, it is often inversely graded. These structures are common on high energy foreshore beach/barrier setting. Also present are abraded arthropod fragments, and highly abraded and highly rounded quartz grains. The arthropod fragments may be either ostracods or trilobites. Near the upper contact of the planar-bedded facies, an association of Skolithos linearis, Barroisella campbelli, and delicate plant leaves and stems strenthens a beach (foreshore) interpretation for the planar-bedded facies. Barroisella was an infaunal burrower with a long, siphon-like peducle and lived with its long axis vertical (Boucot, 1981). It probably lived in intertidal creeks and tidal flats in the Sonyea Group.

The mottled mudrock and sandstone facies stratigraphically above has an association of abundant fish debris of Bothriolepis with Barroisella and Skolithos, occurring stratigraphically above foreshore beach deposits of the planar-bedded facies, argues in favor of brackish water conditions. The mottled mudrock and sandstone facies may represent tidal creek deposits.

STOP 1F (OPTIONAL): BLUESTONE QUARRY #1/ TRANSITION OF TROUGH CROSS-BEDDED FACIES WITH THE OVERLYING PLANAR-BEDDED FACIES (See Figure 4)

Bluestone quarry #1 occurs near the gradational contact between the trough cross-bedded facies and the planar-bedded facies. Planar-bedding dominates while trough cross-bedding is rare. The purpose for visiting this outcrop is to establish the contact relationship and to have an opportunity to collect plant fossils. Fossils of carbonized and pyritized plant stems, branches, logs, and bark can be readily collected at this location. Parting lineations and aligned plant fragments are abundant, which represents a strong onshore/offshore subtidal flow pattern.

23.45	0.0	Return to vehicles. Proceed back out to Dunshee Road.
23.6	0.15	Turn right onto Dunshee Road
25.1	1.5	Turn left onto Delaware county Route 23.
27.9	2.8	Turn right onto River Road.
28.9	1.0	Turn right onto I-88 onramp. Take I-88 westbound.

OPTIONAL FOR LUNCH

33.8	4.9	Take exit 9 from I-88 at Sidney.
34.1	0.3	Turn right onto Route 8 North.
34.6	0.5	Turn left at the first traffic light.
34.9	0.3	Turn left into McDonalds parking lot.

BREAK FOR LUNCH.

35.2	0.3	Return to vehicles. Proceed back out to Route 8. Turn right onto Route 8 south.
35.6	0.4	Turn right onto I-88 west

36.2-37.0	0.6	Exposures of the Triangle Fm. (Sonyea Group) consist of hummocky cross-stratified sandstone interbedded with shale.
38.8	1.8	Take exit 8 off I-88 at Bainbridge. Park cars on off-ramp at end of large outcrop.

STOP 2: HUMMOCKY CROSS-STRATIFIED FACIES, TRIANGLE FORMATION (?), EXIT 8 ONRAMP ON I-88, BAINBRIDGE (Figure 3)

Three outcrops afford ample opportunity to examine storm deposits. We will concentrate most on the largest exposure directly south of the eastbound lanes of Interstate 88. Storm beds, numerous soft-sediment deformation structures of variable scale, bioclastic mudstones and sandstones, and listric truncation surfaces are most easily identifiable at the exposure.

Please be extremely careful not to go out on the highway as this spot has limited sight distance for drivers. Also, vehicles are accelerating up the onramp onto the primary lanes. Please be alert at all times while at this stop. You may cross over to examine the middle outcrop north of the eastbound lanes. Be absolutely sure there are no oncoming cars as far as you can see down the lanes and around the bend before you cross the road. Vehicles are moving faster than they appear but if you are careful then tere is plenty of time to cross. Bedding planes are exposed at the very top of the outcrop displaying wave ripples and oriented fossils. They can be reached by carefully ascending the east end of the outcrop at the large culvert that passes under the road. Note also the numerous bedding surfaces with wave (?) or current(?) oriented Mucrospirifers and Tentaculites. East of the culvert are large slabs weathered out of the outcrop that contain numerous fossils and sedimentary structures. Collect all you wish from these loose slabs, but please do not collect in situ fossils or structures as ongoing research is continuing at this time. Please do not disturb any bedrock in place that has elevation or benchmark data recorded on it. These are reference markers for future study and should not be disturbed.

As we work our way up the large exposure, note the changes in lithology. Is the sequence fining or coarsening upward? What about the coarsest sandstone beds? Do they thicken upward over the entire vertical sequence or are there no discernable trends? Notice the slump scars (?) and what's directly above them. Could loading of several storm-deposited decimeter thick sandstone beds above the mudstone have contributed to the slump, or is some other mechanism warranted here such as seismic shaking (see discussion earlier by Bishuk, this paper). Are there any channel forms? We will try to answer these questions at the outcrop.

Note the variability in the morphology of the soft-sediment features. Relative size, orientation, internal stratification, and depth of penetration into underlying beds all vary throughout the section. Does there appear to be a correlation between the size of pillows and the thickness of the overlying sandstones. Are these all just balls and pillows? Consider the possibility of load casts, foundered ripples, and channel forms. Interpretation of these structures at this locality has been a source of lively discussion among the authors. Much work remains concerning the genesis of these sandstones.

Fossil coquinites and bioclastic mudstones are common at the base of storm layers, i.e. very fine to fine sublitharenites. This gives way to planar cross stratification, wave ripples, and thin mudstones with extensive bioturbation at the top of most sequences. These deposits are interpreted as tempestites by Applebaum (in prep.). Internal morphologies vary from one storm bed to another. For a recent discussion of vertical stratification sequences in storm-deposited beds see Myrow and Southard (1991).

39.3	0.5	Return to vehicles. Proceed to end of Bainbridge exit off-ramp. Turn right onto Route 206 westbound.
39.7	0.4	Junction 7. Continue west on Route 206.
40.2	0.5	Intersection of Rte 206 with Mt. Pleasant Road.
40.4	0.2	Reference marker: Road winds uphill to the left. On your left, look carefully for back of sign for eastbound traffic saying "Welcometo Bainbridge".
41.1	0.7	A guardrail can be seen on both sides of the road with a small outcrop on the left and a stream passing beneath the road. A gorge begins on the left. Park cars at the dirt pull-off on the right.

STOP 3A: STREAM GORGE EXPOSURE OF THE HUMMOCKY CROSS-STRATIFIED FACIES, GLEN AUBREY FORMATION (?), WEST OF BAINBRIDGE (Figure 3)

The exposures in the stream gorge south of Route 206 contain massive bedded sandstones, amalgamated sequences of balls and pillows, and most notably several key horizons of calcareous bioclastic conglomerates that may prove useful as lithostratigraphic or biostratigraphic "key beds". These shell beds are undoubtedly thanatocoenose assemblages and contain mostly crinoid columnals, brachiopods, gastropods, and branching bryozoans.

Numerous erosional surfaces within shell beds indicate amalgamation. Numerous reoriented geopetal structures (e.g. mud-filled gastropoda) suggest reworking of bioclastic materials by storms under waning flow conditions during the closing phases of the individual storm event. Fine mud resuspended by weak oscillatory wave currents settles out into areas of shelter porosity (underneath brachiopoda valves or inside Bellerophon chambers, see Figure 23). Similar fabrics have been documented by Kreisa and Bambach (1982).

41.1	0.0	Return to vehicles. Proceed westbound on route 206.
41.7	0.6	Outcrop can be seen on both sides of the road near crest of hill. Park cars on shoulder.

STOP 3B: HUMMOCKY CROSS-STRATIFIED FACIES, GLEN AUBREY FORMATION (?), WEST BAINBRIDGE (Figure 3)

This stop has some interesting features on the south side of the road. Of particular note are the high angle dips on some of the sandstone beds. Are these slump structures or channel forms? At the present time a definitive interpretation is unavailable. We welcome any suggestions. Surprisingly puzzling is the outcrop just north of the road, which contains no such distortion of bedding with all horizontal and undisturbed stratification. Can you explain the sudden change in attitude of the bedding over such a small distance, i.e. 10 meters.

41.7	0.0	Return to vehicles. Proceed on Route 206 eastbound back to I-88.
44.4	2.7	Turn left onto I-88 eastbound.
47.7	3.3	Take exit 9 off I-88 at Sidney. Bear right onto Route 8 south.
48.5	0.8	Large exposure of the Glen Aubrey Formation can be seen on both sides of the road. Turn right into large parking area.

STOP 4A: HUMMOCKY CROSS-STRATIFIED FACIES AND DARK GRAY SHALE FACIES, GLEN AUBREY FORMATION, ROUTE 8, JUST SOUTH OF SIDNEY (Figure 3)

Location: A huge exposure of the hummocky cross-stratified facies and dark gray shale facies can be seen on both sides of Route 8 located approximately 1 mile south of I-88 off-ramp.

Description: At this outcrop, we will look primarily at the dark gray shale, facies, which is found as three interbeds within the hummocky cross-stratified facies. The dark gray, facies records three brief transgressive periods, which represent deposition on deeper portions of the shelf. The hummocky cross-stratified facies represents storm deposits within storm base, and is similar here in many aspects to stops 2 and 3. The dark gray shale, facies is characterized by dark gray shale, a complete absence of invertebrates and biogenic structures, and locally abundant, listric truncation surfaces. The listric geometry of the truncation surfaces is consistent with submarine gravity-slide structures. The truncations are sharp and regular without any obvious local channels or erosional irregularities, which is best explained by shear rather than current scour. The rotational slumps were probably triggered by storm events or overloading causing liquefaction and foundering of the sediment.

48.5	0.0	Return to vehicles. Turn left onto Route 8 south and immediately get into right lane.
49.0	0.5	Turn right onto Thorpe Road.
49.2	0.2	Drive car to the end of the guardrail. Outcrop is at your left. At the end of the guardrail, make a 3-point turn and park cars single file along the right-hand guardrail.

STOP 4B (OPTIONAL): AMALGMATED PORTION OF THE HUMMOCKY CROSS-STRATIFIED FACIES, THORPE ROAD, NEAR CREST OF SIDNEY MOUNTAIN (Figure 3)

The sharp, scoured and loaded contact in the middle of the outcrop is interpreted as the contact between lower shoreface deposits (amalgamated hummocky cross-stratified facies-above fair weather wave base) and shelf deposits (hummocky cross-stratified facies-below fair weather wave base and above storm wave base). The predominance of fine sand sublitharenite above this contact, and a marked upward decrease in fauna suggest that the amalgamated hummocky cross-stratified portion of this facies occupies the lower shoreface and indicates increasing environmental stress associated with the nearshore. Siltstone and shale interbeds are rarely preserved, indicating that waves or currents (e.g., longshore-, tidal-, and storm-driven) had effectively winnowed the fine fraction. Hummocky cross-stratification is the predominant structure in all sandstone beds at the outcrop. The top surface of the outcrop displays excellent HCS beds in plan view and a variety of biogenic and sedimentary structures which cap HCS beds. Some of the observable features include wave ripples exibiting chaotic patterns, asymmetrical ripples with transverse ribs, conglomerate sheets consisting of flat shale intraclasts, paired vertical burrows of Arenicolites, and "figure 8"-shaped burrow of unknown ichnogenera. To reach the top of the outcrop, please walk around to the extreme right-hand side of the outcrop.

Near the base of the outcrop, an unusual interbed of grain-supported conglomerate grades to fine sand sublitharenite. The conglomerate is the coarsest unit found in the study area. Conglomeratic clasts are subrounded, with an abundance of siltstone and shale clast lithologies. These clasts commonly have Paleozoic weathering rinds. Other components include fish fragments, pockets of diagenetic galena, and plant debris occassionally replaced by galena. Graded bedding is the predominant structure consisting of alternating coarse sand and fine sand sublitharenite. Interpretation of this horizon is still pending, although it is suspected that it is a storm-generated deposit.

49.3	0.0	Return to vehicles. Turn right onto Route 8 south and get immediately into left lane.
49.6	0.3	Make first left onto Delaware County Route 4.
49.7	0.1	Bear right and proceed straight onto unnamed road. (Notice satellite dishes on unnamed road on the left).
49.8	0.1	Just after satellite dishes, bear left onto gravel and dirt road.
49.9	0.1	Bear left onto most used road and park cars at gate to Sidney Mountain bluestone quarry.

STOP 5: PLANAR-BEDDED FACIES, MOTTLED MUDROCK AND SANDSTONE FACIES, PLANAR-LAMINATED AND TROUGH CROSS-BEDDED REDBED FACIES, AND THE HUMMOCKY CROSS-STRATIFIED FACIES, SIDNEY MOUNTAIN QUARRY (Figure 3)

Location: From the gate described above in the road log, proceed on foot on dirt road for 200 meters. Walk around left side of quarry storage shed. Road will lead to the bottom of the quarry at its northern-most end. Each facies noted above can readily be seen by traversing back toward the south along the east wall of the quarry. Proceed with caution while walking on the quarry grounds. The terrain is rough and there is an open pit.

Description: Sidney Mountain quarry has the most complicated stratigraphy in the study area and exibits a multitude of sedimentary features. As a result of its complexity, we direct you to the sections on planar-bedded facies, mottled mudrock and sandstone facies, and hummocky cross-stratified facies (especially Fig. 7).

HYDROGEOLOGY OF GLACIAL DRIFT IN THROUGH VALLEYS NEAR DRYDEN AND CORTLAND, NEW YORK

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U.S. Geological Survey

INTRODUCTION

The Appalachian Plateau physiographic province encompases nearly all of southwestern New York (fig. 1, inset). About 85 percent of this region consists of till-mantled bedrock uplands; the remainder consists of broad valleys that are partly filled with stratified drift. The stratified drift includes coarse-grained sand and gravel deposits that are the only highly productive aquifers in the region; it also includes extensive fine-grained sediments that do not yield usable amounts of water. The divide between streams that drain northward to Lake Ontario and those that drain southward to the Susquehanna River crosses some of these broad valleys, which are termed "through valleys" because they are continuous across a major topographic divide. An area of unusually hummocky topography near the divide in each through valley suggests the former presence of buried ice; these areas were collectively termed the Valley Heads moraine by Fairchild (1932).

This article briefly describes current concepts of (1) the stratigraphy of stratified-drift aquifers and confining units in valleys within and south of the Valley Heads moraine, (2) recharge to those aquifers, and (3) their potential for use as sources of large water supplies during dry periods. These concepts were developed over the past few years from geohydrologic studies, cited herein, that emphasized or included valleys near Dryden, Harford, and Cortland (fig. 1). The article concludes with the log of a field trip that is intended to illustrate several of these concepts and to offer an opportunity for discussion of what is and is not known about these aquifer systems.

SOME GENERALIZATIONS ABOUT STRATIGRAPHY OF THE DRIFT IN VALLEYS OF THE APPALACHIAN PLATEAU

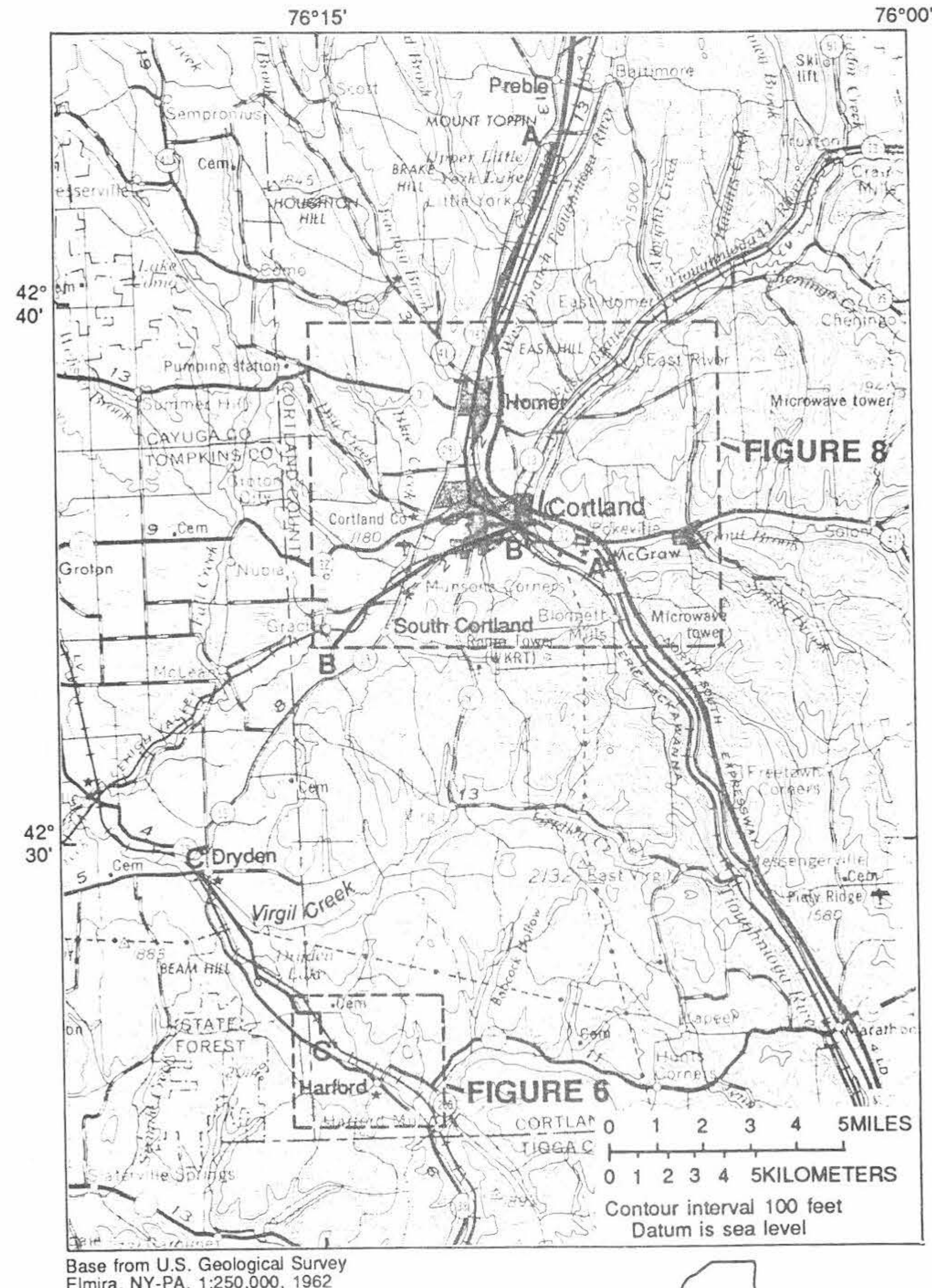
Valley fills south of the Valley Heads moraine consist of three facies:

1. Early facies: coarse sand and gravel that was deposited against the ice sheet as deltaic kame terraces along the valley sides and(or) as subaqueous fans atop bedrock in midvalley.

2. Middle facies: fine-grained sediment that was deposited in large lakes beyond the edge of the ice sheet.

3. Late facies: outwash and(or) postglacial alluvial deposits that constitute a surfical layer of coarse sand and gravel.

These facies are time-transgressive; that is, while meltwater was depositing early ice-contact sand and gravel in one reach, it was also depositing middle and(or) late facies in other reaches downvalley. The three facies form a characteristic stratified-drift stratigraphy (coarse over fine over coarse) in many valley reaches, although all three facies are not present at every location. In general, the greater the depth to bedrock, the greater the thickness of lacustrine clay, silt, and fine sand within the valley fill. Although depth to bedrock in valleys of the Susquehanna River basin ranges from 70 feet to as much as 500 feet, the thickness of water-yielding coarse sand and gravel rarely exceeds 150 feet and is commonly much less. Therefore, thickness and transmissivity of stratified-drift aquifers in these valleys is unlikely to correlate well



Base from U.S. Geological Survey
Elmira, NY-PA, 1;250,000, 1962

EXPLANATION

B———B' GEOLOGIC SECTION--
shown in figure 2 or 3

Figure 1.--Location of geologic sections and places discussed.

with depth to bedrock. Geologic sections at Cortland (fig. 2) are typical of many valley reaches south of the Valley Heads moraine. Surficial outwash tends to be especially thick close to the moraine, as shown in section B-B' (fig. 2).

Valley fills within the Valley Heads moraine are commonly a few hundred feet thick, and have been described as consisting mostly of till and fine-grained lacustrine sediment, with scattered thin lenses of coarse sand and gravel (Crain, 1974, p. 70-71; MacNish and Randall, 1982). The inferred distribution of surficial and buried stratified-drift aquifers is depicted on a map by MacNish and Randall (1982) that encompasses the Susquehanna River basin and on a map by Miller (1988) that encompasses all of central New York. Several geologic sections representing aquifer geometry in valleys near Cortland are given by Miller and others (1981) and Reynolds (1987).

STRATIGRAPHY IN THROUGH VALLEYS WITHIN AND NEAR THE VALLEY HEADS MORAINE

This article describes two through valleys within and near the Valley Heads moraine; one is in the towns of Harford and Dryden, the other extends from the city of Cortland southwestward into the town of Cortlandville.

Valley at Harford and Dryden

The stratigraphy of the valley fill in Harford and Dryden is depicted in figure 3 in a geologic section and a summary diagram. The drift in this valley is predominantly till and lake deposits interlayered with less abundant sand and gravel. From land surface to a depth of 100 feet, till forms about 50 percent of the drift, lake deposits 20 percent, and sand and gravel 30 percent. From depths of 100 to 300 feet, lake deposits are slightly more abundant than till, and only small amounts of sand and gravel are present. Coarse sand and gravel is present as multiple discontinuous layers, each generally less than 15 feet thick. The dashed lines in figure 3 are speculative and suggest more continuity than may actually be present. The sand and gravel might have originated in several ways:

1. As Collapsed Outwash. Meltwater might have deposited south-sloping outwash (valley trains), in part over buried ice. When the ice melted, the outwash collapsed to altitudes lower than the present divide. The stratigraphy and altitude of layers penetrated by the five southernmost boreholes in figure 3 are similar enough to suggest such an origin.

2. As Interstadial Alluvium. Alluvium must have been deposited by generally northward-draining local streams during intervals between ice advances. Such deposits could be expected to slope toward the valley axis (where deposited by tributaries) and northward along the valley axis. Also, they probably contain mostly fragments of local bedrock, because till in the uplands and alluvium of tributaries from the uplands contain few exotic stones derived from regions to the north. The slope and pebble content of the thin surficial deposits labeled Sd (fig. 3) and discontinuous lenses labeled C1 (fig. 3) are generally consistent with such an origin.

3. As Subglacial Channels or Fans. The sand and gravel layers north of the divide that lie below the altitude of the divide might have been deposited by south-draining meltwater, in subglacial channels, or as

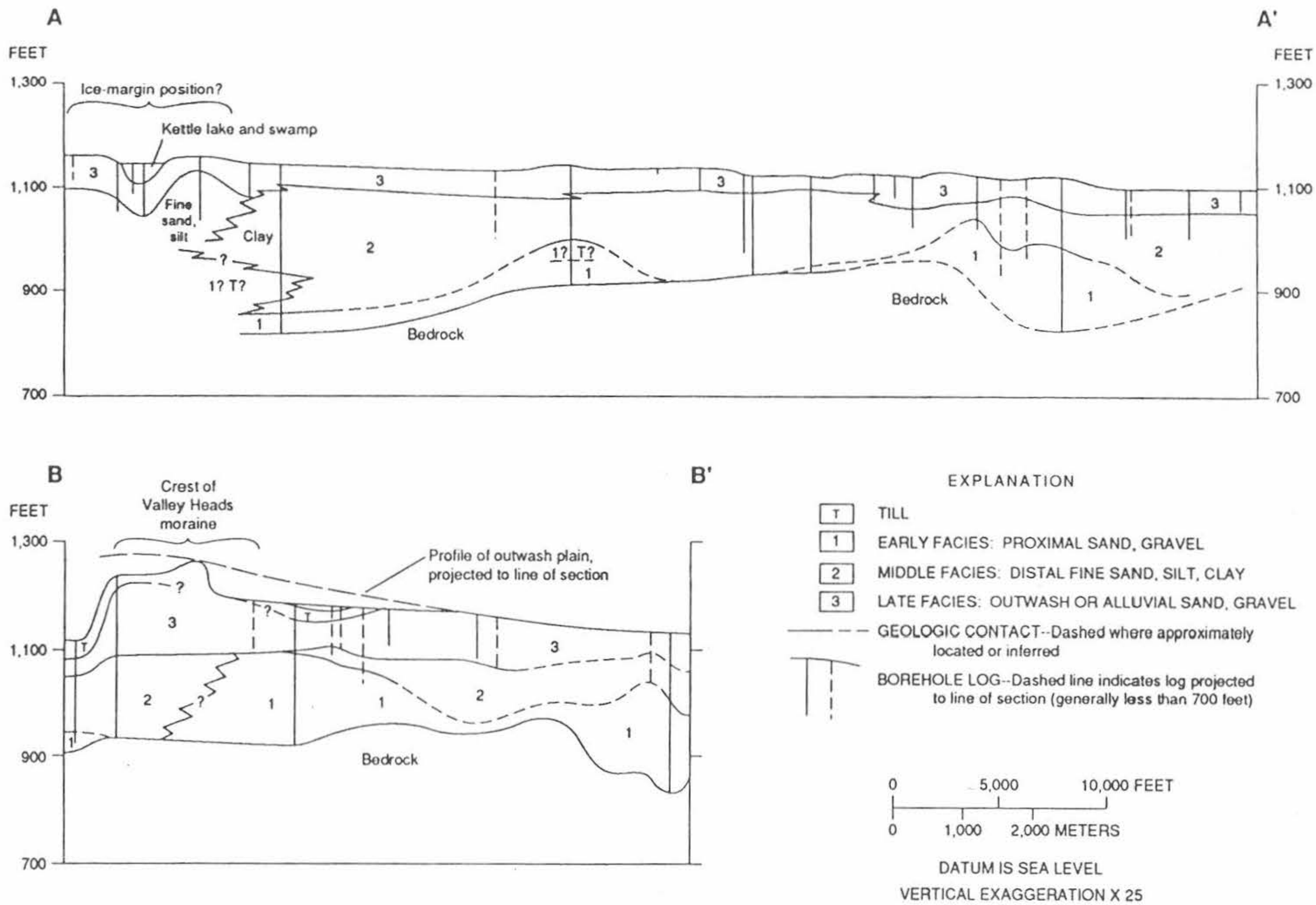


Figure 2.--Geologic sections near Cortland. Traces of sections are shown in figure 1. Logs and exact locations of most boreholes are given in Reynolds (1987), Miller and others (1981), or references cited therein; a few are recent unpublished data (T.S. Miller, U.S. Geological Survey).

A-A': Along West Branch Tioughnioga River valley at Homer and Preble, and Tioughnioga River Valley at Cortland.

B-B': Along Otter Creek-Dry Creek valley at Cortland and South Cortland. Meltwater spilling across the moraine late in deglaciation lowered land surface below the outwash plain in midvalley.

subaqueous fans where the channels emptied into a proglacial lake. Fans deposited in large proglacial water bodies during deglaciation have been widely reported in the literature (Thompson and Smith, 1983; DeSimone and LaFleur, 1986; Miller, in press) but are probably not a major component of the drift here. The downward-coarsening stratigraphy (lake fines over sand and gravel over till) that characterizes subaqueous fans at a retreating ice margin is not prominent in figure 3, and, as noted earlier, lake deposits form only 20 percent of the upper 100 feet of drift.

4. As Kame Moraine Derived from Stagnant Ice. North of Dryden Lake is an area of hummocky moraine that consists of diamicton (till or debris-flow deposits) and fluvial and lacustrine deposits. The composition and complex interlayering of these deposits (fig. 4) suggests ablation, resedimentation by mass movements, and the inversion of topography that occurs when debris slides off high parts of the ice surface and accumulates in low places, only to become topographic highs after the ice melts. Lake deposits and subaqueous fans are not prominent in the section; if proglacial lakes formed during retreat of the later ice sheets, perhaps the water drained away through tunnels or crevasses before most of the ice melted. Thereafter, localized deposition from and upon stagnant ice would have predominated. Layers C4, C5, and parts of C3 (fig. 3) contain a substantial percentage of exotic pebbles, suggesting deposition by streams flowing from a melting ice tongue (rather than by tributaries from the uplands). They also have hummocky top surfaces that include southward slopes and are suggestive of kames. The water level in layer C4 in the spring is as much as 12 feet higher than that in C3, suggesting that C4 is connected to some source of recharge at higher elevation, most likely kame terraces or kame moraine on the valley side.

Among the notable geologic aspects of the stratigraphy in Dryden Lake-Harford valley is the surficial or near-surficial till layer near the divide (D1 in figs. 3 and 4). After more than 100 feet of sand and gravel had accumulated in Harford valley south of the divide, the last glacial event was a readvance that extended more than a mile south across outwash (fig. 3). The ice then dissipated, leaving no lacustrine sediment and only minimal outwash. (Much of the sand and gravel that locally overlies the till layer is postglacial alluvial-fan deposits). The absence of surficial lake beds implies that the ice of the last readvance was thin and that, before a large lake could form between the ice and the saddle at Harford, the ice stagnated to the point that meltwater drained northward and westward through crevasses to some lower saddle. Surficial or near-surface diamicton layers that extend a short distance south over outwash have been observed in several other valleys along the Valley Heads moraine, which suggests some regional pattern of ice dynamics, perhaps a surge (Randall and others, 1988).

Otter Creek-Dry Creek Valley at Cortland

At the crest of the Valley Heads moraine southwest of Cortland, 60 to 100 feet of surficial outwash and ice-contact sand and gravel were deposited during the Valley Heads glaciation (unit 3 in fig. 2B). These sediments overlie less permeable silty sand and gravel that may correlate with kame deposits that mantle the bedrock hillside at South Cortland and were deposited during the older (Olean) glaciation. Records of wells that penetrate the moraine indicate a discontinuous till layer near or at land surface that was deposited during a readvance of ice during the late stages

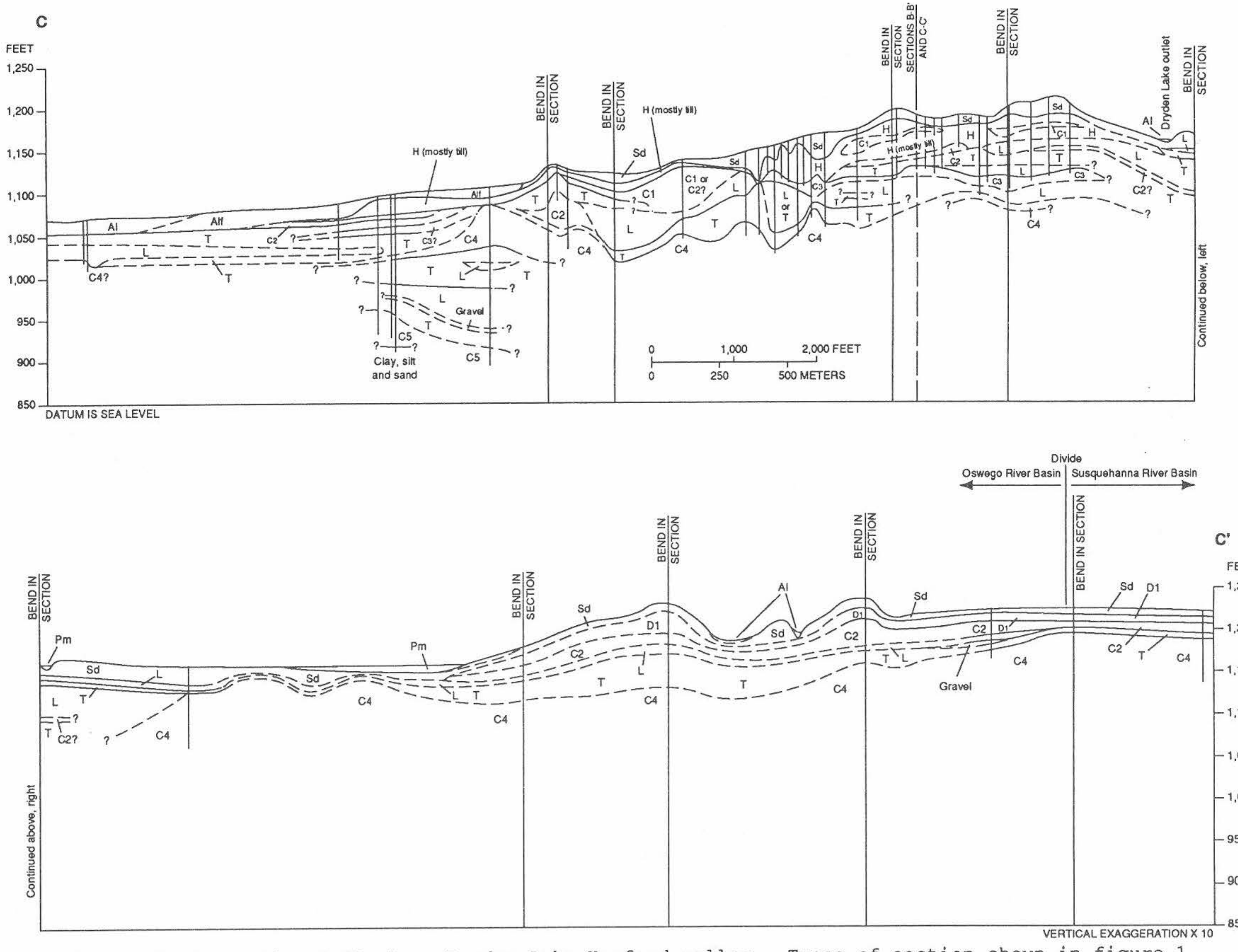


Figure 3.--Geologic section C-C' along Dryden Lake-Harford valley. Trace of section shown in figure 1. Diagram depicts correlation of deposits; arrows and compass points indicate direction of meltwater drainage away from each deposit. (Modified from Miller, in press; identification, narrative logs, and exact locations of boreholes are given therein.)

E X P L A N A T I O N

Sd STRATIFIED DRIFT--Alluvial, kame, outwash, and inwash sand and gravel deposited during retreat of last Valley Heads readvance; typically 3 to 10 feet thick; generally unsaturated except in low areas which have unconfined conditions

Al ALLUVIAL CHANNEL AND FLOOD-PLAIN DEPOSITS--Mostly gravel and sand that may be overlain or interbedded with overbank silt

Alf ALLUVIAL FAN--Gravel and sand

Pm PEAT AND ORGANIC-RICH SILT--Deposited in kettles

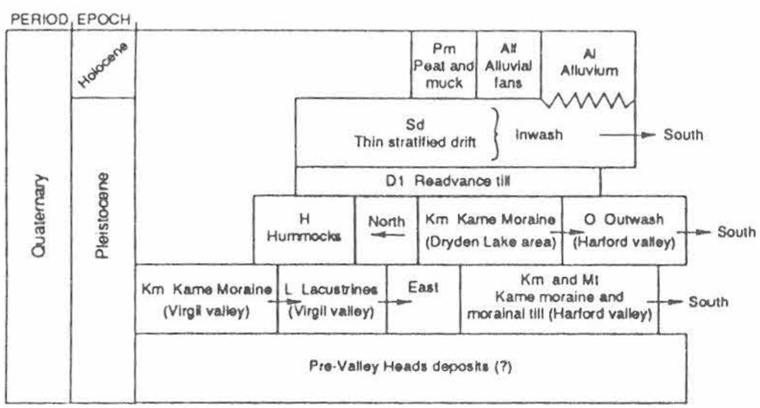
D1 DRAB TILL--Uppermost till deposited during last readvance

H HUMMOCKS--Complex of till and debris flow, glaciofluvial and glaciolacustrine deposits

T TILL--May be any till unit older than D1; moderate to bright clasts

L LAKE DEPOSITS--Fine sand, silt, and clay

K KAME--Ice-contact deposit of sand and gravel



CONFINED WATER-YIELDING ZONES

C1 CONFINED ZONE 1--Thin and discontinuous sand and gravel lenses that underlie the upper till in the hummocky area in the central part of the aquifer; drab pebbles suggest an alluvial or inwash origin; partly or seasonally saturated

C2 CONFINED ZONE 2--Thin and discontinuous sand and gravel; bright pebbles suggest a glaciofluvial origin such as outwash or ice-contact deposits

C3 CONFINED ZONE 3--Semicontinuous to continuous sand and gravel in the northern part of the study area; origin is uncertain

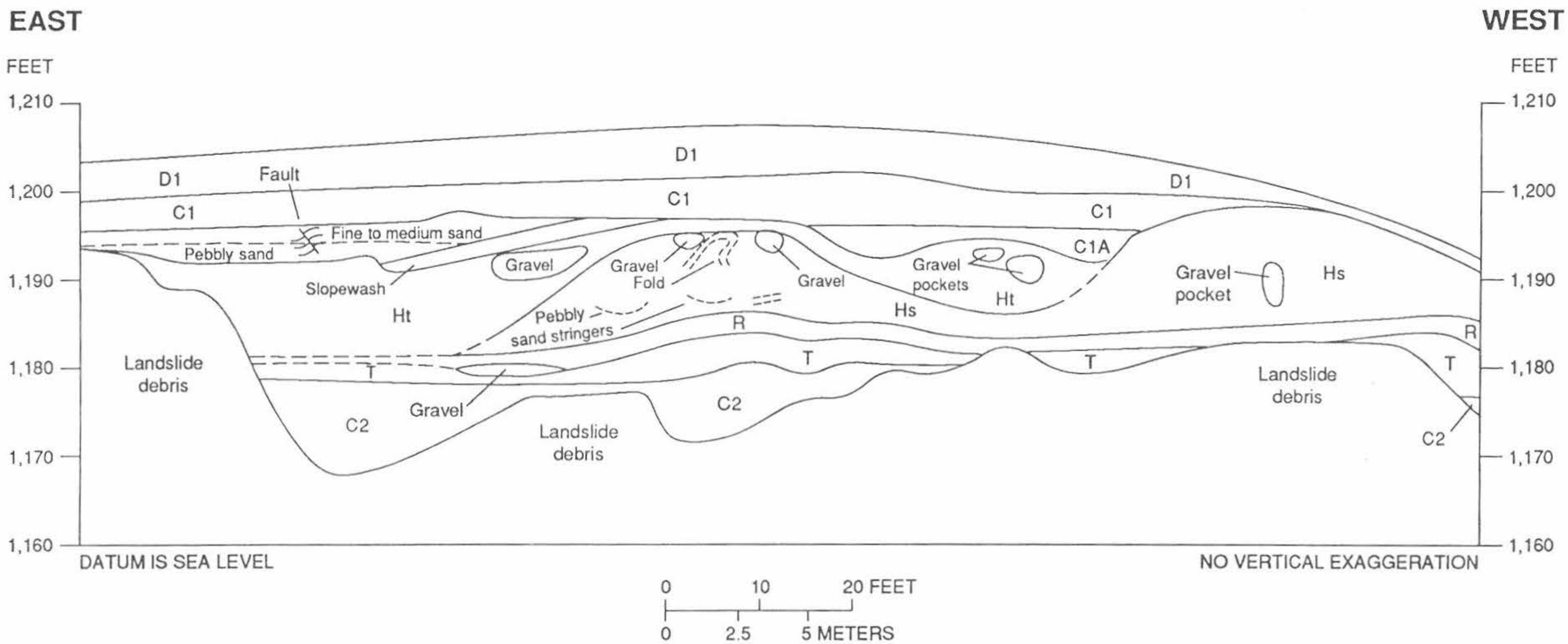
C4 CONFINED ZONE 4--Continuous and sand and gravel; an undulating surface and bright clasts suggest an ice-contact origin such as kames

C5 CONFINED ZONE 5--Extent determined only in a small area in the northern part of the study area; consists of silty sand and gravel; bright pebbles and an upper surface that slopes to the southeast suggest an ice-contact origin

GEOLOGIC CONTACT--Dashed where approximate

BOREHOLE--Log available

Figure 3.--Geologic section C-C' along Dryden Lake-Harford valley. Trace of section shown in figure 1. Diagram depicts correlation of deposits; arrows and compass points indicate direction of meltwater drainage away from each deposit. (Modified from Miller, in press; identification, narrative logs, and exact locations of boreholes are given therein.) (continued)



EXPLANATION

- D1 — TILL--Drab, sandy clayey silt matrix, sparse stones
- C1 — CONFINED ZONE 1 (upper part)--Coarse cobble gravel, moderately bright
- C1A — CONFINED ZONE 1 (lower part)--Pebbly sand, moderately bright, approximately horizontal bedding
- Ht — HUMMOCK TILL--Silt matrix, moderately stony, moderately bright, some gravel pockets
- Hs — HUMMOCK SILT--Massive bedding with some sand stringers; rare pebbles, gravel pockets, and folded bedding
- R — RHYTHMITE--Varved silt and clay with some dropstones
- T — TILL--Moderately bright, very bright, similar lithology as below
- C2 — CONFINED ZONE 2--Coarse cobble gravel, very bright, poorly stratified
- ——— — — GEOLOGIC CONTACT--Dashed where inferred

Figure 4.--Sketch of cutbank along Virgil Creek at Southworth Road, Dryden (stop 8), as exposed in 1984. (From Miller, in press, fig. 8.)

of Valley Heads glaciation. The till layer also mantles the lower parts of the kame deposits on the hillsides. The buried kame deposits consist largely of poorly sorted, silty sand and gravel with silt lenses; therefore, hydraulic conductivity and well yields are smaller than those typical of the surficial outwash. For example, a test-drilling program to locate a water supply at an industrial property included installation of eight test wells and test borings, but the two most productive sites yield only modest amounts of water -- 150 gallons per minute with 22 feet of drawdown, and 235 gallons per minute with 17 feet of drawdown.

The valley fill at the proximal (back or western) side of the Valley Heads moraine at South Cortland consists mostly of fine-grained sediments (such as till and lacustrine fine sand, silt, and clay) with only small amounts of sand and gravel. Water-supply wells at a fish hatchery at Gracie Road (fig. 1) penetrated mostly till and lacustrine sediment, then relatively thin confined aquifers at depths of 130 to 200 feet below land surface. These confined aquifers also yield moderate amounts of water (100 to 200 gallons per minute) to production wells.

Meltwater issuing from the Valley Heads ice deposited 45 to 100 feet of well-sorted coarse sand and gravel as outwash in front of the moraine. The outwash extends northeastward from the crest of the moraine at South Cortland to the eastern part of Cortland and then follows the Tioughnioga River valley to the southeast (fig. 2). Outwash overlies an extensive fine-grained lacustrine layer 60 to 150 feet thick that, in turn, overlies a basal sand and gravel zone atop bedrock. The basal sand and gravel is 10 to 30 feet thick in the Cortland-Homer-Preble valley, but is 50 to 170 feet thick in Otter Creek valley, southwest of Cortland (Section B-B', fig. 2; see also Miller and others, 1981). The surficial outwash aquifer is highly productive and capable of yielding several thousand gallons per minute to large-diameter wells. For example, the city of Cortland wellfield pumps 4.0 million gallons of water per day, and individual wells at the well field can pump 2,000 to 4,000 gallons per minute. Horizontal hydraulic conductivity of the surficial outwash aquifer is greatest, 1,000 to 2,000 feet per day, near the head of the outwash at South Cortland and decreases with increasing distance from the moraine. It is 1,000 feet per day in the central part of Cortland and 500 feet per day further east near the Tioughnioga River (Cosner and Harsh, 1978; Reynolds, 1987).

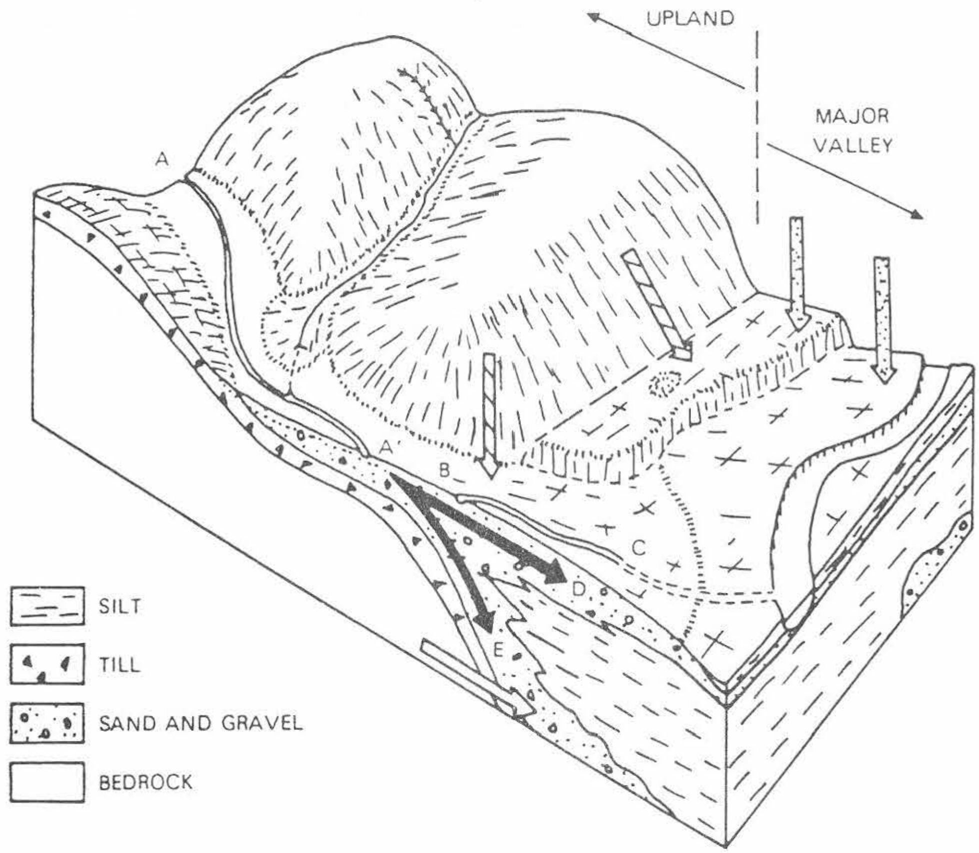
SOURCES OF RECHARGE TO STRATIFIED DRIFT IN VALLEYS OF THE APPALACHIAN PLATEAU

Under natural (unpumped) conditions, surficial stratified-drift aquifers in the Appalachian Plateau receive recharge from the three sources listed below, as illustrated in figure 5 and explained by MacNish and Randall (1982), Morrissey and others (1988), and others.

1. Infiltration of precipitation on the aquifer. Part of the precipitation on surficial sand and gravel is returned to the atmosphere by evapotranspiration, but the remainder infiltrates to become recharge, except in ponds or swamps, where the water table is at land surface and precipitation runs off directly to streams.

2. Runoff from upland hillsides that border the aquifer. Most stratified-drift aquifers are bordered by hillsides of till-covered bedrock. Till in the Appalachian Plateau contains a large percentage of silt and clay and is poorly permeable, so only a small part of rain and snowmelt can

infiltrate beyond the top foot or two. The excess water moves downslope in rivulets or through shallow openings in the soil. Where a stratified-drift aquifer lies at the base of the hillside, runoff infiltrates the permeable sand and gravel and percolates to the water table. Where a stream channel abuts the base of the hillside, the runoff enters the



SOURCES OF RECHARGE TO SAND AND GRAVEL IN MAJOR VALLEYS

Unchanneled storm runoff and (or) ground-water flow from upland hillsides

Regional ground-water flow through bedrock

Precipitation on the valley floor

Tributary-stream infiltration (see letters A - E)

A ═ A Tributary reach that gains water from ground-water discharge

B Edge of major valley, where wider, thicker sand and gravel deposits transmit more water than alluvium in tributary valley, so that water table drops below tributary channel and tributary loses water to the underlying aquifer

C Tributary goes dry here, where cumulative seepage loss equals streamflow at A'

D Recharge that moves laterally to river

E Recharge that moves downward to deeper aquifers

Figure 5.--Sources of recharge to stratified drift in valleys of the Appalachian Plateau. (From Morrissey and others, 1988, fig. 1.)

stream. In addition to hillside runoff, a small but steady flow of ground water moves through the bedrock from upland areas toward the major valleys and into the stratified drift.

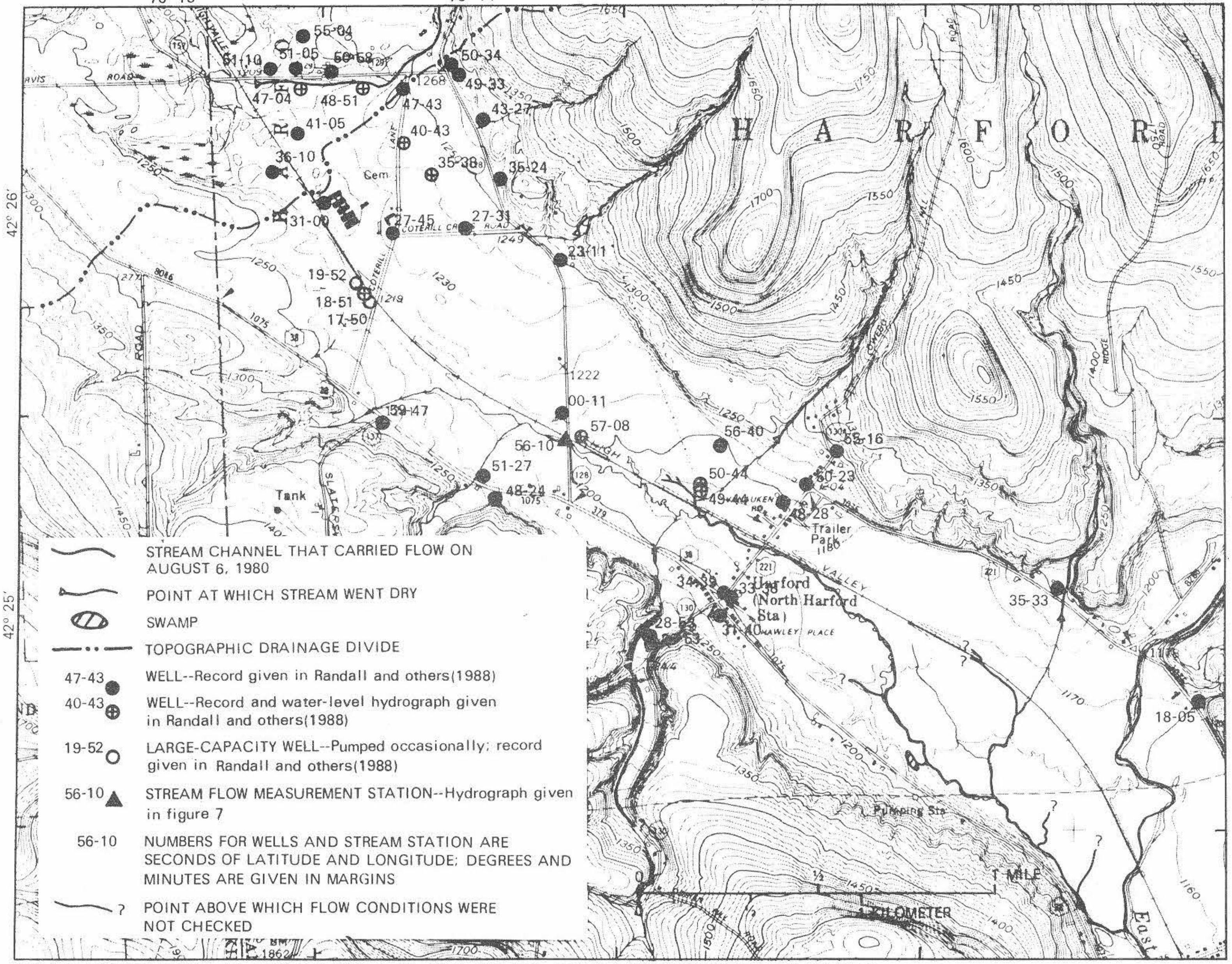
3. Infiltration from upland tributary streams. Several studies have shown that tributary streams draining upland basins can be important sources of recharge to stratified drift. Small streams in the Appalachian Plateau of New York go dry seasonally because their flow seeps into the streambed where they leave their upland valleys and cross stratified drift or alluvial fans in the larger valleys. This phenomenon was noted by Wetterhall (1959) as a source of recharge and described by Ku and others (1975) as a typical feature of the region. Crain (1966) demonstrated that seepage from tributaries on alluvial fans near the sides of Cassadaga Creek Valley in southwestern New York is a principal source of recharge to a gravel layer tapped by municipal wells beneath 100 feet of silt and clay. Randall (1978) investigated the magnitude and distribution of seepage from tributary streams in south-central New York and concluded that seepage rates were small at the edges of the main valley but were at least 1 cubic foot per second per 1,000 feet of channel farther downstream. More recent studies in the Appalachian Plateau of Pennsylvania (Williams, 1991) led to similar conclusions.

After reviewing studies of recharge in several localities, Morrissey and others (1988) concluded that upland runoff can be the largest source of recharge to stratified-drift aquifers under natural conditions in much of the glaciated Northeast. The percentage of recharge derived from upland sources tends to increase as topographic relief increases and valley width decreases. In areas of moderate to high relief, such as the Appalachian Plateau, upland runoff typically provides at least 75 percent of total natural recharge to valleys $^1/_2$ mile wide, and at least 60 percent to valleys 1 mile wide.

Recharge from Tributary Streams in Through Valley at Harford

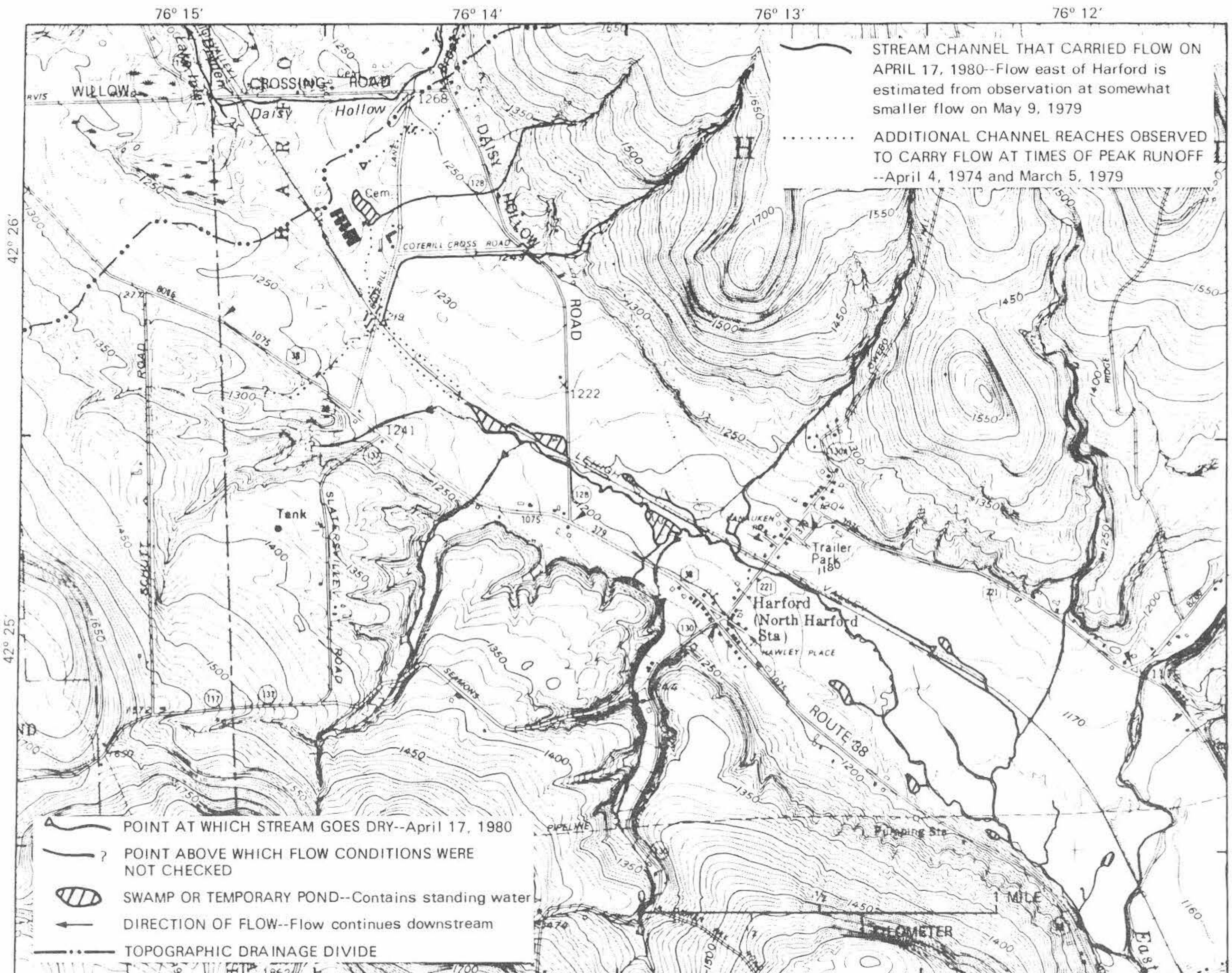
The distribution of streamflow in Harford valley is a function of geologic conditions (Randall and others, 1988). Small streams originate in the uplands, lose water by seepage as they flow across stratified drift in the valley, and during much of the year cease flowing a short distance downstream from where they begin to cross the stratified drift (fig. 6A). Occasionally, however, during the spring snowmelt period and during periods of unusually heavy rain, water becomes ponded in low-lying areas, upland tributaries flow continuously to the valley axis, and streamflow eastward along the valley axis begins just west of Cotterill Lane (figs. 6A, 6B). The extent of the stream network during peak runoff conditions and a few days after the last of several runoff peaks in March and early April 1980 are shown in figure 6B.

Surface runoff along the valley axis normally begins in a narrow wetland about 0.4 mile upstream of the bridge at Harford (fig. 6A), although it may begin further upstream for a time in the spring (fig. 6B). Whenever the tributaries go dry, all flow in the mainstem along the valley axis is derived from ground-water discharge. The mainstem becomes a source of recharge during most periods of high runoff, however, as illustrated in figure 7, which compares stage in the mainstem at station 56-10 (fig. 6A) with water level in nearby well 57-08 from February 1979 through February 1980. The annual flow regime observed at this location may be divided into six periods, as described in the following paragraphs:



Base from NYSDOT, Dryden, NY, 1978, and Harford, NY, 1973, 1:24,000, contour interval 10 feet, datum is sea level

Figure 6A.--Stream network in Harford valley on August 6, 1980, and location of wells. (Modified from Randall and others, 1988, fig. 6A.)



Base from NYSDOT, Dryden, NY, 1978 and Harford, NY, 1973, 1:24,000, contour interval 10 feet, datum is sea level

Figure 6B.--Stream network in Harford valley on April 17, 1980. (From Randall and others, 1988, fig. 6B)

1. February 1979. Surface runoff from the uplands was negligible because air temperature was mostly below freezing, and precipitation fell as snow. The water table declined steadily.

2. Late February through early March. Warmer temperatures after February 24 and heavy rain on February 25 and March 5 resulted in snowmelt and abundant runoff. The mainstem carried surface runoff from the uplands downvalley past the measurement station, where stage rose to a peak about 3 feet above the channel bed. Seepage from the stream into the aquifer caused an abrupt 4-foot rise in the water table.

3. March 5 through at least April 22. Surface flow past the measurement station was continuous, and the water table near the valley axis remained within 0.5 feet of stream grade.

4. Late April through early October. Streamflow at the measurement station ceased near the end of April. The water table declined until early October as ground water flowed downvalley toward the wetland 0.4 mile northwest of Harford, the nearest point of discharge.

5. October through December. An inch or more of rain fell during each of four storms after September 1. The first storm had little effect on the water table at well 57-08, but each of the next three storms caused the water table to rise abruptly 1 to 3 feet (fig. 7). A smaller storm on December 24-25, perhaps augmented by snowmelt, produced a similar effect. The magnitudes of the water-table rises indicate that a principal source of recharge was seepage when upland runoff from these storms flowed briefly past the measurement station. This conclusion is supported by the following reasoning: Unsaturated sand and gravel typically contains 10 to 25 percent air-filled pore space available to be filled with water as the water table rises. If this pore space (specific yield) were 10 percent, the water table would rise 10 times the amount of rainfall; if 25 percent, it would rise 4 times the amount of rainfall. This relation, and the fact that some precipitation would not reach the water table if soil moisture had been depleted, indicate that a water-table rise of much less than 10 times the amount of rainfall could be expected if recharge were derived only from local rainfall. Water-table rises much more than tenfold were observed, however. Two measurements of stream stage indicate that runoff did in fact flow past the measurement station during the November 25-26 storm (fig. 7).

6. January-February 1980. Negligible precipitation and subfreezing temperatures resulted in little recharge. The water table declined steadily as ground water continued to drain downvalley.

Recharge from Tributary Streams in Otter Creek-Dry Creek Through Valley at Cortland

Two small streams, Otter Creek and Dry Creek, occupy the broad valley that extends from the Valley Heads moraine at South Cortland to the Tioughnioga River in the eastern part of Cortland. Before glaciation, valleys draining from the north, northeast, east, and south converged at Cortland and drained southwestward through that valley (Muller, 1966). Glaciation diverted the Tioughnioga River to a southeastward course through Blodgett Mills along a former north-draining tributary.

The headwaters of Otter Creek include an upland drainage basin and Stupke pond, which is fed by ground water in the middle of the valley at South Cortland (fig. 8). Otter Creek loses water to the aquifer between the valley wall and its confluence with Stupke Pond outlet, then gains water from the aquifer between that confluence and the bedrock hill (umlaufberg) at Cortland. The reach between the umlaufberg and the Tioughnioga River is generally a losing reach.

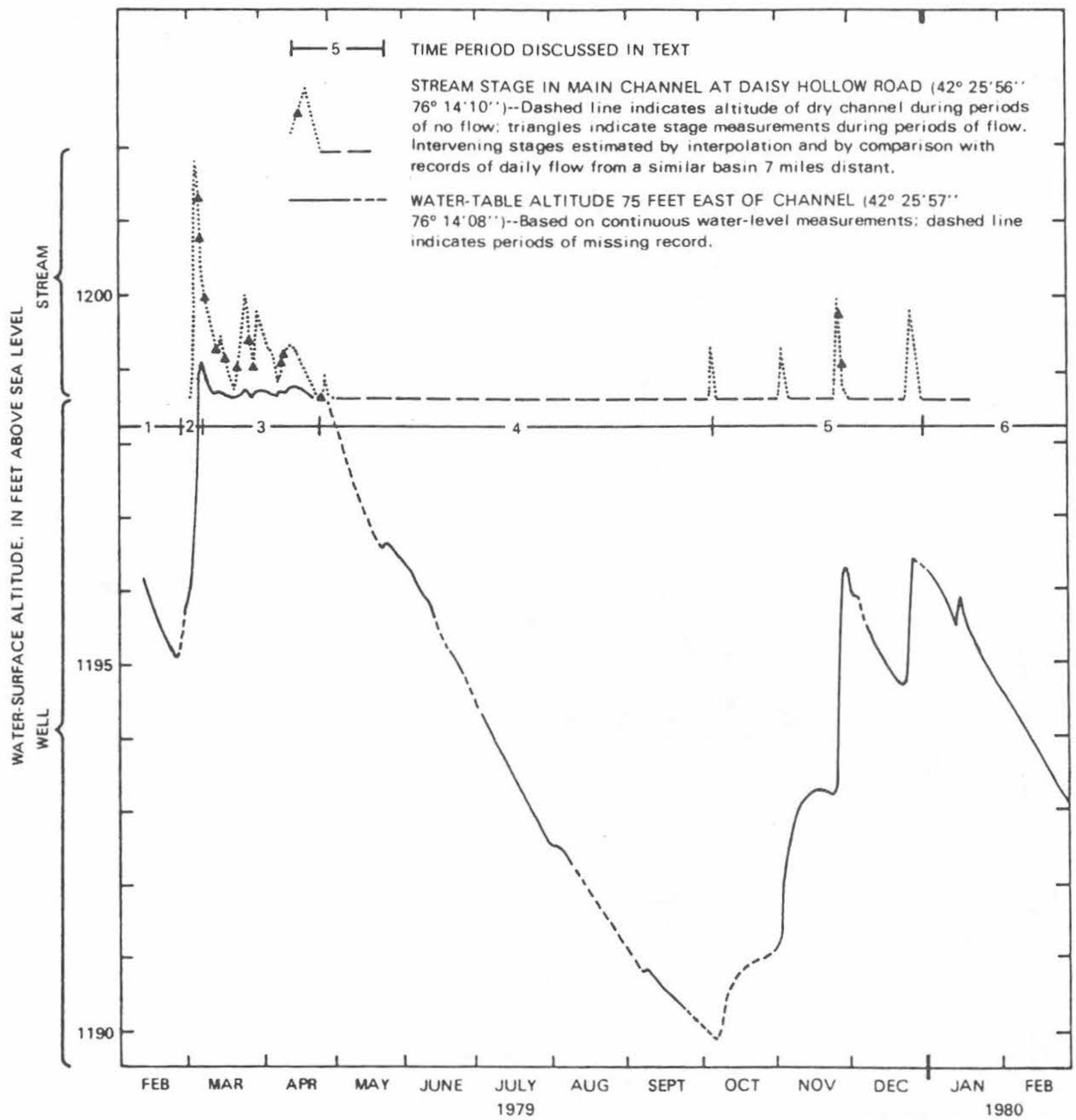


Figure 7.--Stream stage and water-table altitude near the stream in the headwater reach of Harford valley, where streamflow is intermittent. Location of measurement sites shown in fig. 6A. (From Randall and others, 1988, fig. 7.)

Dry Creek is a losing stream throughout the year along its entire reach over the valley floor but carries some flow to the Tioughnioga River during about 11 months in a year of normal precipitation. Dry Creek and the upland branch of Otter Creek dry up from the downstream to upstream direction during low-flow conditions in late summer and fall. By contrast, the main channel of Otter Creek dries up from upstream (near Stupke pond) to downstream, which is typical of small headwater tributaries that follow the axes of through valleys.

Seepage from Dry Creek to the outwash aquifer is controlled largely by the hydraulic conductivity of the streambed or, more likely, the alluvium near the stream. An unsaturated zone beneath the channel of Dry Creek was apparent during construction of two wells adjacent to the creek; at these wells, the water table is below the channel at all times of the year.

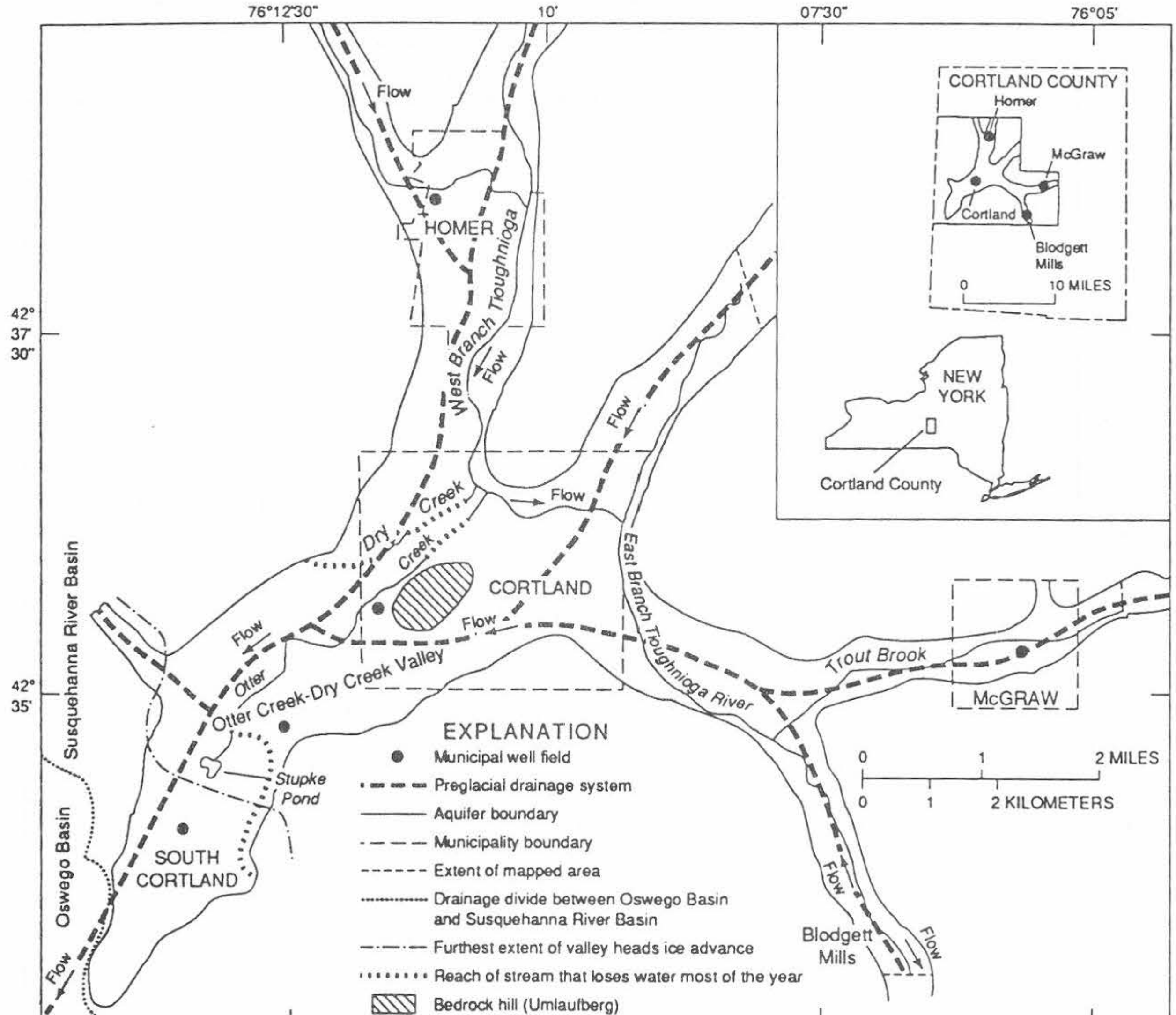


Figure 8.--Location and major geographic features of Otter Creek-Dry Creek valley. (Modified from Reynolds, 1987.)

Thus, if the water table were to decline as a result of increased pumping, seepage from Dry Creek probably would not increase much. By contrast, Otter Creek is hydraulically connected to the aquifer in most places, and large pumpage does affect streamflow. When Cortland municipal well 3 is operated during periods of low flow, the nearby reach of Otter Creek dries up; flow resumes when the well is turned off (James Roberts, Cortland Water Dept., oral commun.). Large pumpage southwest of Stupke Pond would intercept ground water that would normally discharge into the pond, its outlet, and Otter Creek. Therefore, large pumping from this area would cause streamflow to cease sooner in summer and resume later in the fall than during nonpumping conditions.

THROUGH VALLEYS AS POTENTIAL SOURCES OF SEASONAL GROUND-WATER SUPPLIES THAT DO NOT DEPEND ON STREAMFLOW

Streams that follow the axes of large stratified-drift valleys in the Appalachian Plateau normally gain water along their entire length by ground-water discharge from the stratified drift. Pumping from surficial aquifers can reverse the natural water-table gradient toward these streams, however, and thereby induce stream water to recharge the aquifer. This potential for induced recharge far exceeds natural recharge from precipitation and upland runoff in most broad valleys underlain by stratified-drift aquifers. Depletion of streamflow by induced infiltration may be undesirable, however, in periods when streamflow is naturally low and needed for other purposes.

In at least 29 localities along the northern perimeter of the Susquehanna River basin, the drainage divide crosses broad valleys whose floors are underlain by sand and gravel that could provide large yields of ground water from storage during drought periods. Because streams in these localities are small or nonexistent, large withdrawals would not cause equally large concurrent reductions in streamflow downvalley, such as would occur if the same amounts were pumped from aquifers that are crossed by large streams (Randall and others, 1988). These anomalous valley reaches are termed "through valleys." Both the Harford valley and the Otter Creek valley at Cortland are examples, although Otter Creek valley abuts a large stream at one end and hence was classified as a "separated valley" by Randall and others (1988, p. 3). Large-scale seasonal ground-water withdrawals in a through valley would require several wells that tap the aquifer near the divide. The concept is illustrated in figure 9. Large ground-water withdrawals during the summer (fig. 9A) would lower the water table near the divide and would reduce ground-water discharge to the head of the stream that drains the valley axis, and, perhaps, to an equally small headwater stream on the other side of the divide. At the end of the period of seasonal need, the pumps would be shut off, and recharge during the following winter and spring would gradually refill the water-table depression (fig. 9B). Abnormally large or prolonged seepage losses from streams could be expected then, but these losses would be only a small percentage of the large streamflow that occurs then and hence would be of little consequence.

Several possible scenarios for seasonal development in Harford valley were evaluated by use of a numerical ground-water flow model calibrated to transient conditions (Randall and others, 1988). Withdrawal of 10.8 million gallons per day for 2 months in summer near the divide would lower the water table as much as 33 feet near production wells and would cause the point at which streamflow begins along the valley axis to migrate 1,900

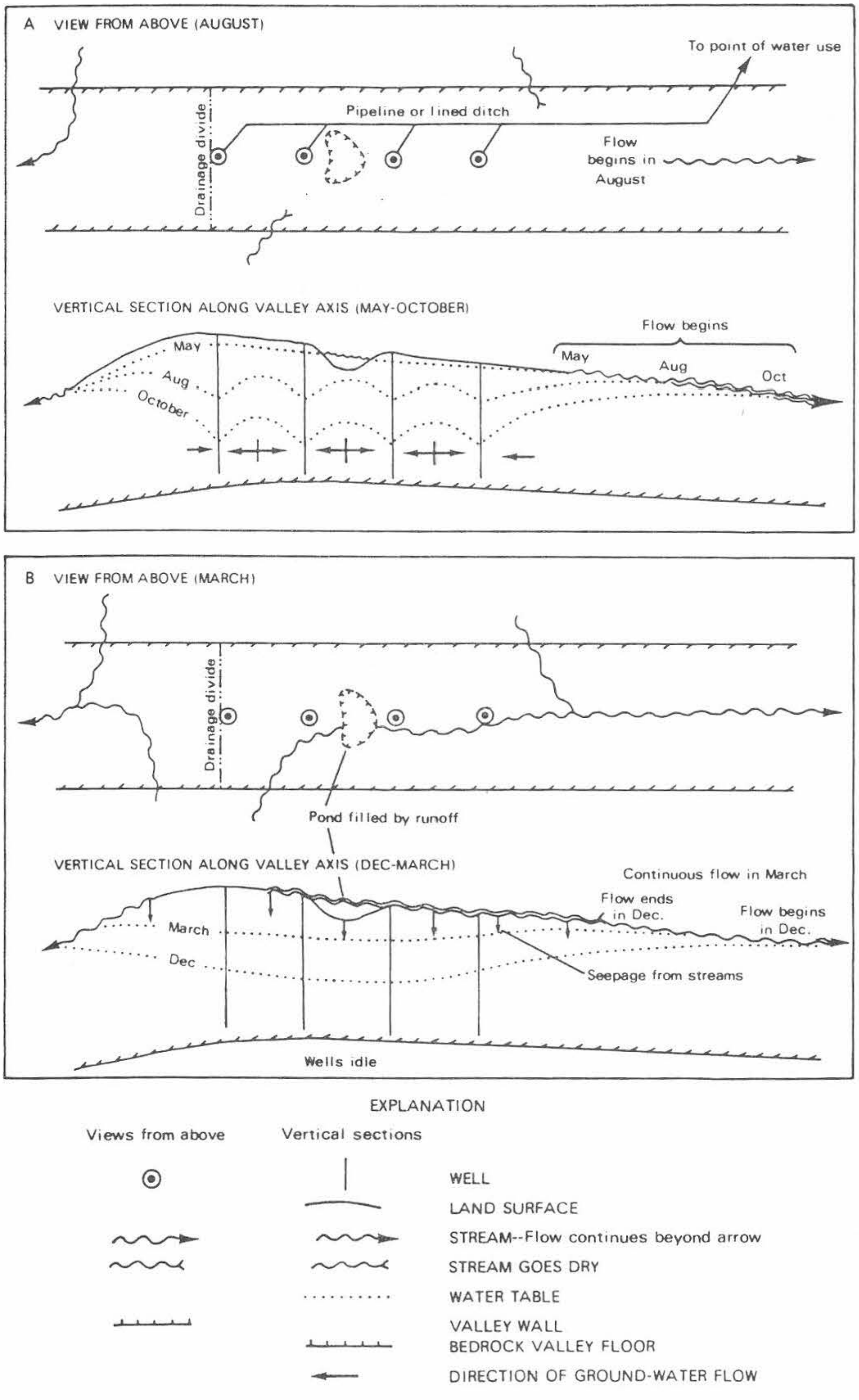


Figure 9.--Ground-water levels and directions of flow in an idealized through valley: A. In summer, with large seasonal ground-water withdrawals. B. In winter, after large seasonal ground-water withdrawals have ceased. (From Randall and others, 1988, fig. 3.)

feet downvalley. Recharge from stream seepage would be greater than normal during the following winter and spring because the aquifer would not fill up to stream grade as quickly as under natural conditions (fig. 9). This increased recharge would allow the same seasonal withdrawals to be repeated each year. The simulated effect of seasonal withdrawal on streamflow in Harford valley is depicted in figure 10, along with the calculated effects

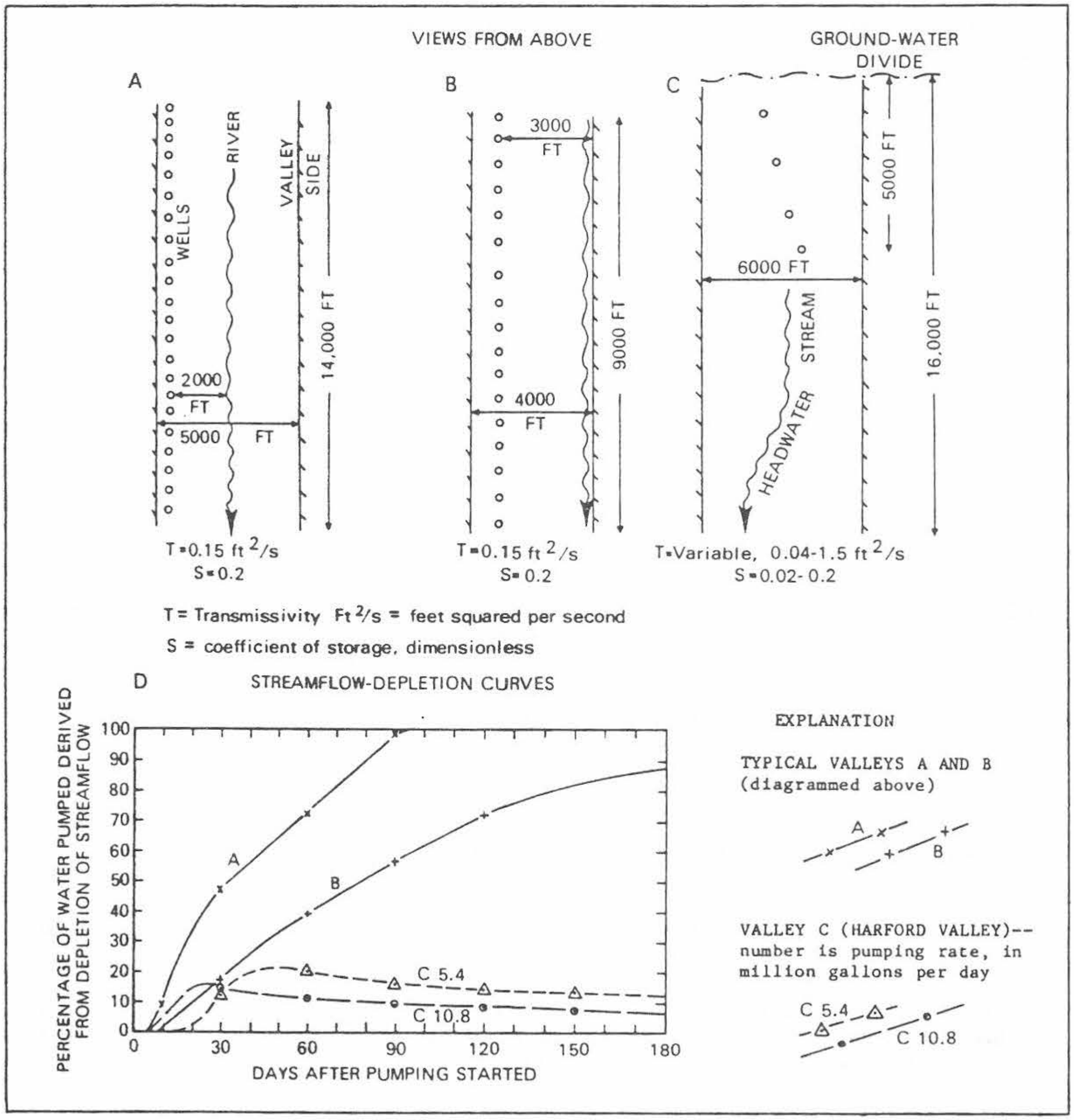


Figure 10.--Streamflow depletion caused by ground-water withdrawals in through valleys compared to that in typical valleys during prolonged drought: A, B. Arrangement of stream and wells postulated by Seaber (U.S. Geological Survey, written commun., 1967) in two typical valley reaches. C. Arrangement of stream and wells postulated in Harford through valley. D. Streamflow depletion as a function of time and pumping rate. Water not derived from streamflow depletion is derived from storage. (From Randall and others, 1988, fig. 17.)

of withdrawals in typical valleys where induced recharge occurs. After 120 days of pumping in the typical valleys, 70 to 100 percent of the water pumped would be derived by depletion of streamflow, and the percentage would increase as pumping continued. After 120 days of pumping in the headwater reach of a through valley such as at Harford, however, the percent depletion of streamflow would be smaller and would decrease as pumping continued. These simulations are discussed in detail by Randall and others (1988).

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The following log identifies several possible stops that illustrate topics of this field trip and other points of geohydrologic interest. Locations of these stops are shown in figure 11. The actual itinerary may differ; stops will be selected at the time of the trip, depending on current streamflow conditions, availability of exposures of surficial deposits, and results of current studies near Cortland.

Cumulative mileage	Miles from last point	
0.0	0.0	Intersection of Routes 221 and 11 at Marathon. Go north on Route 11.
3.6	3.6	Turn left on Route 392 to Messengerville; continue on Route 392 beyond Messengerville
7.3	3.7	Turn left on Tone Road.
7.4	0.1	Cross bridge and park.

STOP 1: SITE OF FORMER USGS GAGING STATION ON GRIDLEY CREEK.

This site is near a preglacial drainage divide now crossed by Gridley Creek, whose flow was reversed by glaciation. It is a good site at which to measure basin yield because the valley fill is only 500 feet wide and probably not thick; therefore, ground-water underflow through permeable sand and gravel is likely to be small, and nearly all runoff leaves the watershed in the stream, where it can be measured.

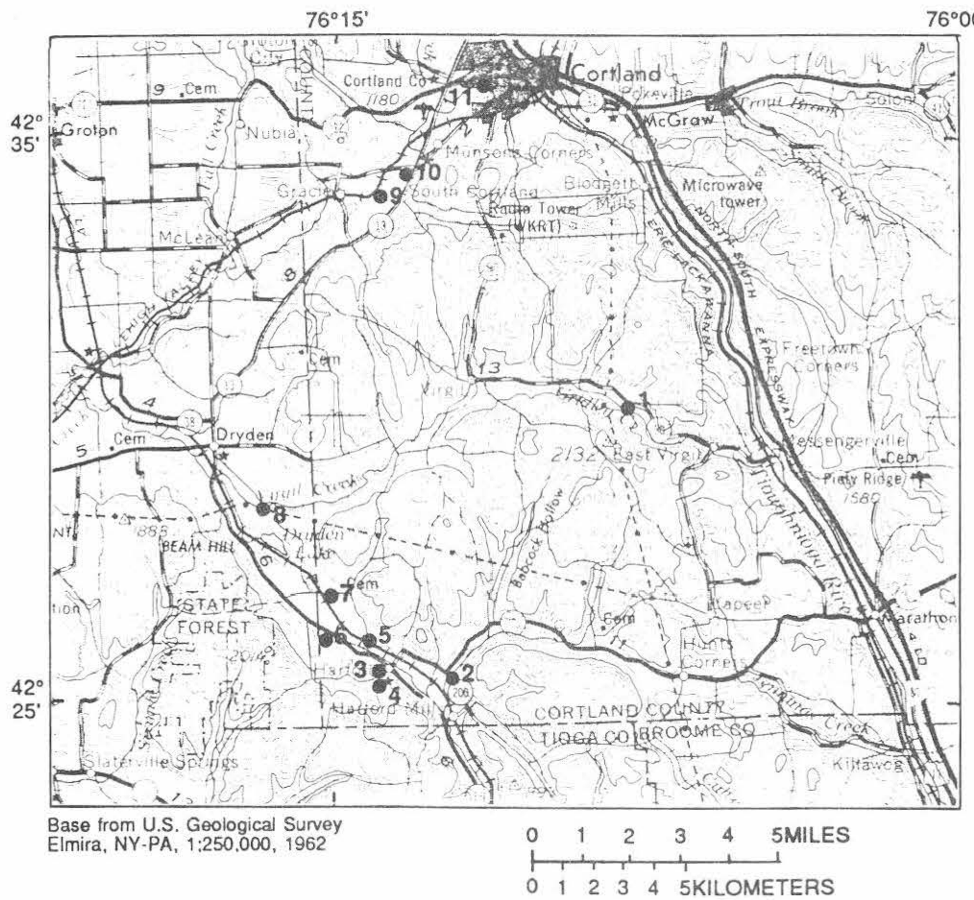

Figure 11.--Location of field-trips stops.

ROAD LOG (continued)

Cumulative mileage	Miles from last point	
7.5	0.1	Return to Route 392, turn left.
8.5	1.0	Ski area on left.
10.0	1.5	Meltwater spillway ahead on left, incised into south end of mound of drift that Route 392 crosses just ahead. The divide between east-flowing Gridley Creek and west-flowing Virgil Creek is here, in the middle of a valley 2,000 feet wide.
10.8	0.8	Approaching hamlet of Virgil, turn left on Vandonsel Road, go past sign for Power Farms.
11.9	1.1	Turn right.
12.2	0.3	Summit, view (to rear) across valley of Virgil and Gridley Creeks.
12.7	0.5	Cross Haucks Road.
14.6	1.9	T-junction, bear left.
16.4	1.8	Turn right on Route 221. Route follows East Branch Owego Creek. Kame terraces on right side of valley.
17.8	1.4	Turn left on Route 200. Kame terraces on both sides at intersection.
18.0	0.2	Pause, view to right.

STOP 2: HESITATION STOP: HARFORD UNDERGROUND GAS-STORAGE PROJECT.

In 1953, a cavity was excavated by solution mining in salt beds at a depth of about 3,000 feet to provide storage space for liquified petroleum gas that would arrive here by pipeline. The brine created by mining was discharged to lagoons excavated in the permeable outwash and alluvium on the valley floor. A few years later, many residents of Harford Mills, visible to the south, began to pump salty water from their shallow wells. Water sampling, resistivity surveys, and a test well disclosed that outwash extended to a depth of 56 feet and contained salty water over a large area near the hamlet and Owego Creek, except close to the water table, where water was fresh (Dunn, 1967, 1968). Gravel layers at depths of 97 and 170 to 194 feet yielded fresh water.

The raised, lined pond visible here is now used to contain brine produced during operation of this facility.

18.6	0.6	Junction, bear right and continue on Route 200 through hamlet of Harford Mills.
19.2	0.6	T-junction, turn right on Route 38.
20.7	1.5	Hamlet of Harford, turn left on Cheese Factory Road.
20.95	0.25	Abandoned creamery; park in driveway on south side.

ROAD LOG (continued)

STOP 3: CHEESE FACTORY BROOK.

Except during the spring freshet and other brief periods of unusually heavy runoff, this brook ceases to flow somewhere near or downstream from this former creamery. We hope to view the point of dryness, note exposures along the brook, and discuss the significance of seepage losses from tributaries as a source of recharge.

Cumulative mileage	Miles from last point	
21.95	0.2	Continue south on Cheese Factory Road, park along road. Walk across field to brook.

STOP 4: SPRINGS ALONG CHEESE FACTORY BROOK.

About 0.2 mile upstream from the creamery, a reservoir (now in disrepair) was built to develop a spring at the base of the bluff along Cheese Factory Brook. Exposures upstream and downstream show that the bluff consists of sandy, somewhat silty gravel. The presence of the spring suggests that impermeable sediment is not far below. On April 25, 1991, a tributary just upstream went dry 300 feet before it reached Cheese Factory Brook. Seepage from the tributary may be a major source of water to the spring.

21.6	0.45	Return to Route 38, turn left.
22.1	0.5	Turn right on Daisy Hollow Road.
22.3	0.2	Cross former railroad; park.

STOP 5: STREAMFLOW AND WATER-LEVEL MEASUREMENT SITES.

This is the location at which the water levels in figure 7 were measured. Most runoff from the 3.42 square-mile watershed above this point occurs as underflow through the stratified drift.

22.5	0.2	Return to Route 38, turn right.
23.2	0.7	Turn left into dirt road, opposite Cotterill Lane. Locked gate ahead, permission from Harford Teaching and Research Center farm manager required to enter; continue 0.1 mile to pit. Brook to left of dirt road flowed past Route 38 April 25,1991.

STOP 6: PIT IN CREVASSE FILLING.

This infrequently used pit is on the property of the State College of Agriculture's Harford Teaching and Research Center. About 10 feet of till that contains only rounded pebbles overlies about 20 feet of stratified, sorted but very silty gravel and, in places, deformed medium to fine sand, silt, and clay. Poorly sorted, silty gravel is characteristic of early-deglacial, proximal stratified drift near the Valley Heads moraine.

ROAD LOG (continued)

Cumulative mileage	Miles from last point	
23.3	0.1	Return to Route 38, cross Route 38, follow Cotterill Lane.
23.8	0.5	Harford Teaching and Research Center main buildings on left.
24.2	0.4	Turn left on Willow Crossing Road.
24.4	0.2	Park near gray house, walk down to channel of Daisy Hollow Brook.

STOP 7: DAISY HOLLOW BROOK.

Daisy Hollow brook flows farther out onto the valley-floor outwash than any other tributary near Harford. One reason why may be observed here. Also exposed is the flat-stone drab gravel that is typical of alluvium along upland tributaries.

24.8	0.4	Junction, bear right.
26.5	1.7	Turn right. This terracelike flat is capped by till.
26.9	0.4	Sharp turn left.
27.0	0.1	Cross Virgil Creek, turn left.
27.1	0.1	Park, cross field to left to view exposure along creek.

STOP 8: SOUTHWORTH ROAD EXPOSURE NEAR DRYDEN.

The large variety of sediments in this exposure illustrate the complexity of the depositional environment at the Valley Heads moraine. Figure 4 shows the appearance of the exposure in 1984. Sediment facies include till, fluvial, and lacustrine deposits. The drab till at the top of the exposure represents a readvance of ice during the late stages of Valley Heads glaciation. The surficial till overlies a coarse, cobbly gravel that in turn overlies either a discontinuous pebbly sand or till hummocks. Till hummocks have pockets of gravel, disturbed sand stringers, and some folded bedding, all of which suggest movement of the sediments such as may occur during topographic inversion. Beneath the hummocks is rhythmic silt/clay with scattered stones, presumably dropped from floating ice. A till containing abundant bright clasts underlies the rhythmite. A bright, coarse, cobbly gravel underlies the bright till.

28.1	1.0	Turn left (west) onto Maclintock Road.
28.5	0.4	Turn right (north) onto Route 38.
28.6	0.1	Straight at intersection at Dryden village, continue north, now on Route 13.
35.4	6.8	Turn left into entrance of gravel pit, stop 9.

ROAD LOG (continued)

STOP 9: GRAVEL PIT AT SOUTH CORTLAND

Coarse outwash and ice-contact sediments deposited by Valley Heads ice. Note boulders 2 to 3 feet in diameter at entrance to pit. Excavation is along the boundary between outwash deposited by northeastward-flowing meltwaters and ice-contact deposits laid down on and adjacent to blocks of ice that were in the middle of the valley. A geologic log of a test well in the gravel pit indicates the following stratigraphy, in feet below land surface: 0-45 feet sand and gravel, 45-50 till, and 50-75 sand and gravel. Mining of sand and gravel will stop 10 feet above the water table so as to minimize the effects on water resources and so that land could be used after mining operations cease.

Cumulative mileage	Miles from last point	
		Return to pit entrance, turn left (north) on Route 13.
36.2	0.8	Turn left (west) on Lime Hollow Road.
36.3	0.1	Pause, view stop 10.

STOP 10: HESITATION STOP: REMEDIATION OF A SPILL OF ORGANIC SOLVENT.

A recovery well pumps about 970 gallons per minute to remediate the source area (O'Brien & Gere, 1991). Pumped water is routed through an air stripper, which brings sufficient air into contact with the water to allow the volatile organic solvent to evaporate. The water is then discharged into infiltration lagoons, where it seeps to the water table.

36.4	0.1	Return to Route 13, turn left (north) on Route 13.
38.2	1.8	Turn left (north) on Broadway.
39.3	0.5	Turn left (west) into entrance to City of Cortland well field.

STOP 11: CITY OF CORTLAND WELL FIELD.

Otter Creek flows through the well field but is usually dry from midsummer to late fall. During low-flow periods, when well 3 is turned on and streamflow disappears in Otter Creek because of induced infiltration. When the well is turned off, streamflow reappears. Abundant vegetation in the channel indicates a relatively large amount of nutrients in the water. In contrast, the channel of nearby Dry Creek contains little or no vegetation, which suggests that runoff from that drainage basin contains relatively few nutrients.

An air-stripping tower was installed beside the main plant in anticipation that the plume of organic solvents (source about 2 miles southwest, stop 10) would reach the well field. Concentrations of trichlorethylene within the plume decrease with distance from the source, and only trace amounts have been detected at the well field, not enough to warrant use of the air stripper. The decrease in concentrations has been attributed to volatilization, discharge to Otter Creek, and biological degradation and transformation of the trichlorethylene.

Well 4 at plant 2 is one of the most productive municipal wells in New York. It is capable of pumping 4,000 gallons per minute. The well taps outwash 63 feet in saturated thickness. At present, it is used only intermittently.