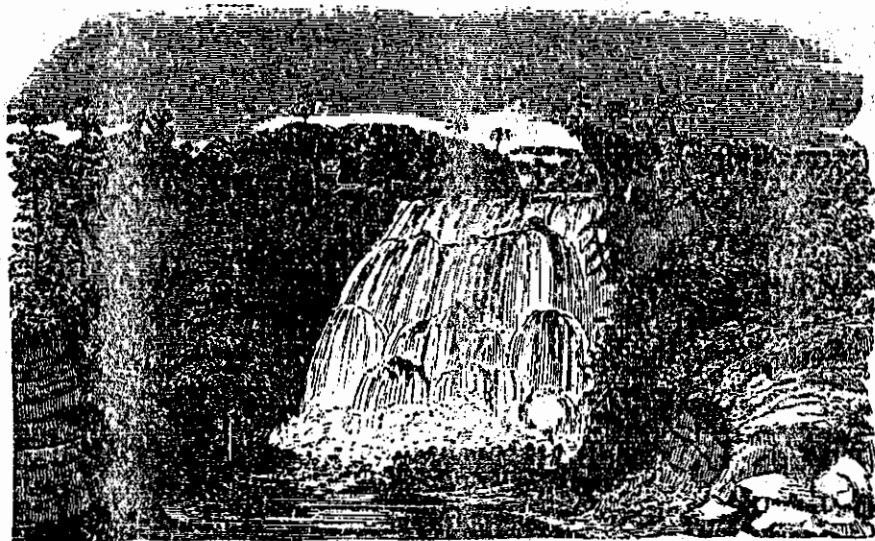


**NEW YORK STATE
GEOLOGICAL ASSOCIATION
58TH ANNUAL MEETING**

FIELD TRIP GUIDEBOOK



Fall creek, near Ithaca. From a sketch by Mrs. Hall.

**CORNELL UNIVERSITY
ITHACA, NEW YORK**

OCTOBER 10-12, 1986

**NEW YORK STATE
GEOLOGICAL ASSOCIATION**

58TH ANNUAL MEETING

FIELD TRIP GUIDEBOOK

**Department of Geological Sciences
and
Institute for the Study of the Continents**

**CORNELL UNIVERSITY
ITHACA, NEW YORK**

OCTOBER 10-12, 1986

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1. The first part of the document discusses the importance of maintaining accurate records of all transactions and activities. It emphasizes that this is crucial for ensuring transparency and accountability in the organization's operations.

2. The second part of the document outlines the various methods and tools used to collect and analyze data. It highlights the need for a systematic approach to data collection and the importance of using reliable sources of information.

3. The third part of the document focuses on the analysis and interpretation of the collected data. It discusses the various statistical techniques and models used to identify trends and patterns in the data, and how these can be used to inform decision-making.

4. The fourth part of the document discusses the importance of communication and reporting in the data analysis process. It emphasizes the need for clear and concise communication of the findings and the importance of providing regular reports to the relevant stakeholders.

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PREFACE AND ACKNOWLEDGMENTS

It is a pleasure to welcome all of you to the 58th annual meeting of the New York State Geological Association and to Cornell University.

Ithaca and the Finger Lakes region offer much to satisfy a wide range of geological interests. Field trips have been arranged to show this diversity. Topics include general geology, geomechanics, geomorphology, glacial geology, kimberlites, paleobiology, paleoenvironments, petrology of continental basement and the mantle, sedimentary petrology, stratigraphy, structure, and tectonics. I hope the field guide proves as lastingly useful as the one from the 1959 meeting, when NYSGA last met at Cornell.

I want to thank the field trip leaders and other contributors for their efforts on behalf of NYSGA.

John L. Cisne, President
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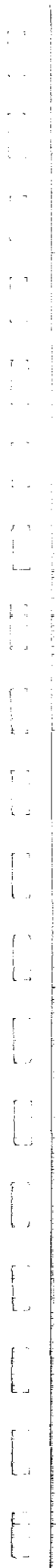


TABLE OF CONTENTS

	Page
PREFACE AND ACKNOWLEDGMENTS	iii
GEOLOGICAL SUMMARY OF THE CAYUGA REGION by D.E. Karig and L.T. Elkins	1
SUBMARINE EROSION ON THE DYSAEROBIC SEAFLOOR: MIDDLE DEVONIAN CORRASONAL DISCONFORMITIES IN THE CAYUGA VALLEY REGION by Gordon C. Baird and Carlton E. Brett	23
SEDIMENTARY CYCLES AND LATERAL FACIES GRADIENTS ACROSS A MIDDLE DEVONIAN SHELF-TO-BASIN RAMP, LUDLOWVILLE FORMATION, CAYUGA BASIN by Carlton E. Brett, Gordon C. Baird and Keith B. Miller	81
THE USE OF JOINT PATTERNS FOR UNDERSTANDING THE ALLEGHANIAN OROGENY IN THE UPPE DEVONIAN APPALACHIAN BASIN, FINGER LAKES DISTRICT, NEW YORK by Terry Engelder	129
FACIES AND FOSSILS OF THE ONONDAGA LIMESTONE IN CENTRAL NEW YORK by Howard R. Feldman and Richard H. Lindemann	145
STRATIGRAPHY AND PALEONTOLOGY - AROUND CAYUGA LAKE ONCE AGAIN by Thomas X. Grasso, Jonathan W. Harrington, and William T. Kirchgasser	167
PALEOENVIRONMENTAL ANALYSIS OF THE FIDDLERS GREEN FORMATION (LATE SILURIAN) IN NEW YORK STATE by Richard D. Hamell and Samuel J. Ciurca, Jr.	199
KIMBERLITES OF THE FINGER LAKES REGION by Suzanne Mahlburg Kay and Brayton P. Foster	219
ICE READVANCE IN THE VALLEY OF FIVEMILE CREEK, STEUBEN COUNTY, NEW YORK AND THE EFFECT OF THE RESULTING DEPOSITS ON POSTGLACIAL STREAMFLOW by Allan D. Randall	239
GEOMORPHOLOGY OF THE CAYUGA LAKE BASIN by Arthur L. Bloom	261



GEOLOGICAL SUMMARY OF THE CAYUGA REGION

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INTRODUCTION

No general summary of the geology in the region around Ithaca and Cornell University has appeared since that in the N.Y. Geol. Assoc. Guidebook of 1959. Not only has much information been gathered since that time but also the geological emphasis has shifted. Earlier studies stress the geomorphic and paleontologic aspects of the area, whereas more recent attention has been toward the physical stratigraphy, environments of deposition and tectonic evolution of the region. In addition, geophysical studies and deep drill holes have greatly expanded our knowledge of the subsurface geology and deep crustal conditions.

This chapter, therefore, is designed to synthesize the geologic information pertaining to the Cayuga region, emphasizing that material collected since 1959, and is oriented toward physical stratigraphy, tectonics and geophysics. The summary aspect of this paper precludes much detail, but extensive referencing in this and the following papers will enable the reader to pursue topics in greater depth.

GEOLOGIC AND TECTONIC SETTING

The Cayuga region (Fig. 1) is physiographically part of the Appalachian Plateau Province, which forms the cratonic flank of the foreland to the Alleghenian collision zone (Engelder, 1979b). In north central Pennsylvania the foreland fold and thrust belt trends nearly east west, as do the related structures on the foreland in central New York (Wedel, 1932; Engelder and Geiser, 1980). This obliquity with the general northeasterly trend of the Alleghenian collision zone marks the northern flank of a large salient in the structural front, which is in part related to the distribution of the Silurian salt (Davis and Engelder, 1985). There is, however, a similar salient in the internal part of the collision belt, which suggests that initial passive margin configuration (e.g., Rankin, 1976) and/or collision complexities were in large part responsible.

The Cayuga region has lain well within a continental interior during all of Phanerozoic time, yet close enough to the continental margin to show subtle effects of more pronounced tectonic events that shaped that margin from the late Precambrian into the Mesozoic (Colton, 1970; Engelder and Geiser, 1980). Thus the stratigraphy of this area bears witness to the formation of a passive continental margin during the late Precambrian

and Cambrian, the Taconic collision in the Ordovician, and the more enigmatic Acadian event of the mid to late Devonian age (Ettensohn, 1985; Bradley, 1983). Although younger sediments have since been eroded from the region, its structural framework is basically a result of the Alleghenian collision. The sum of these processes has produced gently southward tilted and thickening pile of strata (Rickard, 1973), but within that pile are sharply varying trends reflecting earlier events. These trends, primarily in the subsurface, have been delineated by regional stratigraphic and geophysical studies.

GEOPHYSICAL STUDIES

Regional gravity and magnetic data provide a primary source of information concerning the depth and character of Precambrian basement in south central New York. Sparse heat flow, seismologic, and electromagnetic studies give further insights into the deep crust and its behavior. A moderate number of seismic reflection profiles have been run in the Cayuga region, which should clarify structural and stratigraphic relations within the sedimentary column, but none of these have been released for public use.

Gravity

The Bouguer gravity field in New York State (Revetta and Diment, 1971) west of the Taconic zone basically reflects the depth to and density variations of the Precambrian basement. In central New York a very general southward decrease in the gravity field is associated with the 15 m/km southward dip of the basement surface (Hodge et al., 1982; Rickard, 1973). Superimposed on this gradient are sub-circular to more linear NNE trending anomalies. The character of the anomaly field changes markedly across a zone of steep, easterly decreasing gradients that trends NNEerly just west of Rochester (here referred to as the Clarendon-Linden gradient). To the east and including the Cayuga region, anomalies have lower amplitudes (10-15 mgal) and longer wavelengths than to the west. The eastward decrease in density, smoother field, and more circular anomalies have been interpreted as reflecting less mafic Precambrian rocks to the east, which have been intruded by both mafic and silicic plutons (Hodge et al., 1982).

Magnetics

The total field aeromagnetic maps of western New York (e.g., Zietz and Gilbert, 1981) show a good correspondence to the gravity field (Hodge et al., 1982). Thus the magnetic field east of the Clarendon-Linden gradient is more subdued than that to the west. In both areas most magnetic highs correlate with gravity highs and are attributed to mafic intrusions or amphibolites (Hodge et al., 1982).

Lows in both fields reflect granitic or anorthositic bodies. A larger ENE-trending gravity and magnetic low south of Ithaca may imply a syenitic to anorthositic mass in the Precambrian (Sneddon, 1983).

Heat Flow

Heat flow measurements in New York show approximately normal values for stable continental crust (50-60 mw/m^2) but the few measurements in the Cayuga region are significantly higher (65-94 mw/m^2) (Hodge et al., 1982). In general higher heat flow is attributed to the higher radiogenic content of underlying granitic masses.

Variations in temperature gradients differ from those of heat flow because thermal conductivities are a function of lithology. A compilation of temperature gradients in the upper 1 km (Hodge et al., 1982) show east-west trends reflecting surface lithology, but also define a N-S trending band of high gradients near $76^\circ 40'W$ longitude. Near Auburn, values rise to near 40°C/km , which provided the basis for the drilling of 2 geothermal wells (Hickman et al., 1985).

Seismicity

Although several regions in New York show significant seismic activity (Sykes, 1978; Nottis, 1983), central New York, and in particular that region east of the Clarendon-Linden gradient has been almost aseismic during historic time. This lack of seismicity correlates well with the other geophysical fields in outlining major subdivisions of the Precambrian basement in New York (Zoback and Zoback, 1980).

PHYSICAL STRATIGRAPHY AND DEPOSITIONAL SETTINGS

Emphasis in previous guidebooks on the Cayuga region has been placed on the description of the exposed Middle and Upper Devonian section. Beneath these strata, however, are over 2 km of older, Cambrian to Devonian strata that overlie a much older Precambrian basement. Several drill holes have now penetrated these rocks (Kreidler, 1963, 1972; Flagler, 1966), with which it is now possible to describe these rocks and to explore their regional variations (Figs. 2, 3 and 4). These data in turn permit interpretation of depositional and tectonic environments during early Paleozoic time.

Precambrian

The Precambrian basement has been sampled by four wells in the Cayuga region (Fig. 1). Additional inferences about the depth to basement and its lithology are obtained from geophysical data and from crustal inclusions in the Mesozoic dikes exposed in the region. Basement lithologies from wells have been only cursorily described. By well (located on Fig. 1) these are: Shepard #1; pink biotitic gneissic granite (Flagler, 1966, and Isachsen, 1971); Shaeffer #2, amphibolite (Flagler, 1966); Auburn Geothermal wells; grey and pink marble in one and marble and mica schist in the second (Brayton Foster, 1986, per. comm.). Crustal inclusions in the kimberlite dikes around Ithaca suggest mafic syenite and calcsilicate metamorphic rocks (Kay et al., 1983). These lithologies, in addition to the geophysical data, suggest that the Precambrian basement is comprised of a suite of medium-grade metasediments intruded by silicic to feldspathic (syenitic to anorthositic) plutons. This suite has been correlated with the Grenville province exposed to the north and east, and thus should

have a metamorphic age of about 1.1 by.

Cambrian

The oldest Phanerozoic strata in the Cayuga region are the upper Cambrian sandstones and dolostones of the lower Beekmantown Group (Potsdam, Theresa [Galway] and Little Falls formations). These mature shallow water clastics thicken to the east and south, toward the paleoshelf edge (Rickard, 1973).

In addition to the oceanward thickening of upper Cambrian strata, the section adds sequentially older strata in that direction. These Cambrian strata as well as those of the lower Ordovician represent a transgressive wedge of continentally derived material associated with the construction of a late Precambrian passive continental margin that roughly coincided with the present coastline (Fig. 5b).

Ordovician

The lower Ordovician series is represented in the Cayuga region by about 150 m of strata comprising the Tribes Hill Formation of the Beekmantown Group, (Committee on Stratigraphic Correlations, 1966). The Tribes Hill Formation represents a continuation of the pattern established during the Cambrian, with the south and eastward thickening of quartzose sands and dolostones (Rickard, 1973).

Conditions changed markedly at the end of early Ordovician time, when the Knox unconformity was nearly isochronously developed throughout the continental interior of the northeastern U.S. This widespread hiatus in deposition is more easily interpreted as a eustatic event (Vail et al., 1977) than as a result of a regional tectonic event, such as the Taconic collision. This interpretation is reinforced by the subsequent reestablishment of quiet marine shelfal conditions during the early middle Ordovician (Walker, 1973). The deposition of carbonates of the Black River Group testify to the lack of tectonic effect on the continental shelf in the Cayuga region during that period (Rickard, 1973).

Differential relief increased during deposition of the Trenton Group (Cameron, 1972). In the Cayuga region a shallow NE trending trough developed, behind a larger irregular arch to the east (e.g., Jacobi, 1981; Zen, 1973). The arch, which represented the outer swell to the encroaching Taconic Trough and island arc caused the development of westward and migrating unconformities and of the normal faults seen in the Mohawk Valley (e.g., Rodgers, 1971; Cisne et al., 1981).

Except for possible subtle downflexure of the crust on the cratonic flank of the outer swell and a few ash beds in the Trenton limestones, the Taconic orogeny left little imprint on the Cayuga region. Cessation of convergence by the end of the Middle Ordovician led to the filling of the Taconic Trough with clastics, which further depressed the surrounding region, and led to the expansion of basinal conditions. This elastic loading may explain the extension of the easterly derived clastics of the Utica Formation westward across the entire state.

Siliciclastics of the upper Ordovician Queenston Complex mark the continued westward transport of detritus from the Taconic collision zone (Martini, 1971), which appears to have ceased convergence but was still undergoing vertical adjustments. Facies bands within the Queenston, comprising a range from shelfal shales to fluvial or coastal plain sands, trend roughly north-south and mark repeated transgression and regression of the marine conditions that lay to the west (Hughes, 1976; Stone and Webster, 1978). The migrating shorelines during this period produced diachronous lithostratigraphic units that are relatively poorly dated and correlated in the region (Hughes, 1976).

Vertical displacements related to the crustal rebound or relaxation following the Taconic collision might be responsible for the upper Silurian cycles, but late Ordovician glaciation is at least as plausible a cause for the relative changes in sea level (Hughes, 1976).

Silurian

The Early Silurian was a period of continued tectonic quiescence and reduced eustatic variation. A reduced supply of sediment from the east was accompanied by shallow marine to supratidal conditions across the region (e.g., Fisher, 1954). Sediments consist of reworked older strata (e.g., Medina sands, Martini, 1971; Muskatt, 1972), carbonate units, and a striking abundance of iron rich units (e.g., the Clinton oolitic ironstones (Gillette, 1947; Zenger, 1971)). The lateral variations in these environments produced marked facies changes in many of the units of this age.

The Late Silurian saw the remarkable development of evaporates over much of the northeastern interior of the U.S. (Fisher, 1957; Kreidler, 1957; Alling and Briggs, 1961; Rickard, 1969; Treesh, 1972). Central New York lay under part of the Salina Basin, represented by the Vernon, Syracuse, Camillus and Bertie formations. The relative roles of oceanographic and tectonic effects in the creation of this basin are still unclear, but the northeast elongation (Fig. 5c), suggests that some tectonism was involved. Before the end of Silurian time, evaporite deposition had ceased and sedimentation was again dominated by very shallow marine to supratidal carbonates of the lower Rondout formations (Rickard, 1969; Laporte, 1967).

Devonian

Except for the thin Oriskany sandstone, the lower Devonian is represented by widespread, and laterally more uniform carbonates. The Helderberg group represents multiple westward transgressions of neuritic carbonates (Rickard, 1962; Laporte, 1967, 1969). Not only does this group thin westward but the upper units have pinched out westward by the longitude of Cayuga Lake (Rickard, 1969). The Onondaga Formation represents the last major limestone deposit before the westward flood of clastics from the massive uplift associated with the Acadian event. Even during the Onondaga deposition the Acadian event was heralded in the Cayuga region by the Tioga ash beds (Roen and Hosterman, 1982; Rickard, 1984). The extremely widespread distribution of these beds and their silicic and potassic composition suggests that they are the result of caldera eruptions with

continental crustal sources rather than of typical island arc eruptions.

With the deposition of the Hamilton group (Baird and Brett, 1981), clastic deposition in the "Catskill Delta" became dominant over carbonate deposition (Faill, 1985; Rickard, 1984), although the shales generally remain calcareous, and a few thin limestones punctuate the section (Dugolinsky, 1981; McCave, 1973). In eastern New York, the upper Hamilton group records the westward migration of the shoreline and the westward progradation of coarser facies, whereas in western New York the group consists of shelfal strata (Brett and Baird, 1982). The overlying thin Tully limestone marks a singular change in depositional regime, developing along a shelf edge between deeper basinal environment to the east and broad shelf conditions to the west (Heckel, 1973). Basinal conditions, and westward migration of coarser clastics developed in the Cayuga region during deposition of the Genesee formation (de Witt and Colton, 1978; Kirchgasser, 1985). The Genesee and succeeding mid and late Devonian section thicken and coarsen markedly southeastward (Fig. 5d) and are represented by more than 1500 m of sandstones and shales in the Cayuga region (Rickard, 1981; Tetratex, Inc., 1981; Van Tyne, 1982).

No younger strata remain in the Cayuga region, but estimates of overburden above the Devonian strata (see Engelder and Oertel, 1985) suggest that a maximum of 1 km of younger, probably Mississippian to Pennsylvanian sediments were deposited before the Alleghenian collision raised the region above sea level.

Igneous Rocks

The only igneous rocks of Phanerozoic age in the Cayuga region are kimberlite dikes of late Jurassic to early Cretaceous (140 m.y.) age (Basu et al., 1984). These dikes are sub-vertical, trend nearly north-south (Fig. 6), are a few centimeters to a meter wide and can be traced from a few meters to a kilometer along strike. Their mineralogy implies a relatively shallow mantle source and either a higher fossil geothermal gradient or a relatively undepleted mantle (Kay et al., 1983). The largest of these bodies probably extended to the surface and were associated with maar eruptions (Williams and McBirney, 1977). The occurrence of diatreme facies in the Poyer Orchard body (see Foster and Kay, this volume) implies that the level presently exposed is relatively close to the Cretaceous surface. The origin of these dikes, which occur sparsely but over a very widespread area of the eastern U.S., is still conjectural.

STRUCTURES

The Cayuga region lies in the very mildly deformed belt in front of the foreland fold and thrust belt that was developed during the Alleghenian orogeny. This anomalously wide zone of mild deformation is related to the distribution of Silurian salt, which is extremely weak and acted as a low-strength decollement, above which strata were displaced northward (Davis and Engelder, 1985). On the gently south dipping strata, which reflect the flexure of the foreland beneath the load of the Alleghenian fold-thrust belt, are impressed a suite of subtle but significant deformational structures. There are large very gentle folds and a few small thrust faults, but most shortening is absorbed by intragranular strain and pres-

sure solution effects (Engelder, 1979a, 1979b). In addition, the study of well-developed joint sets in the region has led to a number of very fruitful studies of the past and present stress field.

Folds

A series of gentle regional folds forms an arcuate pattern in south central New York. The northernmost of these traverse the Cayuga region with a general east-west trend. Surface mapping of these folds (Wedel, 1932) defined anticlines that rise above the regional dip with amplitudes of tens to near 100 m and wave lengths on the order of 10 km. Although very persistent along trend, the folds often show locally irregular axial traces. The south limbs of anticlines are generally steeper, defining folds that are either symmetric when corrected to the regional dip or even slightly asymmetric toward the south (e.g., Sherrill, 1934). All the folding appears to be restricted to the section above the Silurian salt (Prucha, 1968) but there appears to be some uncertainty as to the nature of the structure between the exposures in the middle Devonian and the decollement. Engelder and Geiser (1980) suggested that folding is replaced downward by thrusting in the brittle carbonates of the lower Devonian, but the data to support this suggestion are sparse and ambiguous. Structural contours on the Oriskany sand have been interpreted by Brayton Foster (pers. comm., 1986) as delineating sets of reverse faults, often with opposing dips, beneath the crests of the regional anticlines. Such opposing faults would define tepee structures and pop-ups, which are common in salt-base thrust systems, and would explain the general symmetry of the overlying anticlines.

The Firtree Point (Portland Point) anticline, which is well expressed in exposures along the shores of Lake Cayuga, is the best studied of these folds, largely because of the salt mine in its core. The anticline trends N80E in surface exposures and is grossly symmetric, with flank dips of less than 2° (Prucha, 1968). However, dips on the south limb (in the Cayuga Crushed Stone Quarry) reach 6 to 8°S, and are probably associated with local thrusting (Scott, 1986). Structural relief on the Tully limestone is about 75 m (Prucha, 1968). The mine workings and borings show that the fold in the Syracuse formation is similar in trend, and more importantly, in amplitude as expressed by the fold in surface exposures. There is, however, evidence of a fault in the Oriskany sand beneath the anticlinal crest (Brayton Foster, pers. comm. 1986), which appears to account for some of the shortening within the Early Devonian strata. No relief exists on units beneath the salt. These observations together indicate that the Firtree anticline is not simply an upward continuation of a blind thrust, but a faulted, salt-cored fold. Minor folds within the salt are much tighter and more irregular than the major anticline, and curiously trend about N50W (Fig. 2 of Prucha, 1968).

Faults

Only a very few faults with displacements greater than a meter have been recognized in the Cayuga region, most of which are south dipping thrusts. The recognition of most of these faults in artificial outcrops suggest that there are probably other similar faults in the middle Devonian strata that have not been recognized.

The largest and best documented exposed fault in the region is a thrust high on the south flank of the Firtree anticline, exposed in South Lansing in the Cayuga Crushed Stone Quarry and on adjacent valley walls. This thrust locally trends N84°E, subparallel to the axis of the anticline, and has a southerly dip of 20° to 30° (Scott, 1986). The dip steepens where the fault crosses the Tully limestones, generating a small ramp-flat geometry and related hanging wall structures (small folds). The total displacement on this purely dip-slip feature is 65-70 feet. The extension of this fault westward across Lake Cayuga, noted in New York State Geological Association Guide Book, 31 (1959), has not been subsequently substantiated.

Other thrusts in the region occur in the band of carbonates between the north end of the Finger Lakes and Syracuse (Chute, 1964; Wm. Brice, 1986 pers. comm.). These are all north vergent faults with moderate to steep dips and displacements of up to 50 feet. Faulting may either be more common in these strata (Engelder and Geiser, 1980) or may be concentrated where the deforming sheet of strata overlies the original margin of the salt and resistance along the decollement increases. Thrusting within the Salina group is documented by wells in the Watkins Glen Brine Field by Jacoby and Dellwig (1974, p. 231), who feel these thrusts do not penetrate the overlying section.

Other Deformational Features

Folding and faulting account for less than 1% horizontal strain in the strata above the decollement in the Cayuga region (Engelder, 1979a). Far more strain (up to 15%) is recorded by distributed layer parallel shortening, expressed as ductile intergranular flow, intragranular flow, and pressure solution (Engelder, 1979b). Intergranular flow and rotation of clay platelets is suspected to have accompanied an early soft sediment deformation (Engelder, 1985), and intragranular flow is documented by calcite twinning in calcareous fossils and in limestone units (Scott, 1986; Engelder, 1979a). Pressure solution occurred both as stylolitic seams in limestone and by a faint pressure solution cleavage in siliciclastics (Geiser and Engelder, 1983). All this deformation appears to be Alleghenian in age, but Engelder and Geiser (1983) provided evidence to suggest that an early phase, responsible for fabric orientation and for one joint set had a maximum shortening direction that was oriented slightly west of north in the Cayuga region, whereas a later phase shortening was more northerly directed. Shortening, but of a much reduced magnitude was also noted in the strata beneath the decollement (Engelder, 1979b, Engelder and Geiser, 1979).

Jointing and Stress Studies

The very regular joint sets in the Cayuga region have been noted for years, but new understanding has been provided by the studies of Engelder and colleagues. Four regional joint sets are recognized; two cross-strike sets related to separate phases of the Alleghenian orogeny (Sets IA and IB), a strike set that is a post-Alleghenian "release" set (Set II), and an oblique set that reflects the contemporary stress field (Set III) (Engelder, 1985). The contemporary maximum principle stress, which trends ENE in New York State has been established by seismic focal mechanisms

(e.g., Yong and Aggarwal, 1981), by hydraulic fracturing in wells (Zoback and Zoback, 1980), and by analysis of well "break-out" geometries (Hickman et al., 1985). Measurements of residual stress in shallow or exposed strata by such techniques as overcoring show that the elastic stresses impressed during the Alleghenian orogeny remain in the rocks (Engelder, 1982).

CONCLUSIONS

The geology of central New York has been studied for over a century and has been the setting for several seminal geological concepts. It was here that Hall (1959) developed the idea of a geosyncline and here also that Williams (1894) worked out the modern concept of sedimentary facies. With the more recent bias of studies toward the geology of plate boundaries, however, many of us have considered the Cayuga region to be so simple as to warrant little further attention.

The short sightedness of such an attitude has been illustrated by the studies of jointing by Terry Engelder, who has pointed out that the mild deformation of the region affords a unique opportunity to understand deformation mechanisms and the state of stress. With the availability of new tools and techniques, several other regional problems might usefully be addressed. The origin of the kimberlite dikes, which are relatively abundant here, still begs an acceptable explanation. The Silurian salt basin remains enigmatic, and even the contentious nature of the Acadian orogeny might be constrained by regional sedimentary petrographic studies.

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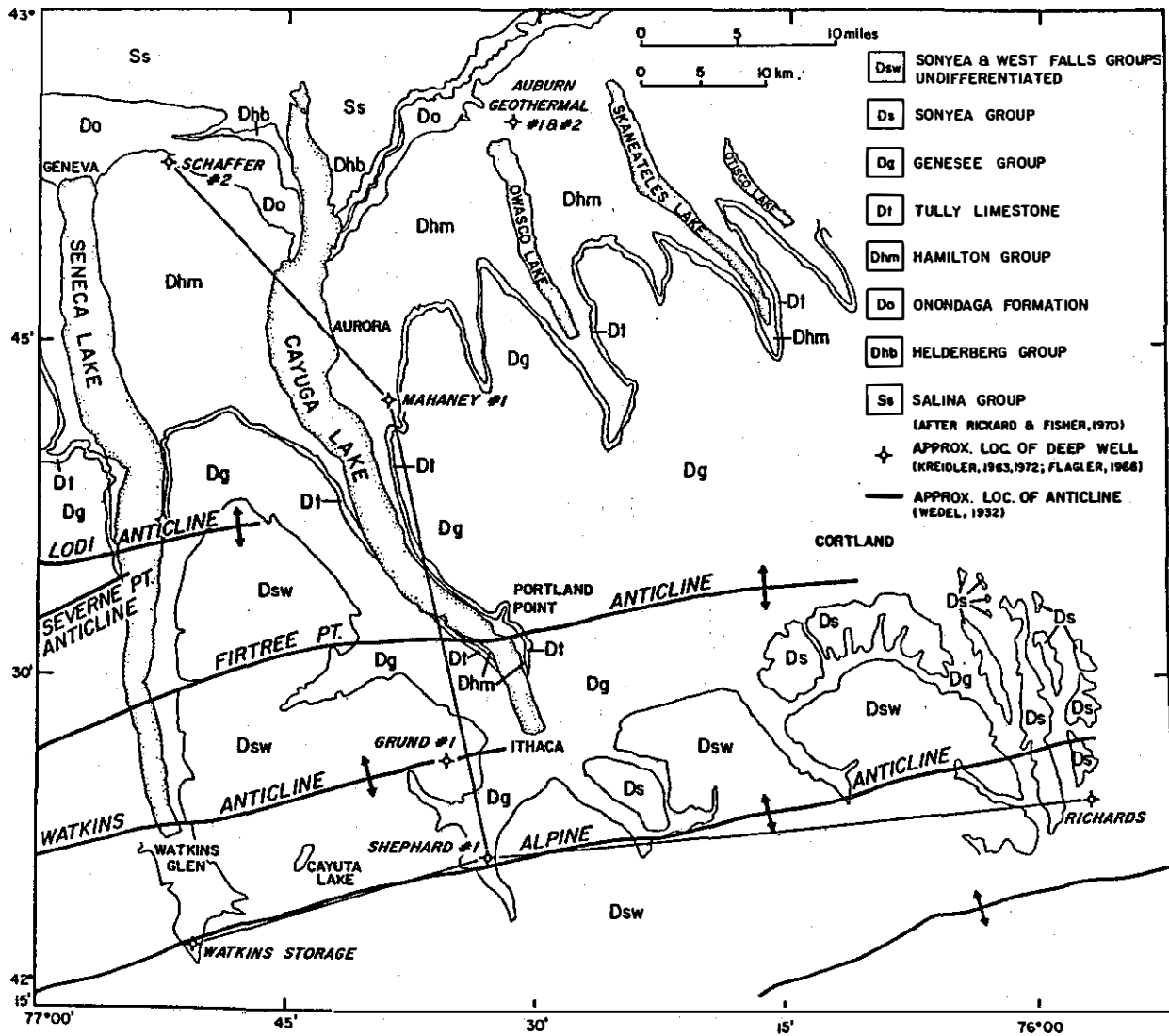


Figure 1: Generalized geologic map of the Cayuga region after Rickard and Fisher, 1970, showing locations of deep wells and anticlines. Sections presented in Figures 3 and 4.

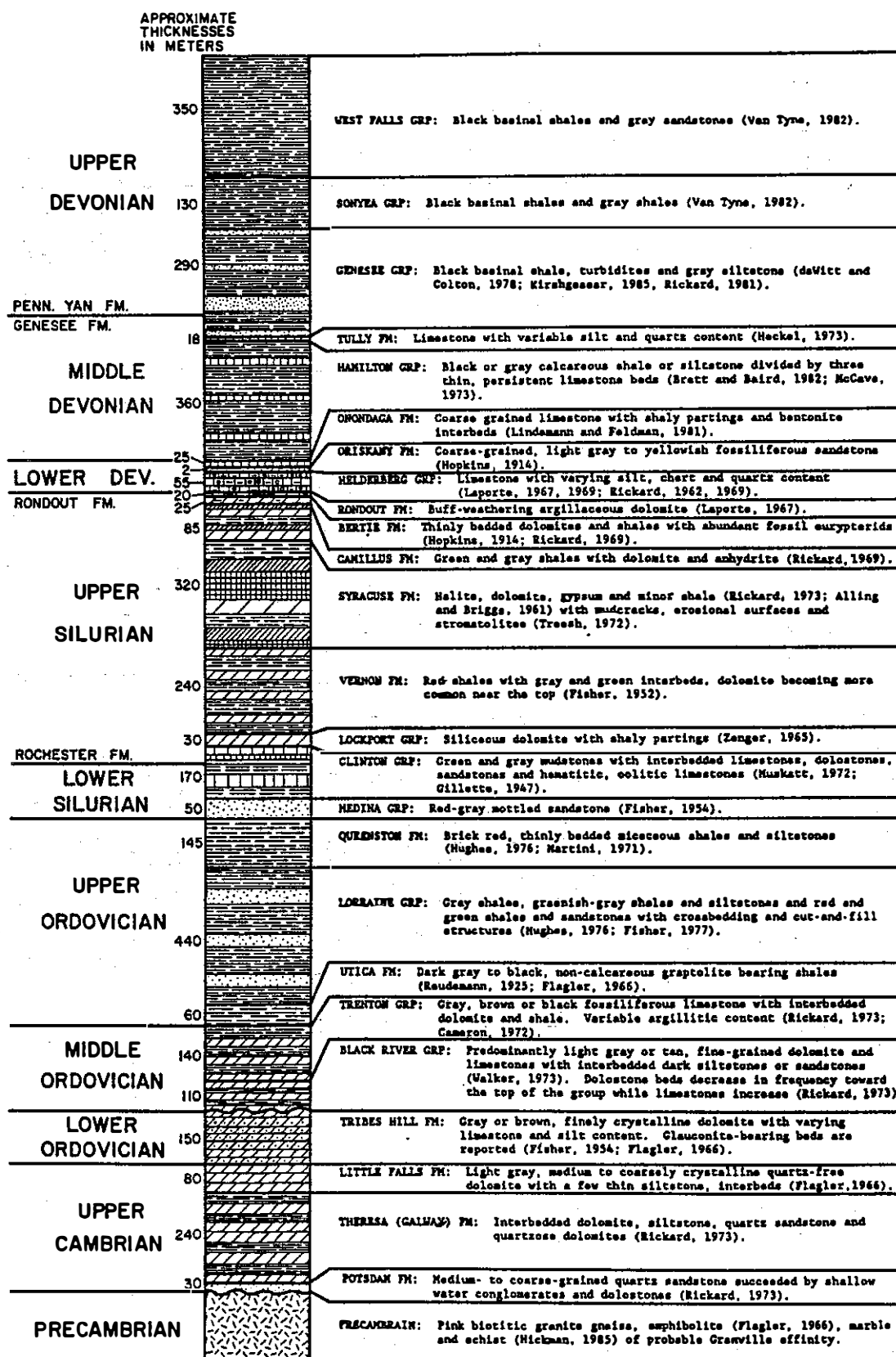


Figure 2: Stratigraphic column near Ithaca, New York. After Committee on Stratigraphic Correlations, 1966, and Rickard, 1969, 1973, 1981.

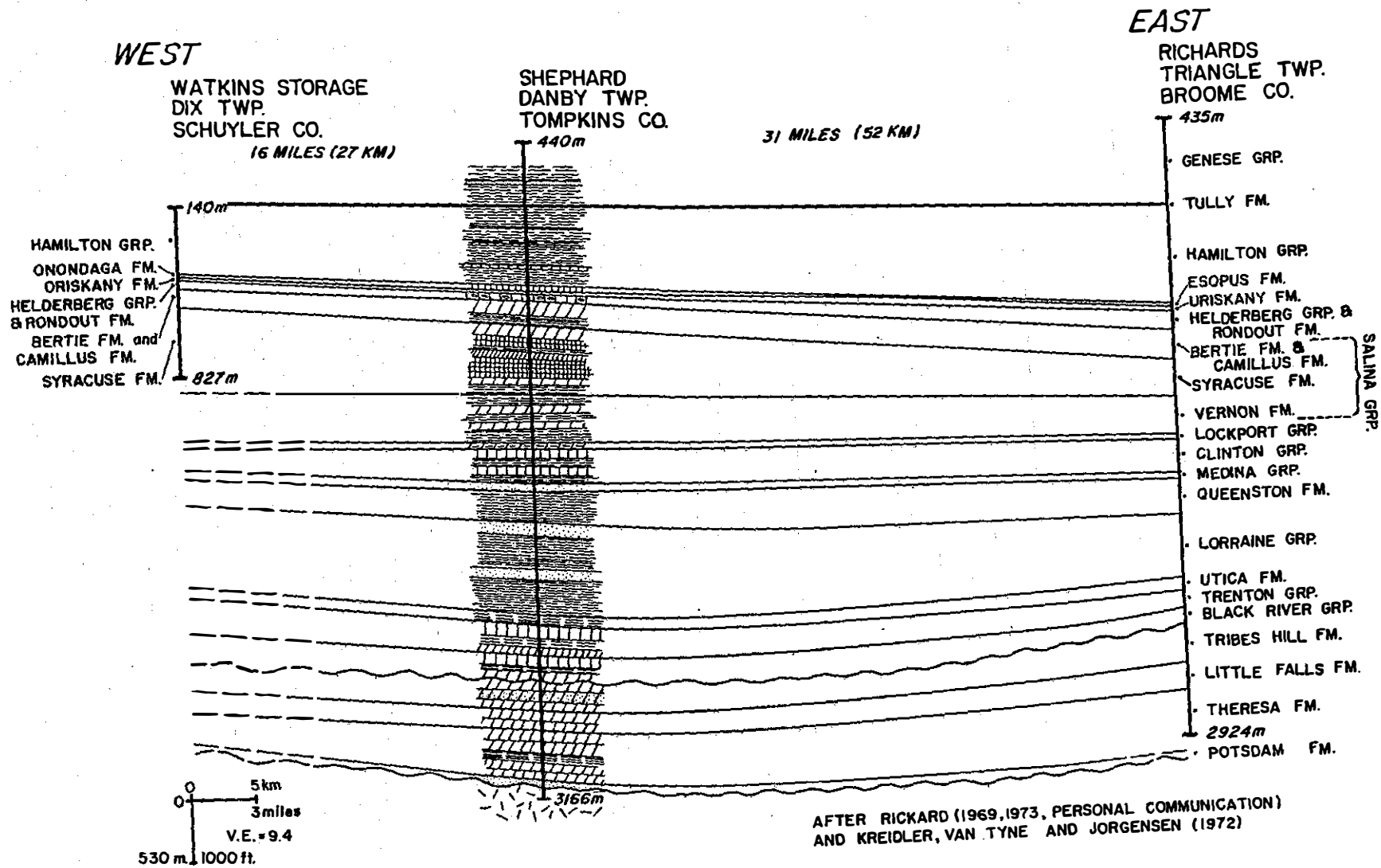


Figure 3: East-west stratigraphic cross-section just south of Ithaca, New York. After Rickard, 1969, 1973, personal communication, and Kreidler et al., 1972.

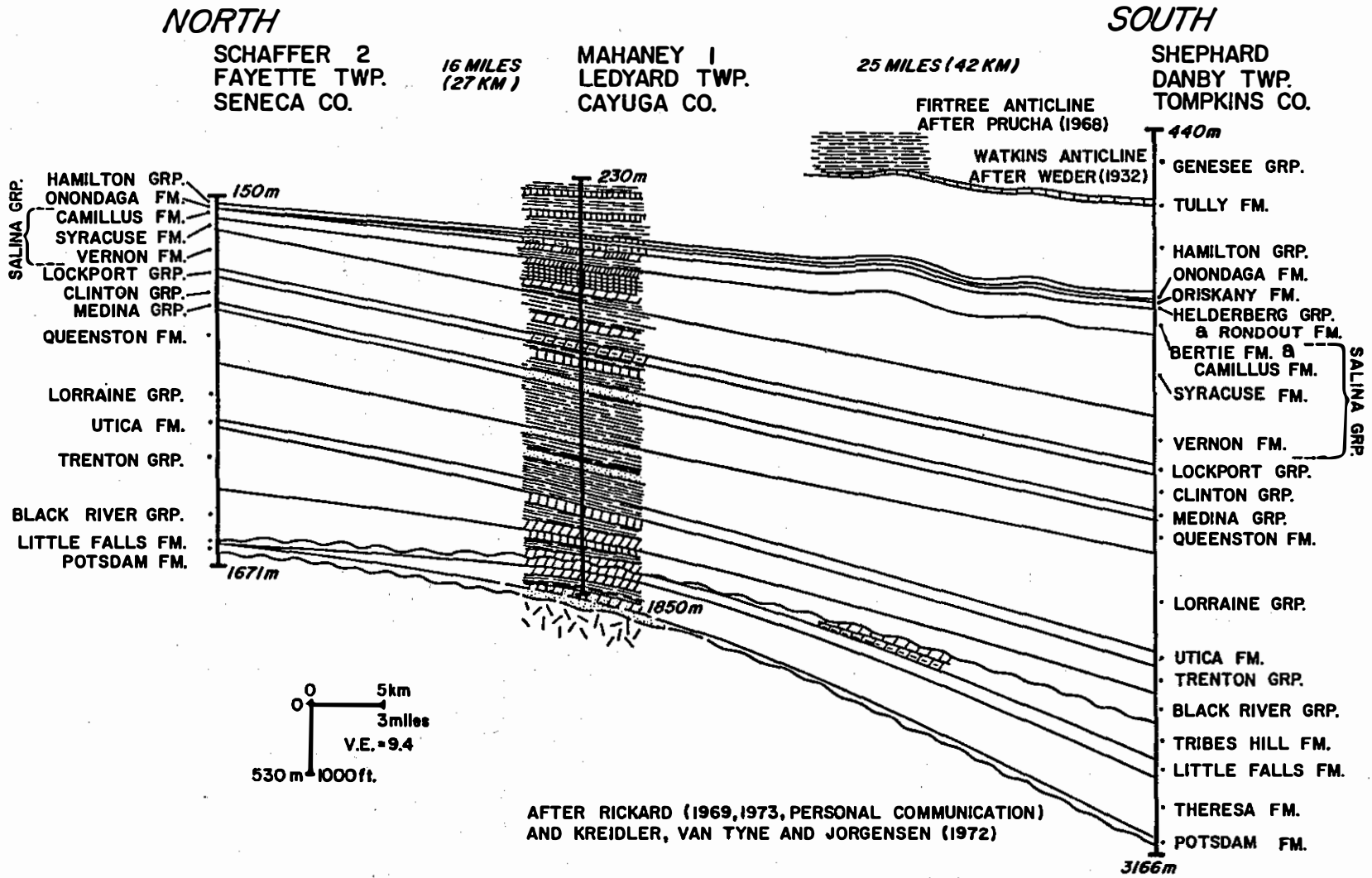


Figure 4: North-south stratigraphic cross section along Cayuga Lake. After Rickard, 1969, 1973, personal communication, and Kreidler et al., 1972.

FIG. 5a

Approximate structure contours
on top of Precambrian basement
after Rickard (1973)
in feet below sea level

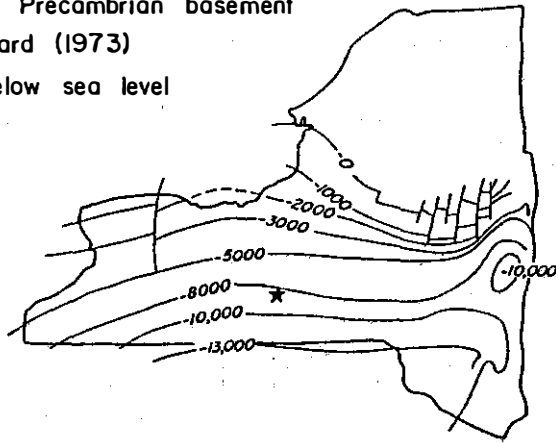


FIG. 5b

Approximate isopach map of
Cambrian and Lower Ordovician
(all units below the Knox
unconformity) in feet, restored
over the Adirondack massif
after Rickard (1973)

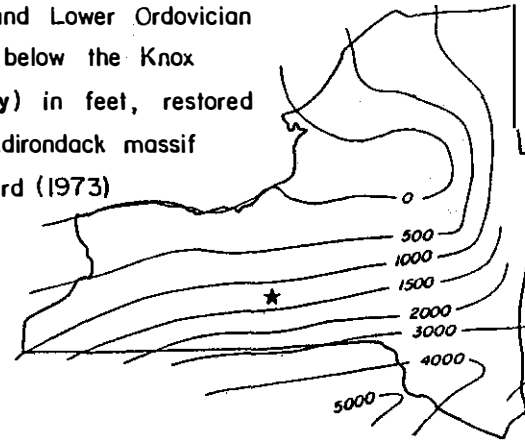


FIG. 5c

Approximate isopach map of
the Salina group in feet
after Rickard (1969)

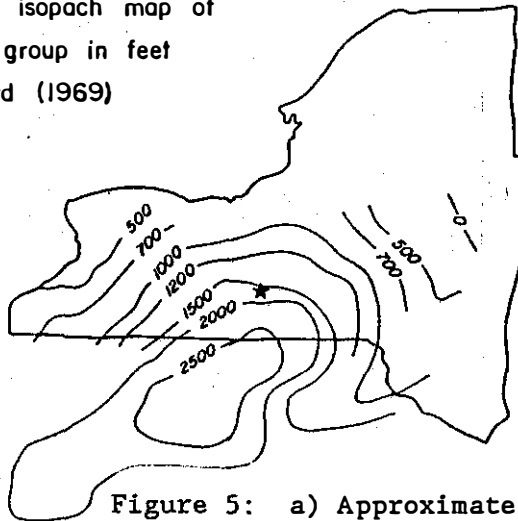


FIG. 5d

Thickness of Genesee in feet
after de Witt and Colton (1978)

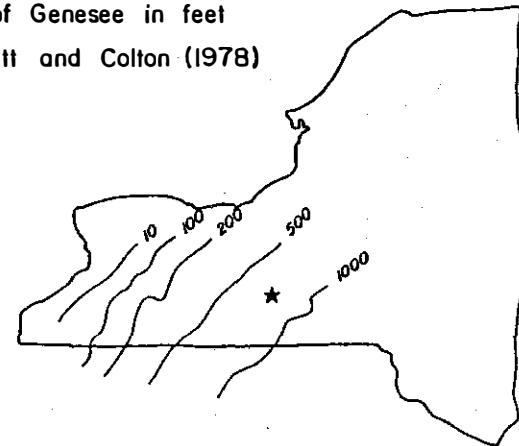


Figure 5: a) Approximate structure contours on top of Precambrian basement in feet below sea level. After Rickard, 1973.

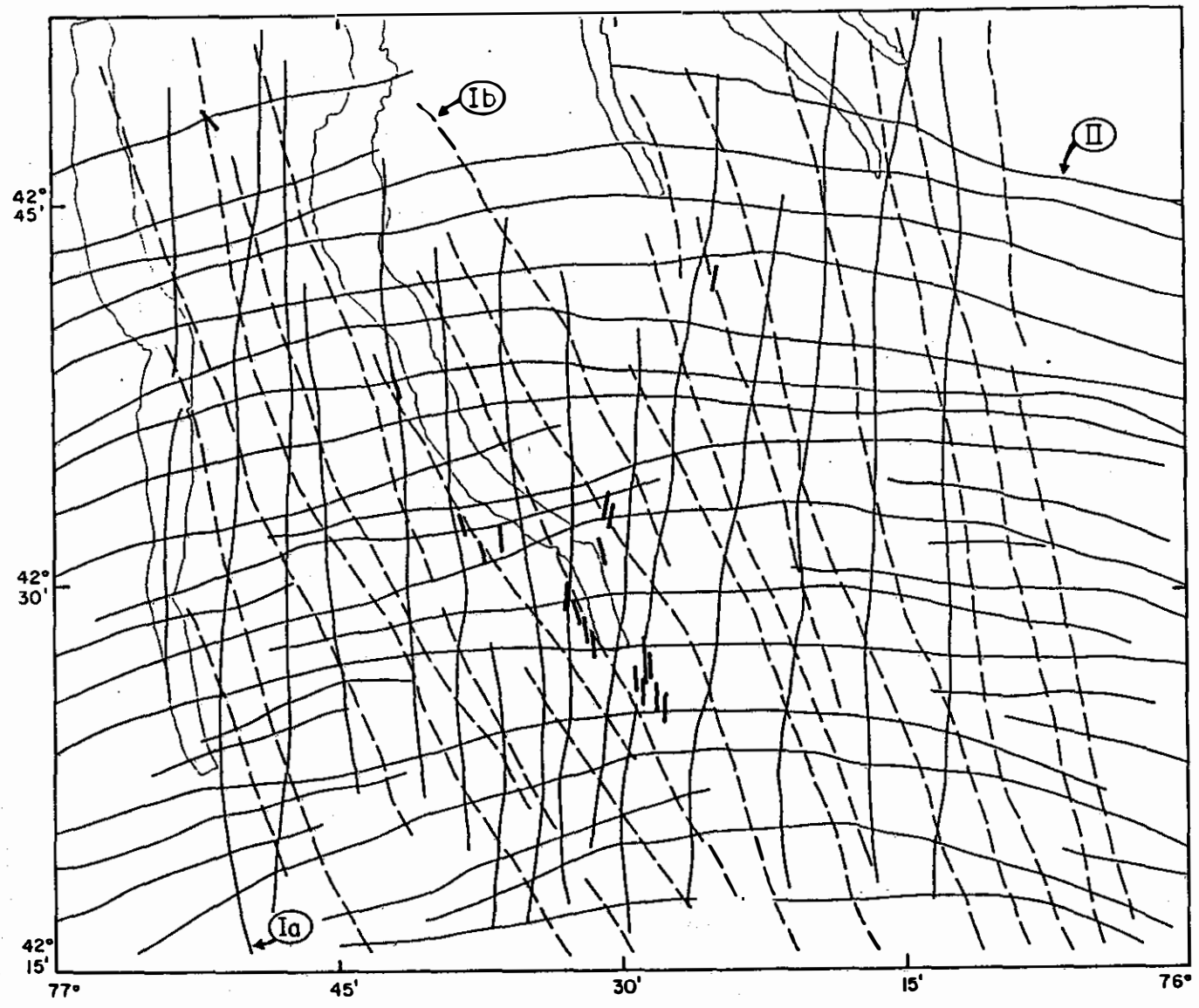


Figure 6: Trajectories of the joint sets I and II (After Engelder and Gieser, 1980) and locations of kimberlite dikes (Foster, 1970) in the Cayuga region.

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SUBMARINE EROSION ON THE DYSAEROBIC SEAFLOOR:
MIDDLE DEVONIAN CORRASIONAL DISCONFORMITIES IN THE CAYUGA VALLEY REGION

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INTRODUCTION

Many Devonian black shale units in New York and adjoining states are marked by erosional unconformities along their bases. Of particular note are distinctive lag accumulations of detrital pyrite, phosphatic debris, and bones along these discontinuity surfaces. This debris is derived from the scour of underlying units and it is often concentrated in erosional runnel-like depressions which explains the typical occurrence of this material as laterally discontinuous lenses in outcrops.

This reworked material is remarkable in that pre-formed pyrite was reworked as pyrite gravel; several lines of evidence in support of this include: similarity of reworked grains to underlying in-situ pyrite, distinctive patterns of pyrite grain breakage within lenses, and the reorientation of geopetal stalactitic pyrite. The absence of detrital carbonate allochems within debris lenses and the facies-specific association of these lenses with black shale deposits, indicate: 1, that detrital pyrite was stable on the anoxic or minimally dysaerobic sea floor; 2, that carbonate material (loose allochems on exposed limestone substrates) was prone to dissolution in these oxygen-deficient settings; and, 3, that strong episodic current activity was important in scouring the sea bed within the basin environment.

Where carbonate dissolution accompanies mechanical abrasion of the sea bed, we refer to the resultant erosion surface as a corrasional discontinuity; although this word has been used to describe several unrelated erosion processes, it has been introduced previously to describe submarine erosion by a combined process of abrasion and corrosion (see Gary et al. eds., Glossary of Geology, 1972).

We believe that discontinuities associated with black shale deposits, particularly those in the Genesee Formation, were produced by current processes operating in circumlittoral outer shelf and basin settings. Current processes probably include a significant component of deep-storm-generated turbulence, particularly on the mid- and upper basin margin slope. However, the occurrence of scour contacts and coarse debris along bases of even thin black shale deposits suggests that erosion may be

closely associated with impingement of the water mass boundary (pycnocline) with the basin margin slope. We include a discussion of the possible role of internal waves in generating these peculiar contacts and deposits during transgression events.

Black Shale Environments

A persistent myth regarding black shales is that they were deposited extremely slowly in "stagnant" settings by the gradual sedimentation of suspended sediment. Recent studies of black shale units indicate that basinal environments recorded in these deposits were more dynamic with respect to bottom circulation than the older "silled basin" model would imply. Williams and Richards (1984) note abundant evidence of current-induced graptolite alignment, as well as entrainment of silt in ripple bedforms in Ordovician black shales, a phenomenon which is also beautifully displayed in the medial Clinton Williamson Shale, an analogous Middle Silurian black shale in western New York. Likewise, Brenner and Seilacher (1978) and Kauffman (1981) cite evidence for moderate to strong (> 20 cm/sec) currents during deposition of the Jurassic Posidonia shales. Similarly, Devonian black shales in western New York display numerous levels of current-aligned, conical Styliolina shells, as well as thin beds of siltstone which contain bedforms suggestive of pervasive winnowing by bottom currents on a sediment-starved sea floor rather than turbidite deposition per se (Baird and Brett, in press). The presence of erosional unconformities and associated reworked pyrite gravel in this facies further indicates that strong currents were periodically active in this environment.

Most workers, including the present authors, interpret the Devonian black shales as offshore facies recording deposition in an intracratonic basinal setting; depth estimates for Frasnian and Famennian black shales range up to 530 meters (Lundegard, et al. 1980) but several authors favor depth ranges of 50 to 230 meters (Thayer, 1974; Broadhead, et al. 1982; Woodrow and Isley, 1983). The eastward facies spectrum from black shale, through greatly thickening turbidite wedges, into shell-rich, aerobic platform facies of the Catskill Delta Complex favors a deeper water (circumlittoral) interpretation for the black shale deposits (Thayer, 1974). Westward thinning of the Genesee deposits from 400 meters at Ithaca to 3 meters in eastern Erie County (deWitt and Colton, 1978) also indicates that black shale facies in western New York record extremely slow net deposition. The Devonian black shale sea was presumably stratified; this stratification may have produced a "silled basin" effect (e.g. model of Byers, 1977) with a strictly horizontal layering of water zones, but we feel that it may have been more a sub-horizontal, vertically-shifting pycnocline as proposed by Ettensohn and Elam (1985).

Discontinuities Flooring Black Shale Units

Discontinuities flooring black shales in New York State can be observed in the Middle Ordovician (Trenton Limestone-Utica Shale contact at several localities), the Middle Silurian Clinton Group (Sodus Shale-Williamson Shale contact in the Rochester region), and at numerous levels in

the Middle and Upper Devonian (Eifelian and Givetian) examples include the base of the Bakoven Shale Member in eastern New York (Goldring, 1943), the base of the Oatka Creek Member in western New York (this paper), the base of a black shale sequence in the medial Levanna Member in western New York, the base of the Ledyard Shale Member in the Seneca-Cayuga Valley (Brett and Baird, 1985; this paper), the base of the Genesee Formation (Taghanic Unconformity: Fig. 1) in western New York (see Brett and Baird, 1982; Baird and Brett, in press, numerous other authors), and at the base of thin black shales in the lower Genesee Formation in the Seneca Valley-Salmon Creek Valley region (Baird and Brett, 1986; this paper). Upper Devonian (Famennian) examples include the base of the Dunkirk and Cleveland black shales in western New York and Ohio, respectively (see Mausser, 1982; Baird and Brett, in press). The present field trip discussion and road log is an outgrowth of detailed studies of the Leicester Pyrite Member and two stratigraphically higher detrital pyrite occurrences within the lower part of the Genesee Formation.

Significance of the Leicester Pyrite

The Leicester project involved the study of lenses of pyritic debris (Leicester Pyrite Member) on the post-Hamilton-post-Tully (Taghanic) disconformity which separates richly-fossiliferous shales of the Upper Hamilton Windom Member (Moscow Formation) or Tully carbonates from overlying laminar black shale of the Genesee Member (lower Genesee Formation; see Fig. 1).

This erosion surface marks a hiatus which increases in magnitude westward from the Skaneateles meridian, where it is a mere omission horizon, to Lake Erie, where it is a regional disconformity below which the Tully Limestone and the upper beds of the underlying Windom Member of the Hamilton Group have been removed (Huddle, 1981; Brett and Baird, 1982). This unconformity is associated with a major eustatic transgression which began in the latest Givetian; it marks the Taghanic Onlap event in which basinal black shale deposits accumulated on variably bevelled older units (Johnson, 1970; Huddle, 1981; Johnson et al., 1985). The age of the basinal shales above the unconformity decreases westward from the Cayuga Valley to Lake Erie (Fig. 1); these beds become markedly younger into Ohio and Ontario such that up to four million years of diachroneity is revealed by conodont biostratigraphy due to regional onlap effect (Huddle, 1981; Uyeno, et al. 1982). Detailed mapping of strata within the lower Genesee Formation in western New York State corroborates biostratigraphic studies; individual beds are observed to converge and descend onto the Taghanic erosion surface as they are followed westward (Baird and Brett, 1985).

Findings from the Leicester study include the following: 1, most Leicester grains are reworked detrital pyrite; 2, Leicester pyrite lenses were deposited within the predominantly anaerobic environment recorded by the black shale; lenses are shingled within the basal few centimeters of this shale recording anywhere from one to several pyrite transport events in any given locality; 3, there is stratigraphic evidence for regional diachroneity of Leicester lenses, and, by implication, diachroneity of

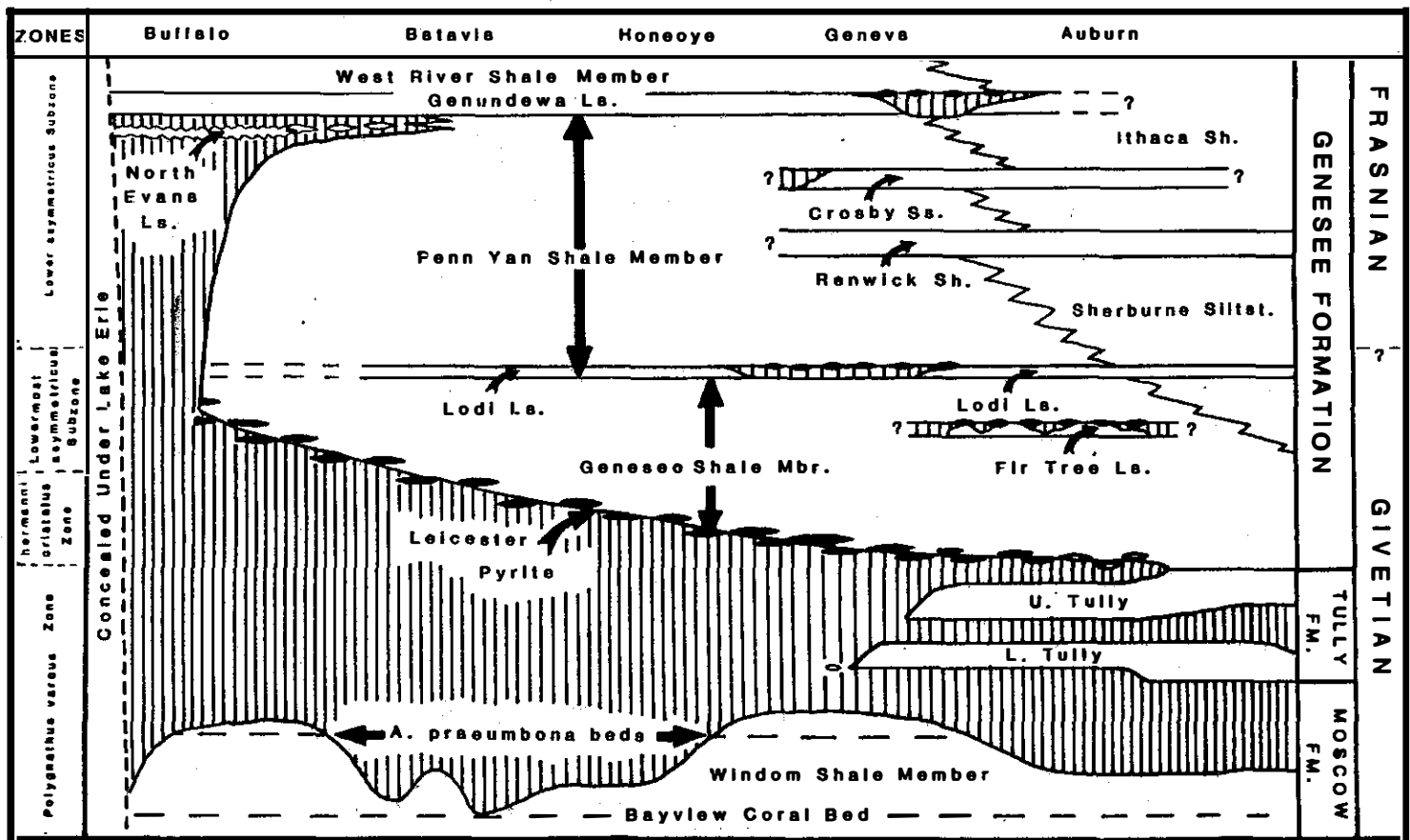


FIGURE 1.--Chronostratigraphic cross-section of lower Genesee Formation and subjacent Moscow Formation (Windom Shale Member). Note positions of the thin, and locally bevelled Fir Tree and Lodi limestone submembers. Large hiatus below the Genesee Formation marks the position of the Taghanic Unconformity; lenses of detrital Leicester Pyrite are derived from this erosion but were deposited through a long period of diachronous overlap of Genesee black muds upon this discontinuity. Based on Brett and Baird, 1982; Baird and Brett, in press, this paper.

basal Genesee muds, involved in the Taghanic Onlap event; this agrees with Huddle's (1981), biostratigraphic conclusions, summarized above (Fig. 1); and finally, 4, that the erosional process which entrained the pyrite and bones on the erosion surface was clearly an ongoing event both up to-, and even synchronous with, initial depositional onlap of Genesee black muds (Brett and Baird, 1982; Baird and Brett, 1985, in press).

Reworking of Pyrite Grains: Lines of Evidence

Examination of pyritic material within Leicester lenses and in the higher Genesee examples examined on this field trip, shows that the grains are essentially a detrital pyritic gravel breccia that was entrained and aligned by bottom currents (Fig. 2). Pyritic grains include nodules, tubular particles, and fossil steinkerns. Evidence for pyrite exhumation is fourfold. Firstly, Leicester tubular grain fragments correspond closely to in-situ pyritic burrow tubes in the underlying Windom Member. The absence of comparable pyritized lebensspuren in the Genesee black shale indicates that Leicester tubes are derived from below. Moreover, we have observed partially exhumed (upright) tubes both along the Taghanic unconformity and above two similar but younger discontinuities (post-Fir Tree and post-Lodi diastems discussed here) within the lower Genesee Formation (Fig. 2). The in-situ Windom tubes, corresponding closely to pyritized polychaete burrows described from Late Pleistocene Atlantic sediments (Thompson and Vorren, 1984), show evidence of early diagenetic pyritization in near-surface muds; Leicester tube debris was certainly pyritic at the time of erosion, up to approximately 500,000 years after its formation in Windom sediment, based on the zonal magnitude of the post-Windom hiatus.

Other types of pyritic grains in the Leicester also closely resemble those found in situ in the underlying shales. For example, reoriented, compressed, pyritic steinkerns of nautiloids in the Leicester correspond to in situ, vertically-flattened pyritic molds in the subjacent Windom (Baird and Brett, in press); in order for such molds to have been reworked, they would have to have been reinforced by diagenetic mineralization before exhumation. Leicester brachiopod and bivalve steinkerns additionally display "meniscus" collars of coarse crystalline exterior pyrite ("overpyrite") identical to those found in pyritized shells from the underlying Windom Shale. Regional species correspondence of Leicester steinkern fossils to those in successively bevelled Windom strata supports our contention that Windom-derived pyrite constitutes the bulk of Leicester grains (Brett and Baird, 1982; Baird and Brett, in press).

Secondly, pyritized steinkerns of tubes and fossils show arcuate, sharp-edged mechanical breakage surfaces, indicating that the grains were hard and brittle when broken.

Thirdly, to close any argument that Leicester grains may have been diagenetically altered to pyrite from a different earlier mineral, we document reorientation of early diagenetic geopetal pyritic stalactites identical to those described by Hudson (1982). We observe vertical

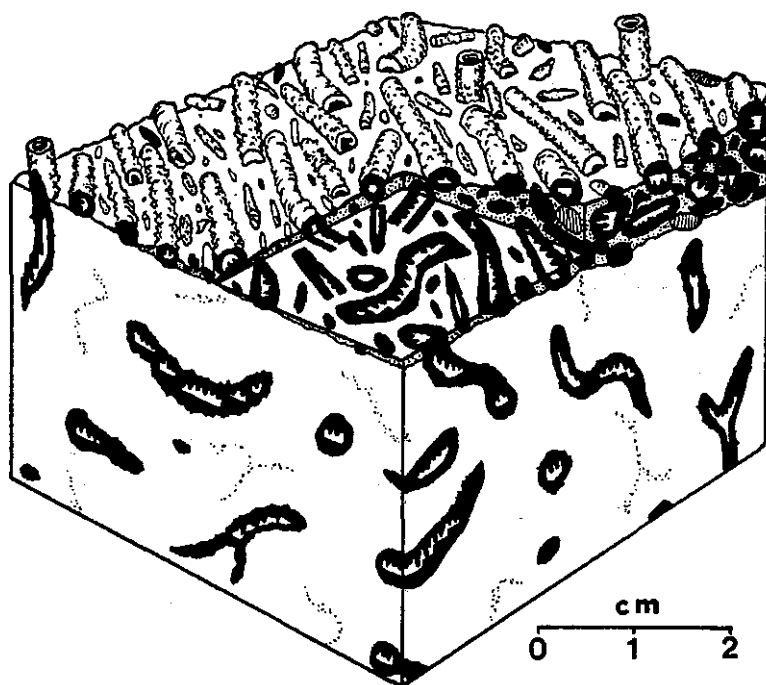


FIGURE 2.--Reorientation of pyritic geopetal structures on top Lodi discontinuity: schematic reconstruction. Note random attitudes of reworked burrow tubes above discontinuity.

stalactites within interior axial voids of in-situ pyritic burrow tubes below both the Leicester and a younger reworked pyrite accumulation associated with the Lodi Limestone Bed, higher in the Genesee Formation (this paper); tube fragments in the overlying lenses show stalactites in random orientations (Fig. 2).

Dissolution of Calcareous Grains

Pyritic lenses, conversely, generally show complete absence of calcareous grains associated with the pyrite although fish bones, phosphatic orbiculoid fragments, and conodont debris are typically abundant. Although the Leicester pyrite regionally oversteps the Tully Limestone as well as numerous underlying Windom concretion and shell beds, calcareous clasts are rare in lenses (Brett and Baird, 1982).

We hold that Leicester calcareous debris material underwent dissolution on the oxygen-deficient sea floor following exhumation. Where we observe occasional calcareous grains in lenses, they are almost always badly corroded. Examination of debris on the younger (basal Genesee) North Evans discontinuity (medial Genesee Formation; Fig. 1)) adds further support to our contention; where reworked debris is overlain by a limestone (Genesee Member) it is calcareous, but where a black shale

unit intervenes between the debris layer and the limestone, the reworked material changes laterally from a continuous blanket of pelmatozoan and hiatus-concretion debris to a thin, discontinuous deposit, rich in both nodular and tubular pyrite. This pattern suggests that the pyritic beds are chemical residues of a far greater volume of reworked material. Thus, Leicester, and analogous pyritic deposits are the chemical reverse of normal detrital carbonate accumulations; in aerobic settings, carbonate is stable but pyrite will oxidize and disintegrate. In the basinal settings discussed here, however, the pyrite is stable, and it is the carbonate fraction which is lost.

The precise process of dissolution remains to be determined. We believe that prolonged exposure of carbonate on the basin slope and floor combined with periodic development of low pH conditions are responsible for its disappearance. Similar interpretations have been made for other instances of fossil dissolution in black shale sequences (Seilacher, et al. 1976; Tanabe, et al. 1984). Reaves (1984) argues that repeated oxidation of pyrite or iron monosulfide produces sulfuric acid leading to dissolution of adjacent carbonate, and Sholkovitz (1973) interestingly shows that shells exposed on the anaerobic sea floor are better preserved than those found on dysaerobic substrates. Locally observed multiple-stacking (shingling) of Leicester lenses within basal Genesee black muds indicates that episodic strong currents swept the basin substrate; this suggests that prevailing conditions were anoxic but that brief periods of oxygen incursion could have lowered the pH level near the reworked materials.

Geologic Units to be Discussed

Three important and distinctive examples of submarine discontinuities associated with the onset of black mud deposition will be examined on this field trip; these examples, discussed in sequential-, and stratigraphically ascending, order, include: 1) the Cherry Valley Limestone-Oatka Creek Shale contact near the base of the Hamilton Group (corrasional limestone hardground); 2) the Centerfield Limestone-Ledyard Shale contact in the medial Hamilton Group (erosional concentration of phosphate nodules, shells, and detrital pyrite); and, lastly, 3) we describe and discuss the observed downslope erosional bevelling of two thin limestone units (Fir Tree and Lodi beds) in the lower Genesee Formation by discontinuities at the base of black shale units above each limestone.

On the field trip, the first four stops will illustrate both the downslope condensation of the Fir Tree Bed and the eventual downslope erosional overstep of this same unit below a black shale sequence; at stops 2 and 3 participants will also be able to observe and collect detrital pyrite on the discontinuity surface.

GEOLOGIC SETTING

Paleogeography and Tectonic Setting

The Hamilton Group and Genesee Formation are parts of a thick and largely terrigenous sequence within the Appalachian basin of eastern North

America. These sediments accumulated at the northern margin of this basin a few degrees south of the inferred paleoequator, based on paleomagnetic reconstructions (Ettensohn, 1985). The Acadian orogeny commenced during the Middle Devonian; convergent activity east and southeast of New York State involving oblique collision of one or more microcontinents (Avalonia) or possibly an island arc produced mountains in New England and in the Mid-Atlantic states (Dewey and Burke, 1974; Van der Voo et al., 1979).

Initial Acadian disturbance in New York is marked by an upward change from carbonate deposits (Onondaga Limestone) to the predominantly terrigenous succession of Hamilton formations; both the Hamilton and stratigraphically higher Genesee sediment sequence record erosion of the rising mountains to the east (Cooper, 1957; Rickard, 1981; Woodrow, 1985; Ettensohn, 1985).

The Genesee Formation and succeeding clastic divisions of the Upper Devonian record particularly rapid growth of the Catskill Delta Complex, a large tectonic delta system bounded by a sublittoral to uppermost bathyal basin to the west and northwest (Sutton, Bowen, McAlester, 1970; Broadhead et al., 1982; Woodrow, 1985); Genesee deposition coincides with a major pulse of orogenic activity which caused crustal diastrophism and locally high rates of sedimentation in the study area (Dennison and Head, 1975; Johnson et al., 1985; Ettensohn, 1985). The deepest part of the basin during Hamilton and Genesee time was probably centered in western Pennsylvania south of the study area (McIver, 1970; Lundegard et al., 1980), but the basin axis apparently shifted westward during the Frasnian and Famennian through processes of tectonic adjustment and isostatic effects (Ettensohn, 1985). The northern and western boundaries of the basin bordered low relief, cratonic shelf regions which supplied relatively little detrital sediment on that side of the basin; this is reflected in the almost exponential westward thinning of Genesee and higher sedimentary units to the west and northwest across the basin.

Depositional Setting and Facies

The present Finger Lakes Region is near the northern limit of the Devonian basin; during much of Hamilton time it was a variably well-oxygenated, infralittoral shelf which, at numerous different times, supported some of the highest Devonian invertebrate diversity recorded in eastern North America. Basal Hamilton (Marcellus) strata in western New York (see Union Springs and Oatka Creek Shale members: stop 5) are predominantly anaerobic and dysaerobic facies supporting minimal benthos, but numerous middle and upper Hamilton beds (see Centerfield Member: stop 6) are rich in invertebrate macrofossils, including rugose and tabulate corals, brachiopods, bryozoans, trilobites, bivalves and pelmatozoans; facies of this type record shallow subtidal shelf conditions with reduced turbidity (Baird and Brett, 1983; Gray, 1984; Brett et al., this volume).

During deposition of the Centerfield and most of the higher Ludlowville Formation, the New York shelf was bisected by a region of differential subsidence and slightly deeper water conditions (Finger Lakes Trough) which was centered in the Seneca-Cayuga Valley region and which

probably connected southward to more basinal settings in western New York and Pennsylvania, where the largely black Millboro Shale accumulated (Baird and Brett, 1981; Brett and Baird, 1982); conditions in the Finger Lakes Trough explain both the differentially thick and shaley condition of many Hamilton carbonates in this region and the black character of the Ledyard Shale Member at Stop 6, an interval which is grey and more fossiliferous both to the east and west of this area.

After deposition of the upper Hamilton Moscow Formation and succeeding carbonates of the Tully Formation, a period of erosion ensued which removed the Tully Limestone in western New York and up to half of the Windom Shale Member of the Moscow Formation in Genesee County (Brett and Baird, 1982). This pervasive bevelling probably occurred both prior to-, and during the major post-Tully transgression event which saw the development of a deeper, stratified water mass across most of New York State. This period also saw the beginning of the Taghanic sedimentary onlap event, as black, laminated muds (Genesee Member) of the basal Genesee Formation began to accumulate in this basin which was apparently deepest in the Seneca Valley-Canandaigua Valley region. Continued filling of the basin by the prograding Catskill Delta accounts for the overall regressive character of the lower-to-middle Genesee stratigraphic section in the Cayuga Valley.

In contrast to the generally deep-to-shallow infralittoral facies spectrum of the Upper Hamilton Group which includes black or dark grey shales with diminutive fossils on the deep end and carbonate grainstones or shell-rich sandstones on the shallow extreme (see Brett and Baird, 1985 this volume), the Genesee spectrum in the Finger Lakes Region is dominated by less fossiliferous deeper water deposits. Above the Leicester pyrite or uppermost Tully beds, the Genesee Formation commences with hard, laminated, black shale (Genesee Member) followed by a prodeltaic wedge of dysaerobic, turbidite-rich, shale-siltstone facies (Penn Yan and Sherburne Members) which contains thin tongues of black shale and minor shell-rich layers such as the Fir Tree and Lodi beds discussed herein. Only further to the east in the Cortland-Chenango Valley region does the sparsely fossiliferous Sherburne pass into shallow subtidal, coquinite-rich siltstones and sandstones (Unadilla Member) comparable to the more pervasively fossiliferous Hamilton deposits.

CORRASIONAL HARDGROUND SURFACE: CHERRY VALLEY LIMESTONE-
OATKA CREEK SHALE CONTACT

Stratigraphy and Facies

The basal beds of the Hamilton Group which overlie the Onondaga Limestone Formation consist predominantly of black shale with lesser proportions of dark grey styliolinid-rich and concretionary limestone (Fig. 3). Most conspicuous among the limestones is the Cherry Valley Member, a thin (0.5 - 2.0 m thick) unit which is notably widespread across New York State and which is particularly famous for its goniatite and nautiloid fauna (see Flower, 1936; Rickard, 1952; Cottrell, 1972). Between Cherry Valley, New York and the "Five-Points" quarry, northwest of

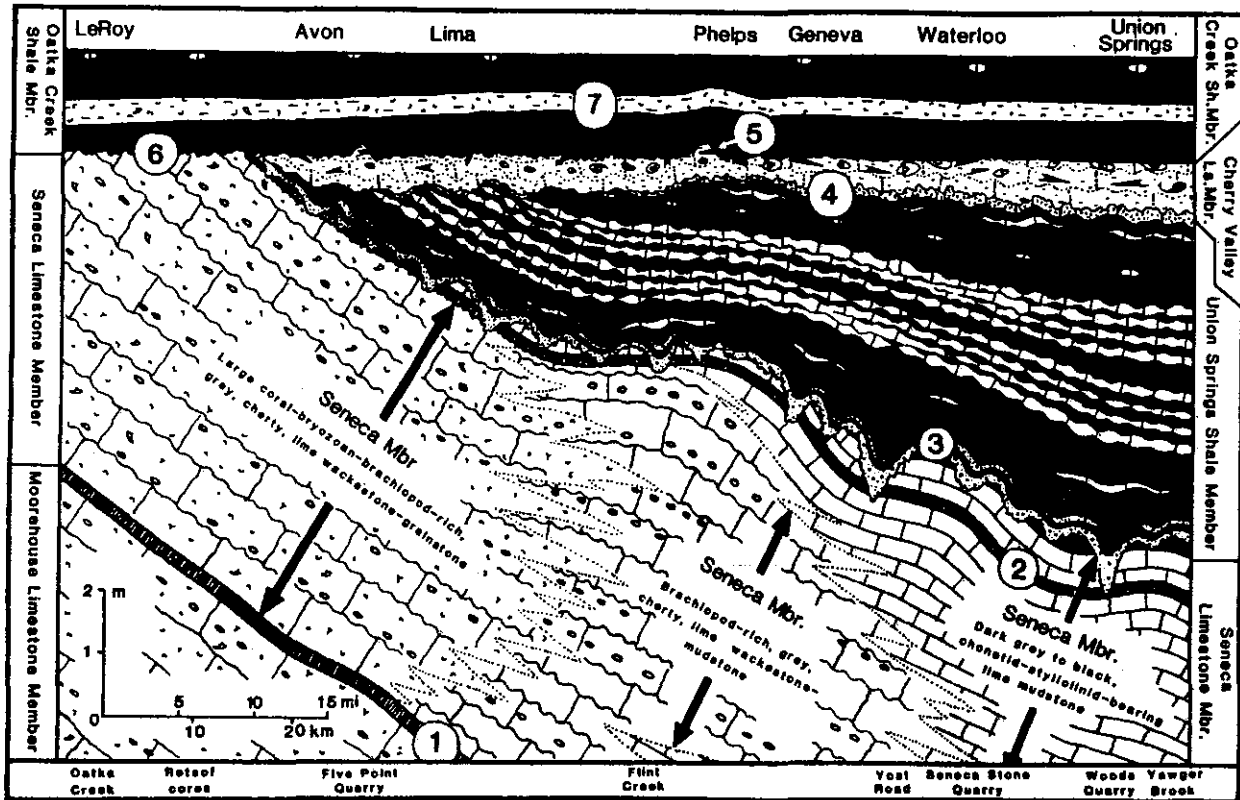


FIGURE 3.--Sub-Oatka Creek corrasion hardground and associated strata. Note conspicuous pattern of westward truncation of Seneca and Union Springs strata below the Cherry Valley and Oatka Creek Members; evidence for bevelling of units below the Oatka Creek is well displayed at Seneca Stone Quarry, Flint Creek near Phelps, Five-Points (Lima) Quarry, International Salt Company (Retsof) drill cores, and on Oatka Creek (see bottom of diagram). In addition, a conspicuous band of thin, bone and onychodid tooth-rich, styliolid limestone beds ("Bone Bed 7" of Conkin and Conkin, 1984), is observed to truncate the uppermost Seneca Member locally; varying completeness of the post-Tioga "A" to top of Seneca section in outcrops between Syracuse and the Five-Points Quarry suggests that the "Bone Bed 7" discontinuity is undulatory, as is shown on figure. Also shown is a conspicuous east to west facies change within the Seneca interval (see text). Units include: 1, Tioga "B" ash (Onondaga Indian Nation Metabentonite of Conkin and Conkin, 1984); 2, Tioga "A" ash ("Restricted Tioga" of Conkin and Conkin, 1984); 3, "Bone Bed 7" of Conkin and Conkin, 1984; 4, "Proetid Limestone Bed" at base of Cherry Valley Member; this unit is partly overstepped by Cherry Valley in this region; 5, corrasional hardground on top of Cherry Valley limestone deposit; raised bosses on this surface at Flint Creek are remnants of higher Cherry Valley deposits which escaped destruction before burial by black muds; 6, knobbly corrasional hardground on partly truncated Seneca Limestone; 7, unnamed, thin, grey, shell-rich bed within basal Oatka Creek Member, which is both widespread and rich in large, distinctive brachiopods.

Valley Member bears some resemblance to the Cephalopodenkalk limestones described from the Devonian of Europe and Africa (see Tucker, 1973; Wendt and Aigner, 1985), and it is similarly rich in pelagic (planktonic and nektonic) organisms, although some benthic taxa (aulopodid corals, bivalves, gastropods) also occur in this unit (Cottrell, 1972). As such, Cherry Valley facies accord well with overlying and underlying black shale sediments; Cephalopodenkalk deposits are generally interpreted as offshore transgressive facies, although not necessarily deep water (Tucker, 1973).

The cephalopod fauna, including large nautiloids and a diverse goniatite fauna, is the best known feature of this unit (see Clarke, 1901; Grabau, 1906; Flower, 1936; Rickard, 1952); one or more levels in the Cherry Valley are locally packed with phragmocones such that certain bedding surfaces are literally pavements of these shells. This is particularly the case for the topmost few centimeters of the Cherry Valley at Seneca Quarry; here, submarine erosion has neatly planed off the top of one of these layers revealing scores of truncated, prefossilized phragmocones at the discontinuity surface.

The juxtaposition of black shale beds on the Cherry Valley upper surface renders implausible the role of boring organisms as the principle erosive agent, as is normally seen on aerobic, organism-coated hardgrounds. Furthermore, the gross character and amplitude of irregularities on the erosion surface both at, and west of Flint Creek, suggest that some process other than burrowing or boring produced these features.

We argue that submarine corrosion of the pre-cemented limestone was an important erosive process prior to Oatka Creek black mud accumulation; current-induced scour was important in exposing the carbonate on the sea floor and in keeping it clear of covering sediment, but the peculiar isolated knobs and irregular, pitted and ridged terrain of the discontinuity surface at Phelps and further west does not resemble the smooth or runneled surfaces of limestones exposed to wave or stream erosion. Rather, it is more like pitted carbonate surfaces observed in areas where dissolution is known to occur (Freeman-Lind and Ryan, 1985; Williams et al. 1985). Additional evidence for carbonate dissolution is indicated by the concentration of phosphatic, but not calcareous, reworked debris on the erosion surface; this pattern is identical to that observed with the Leicester Pyrite, an accumulation of solution-distilled detrital pyrite and phosphatic debris that is similarly associated with initial black mud deposition or a submarine discontinuity surface (Brett and Baird, 1982; Baird and Brett, 1986, in press).

CENTERFIELD-LEDYARD BOUNDARY SURFACE

The Centerfield Limestone Member (basal Givetian) is a widespread key bed that marks the base of the Ludlowville Formation across western and west-central New York State. This unit, which is correlative with the Chenango Sandstone Member in central New York State (Gray, 1984) and probably correlative with the upper part of the Hungry Hollow formation in

Ontario (Landing and Brett, in press). The Centerfield, originally interpreted as being transgressive relative to overlying and underlying shales (see Cooper, 1957), is now known from biofacies and taphonomic evidence to record a regressive maximum in its medial beds (Gray, 1984; Brett and Baird, 1985).

Most significant within the Centerfield is a lithologic and biofacies sequence which is arrayed in a subsymmetrical pattern about the middle carbonate beds of this member (see Fig. 4A); in western New York, the medial beds of the Centerfield Limestone yield large rugose corals, domal Favosites colonies, and crustose stromatoporoids, as well as numerous bryozoans, brachiopods, and echinoderms. Both in the downward and upward directions away from the middle Centerfield limestone beds, there is a sequential outward biofacies transition from the above shallow subtidal fauna through respective intervals of large corals (Heliophyllum, Cystiphyllodes) in calcareous mudstone, in turn, through opposing belts of soft shale rich in small corals, fenestrate bryozoans, and diverse brachiopods, and finally to medium to dark grey dysaerobic shales above and below the Centerfield which yield a meager biota of diminutive brachiopods and mollusks (Gray, 1984).

This same regressive cycle is present in the Cayuga Valley area, except that the Centerfield sequence is much thicker in this region (Fig. 4A, B). It is also, distinctly less calcareous than the Centerfield Member of western New York, and is dominated by a greater number of mud- and silt-tolerant organisms than is observed in western New York Centerfield sections (Brett and Baird, 1985). Otherwise, a similar complete "mirror image" biofacies spectrum is observable both up and down from medial Centerfield calcareous mudstone beds in the Cayuga Valley with the sole exception of the Moonshine Falls section (Stop 6).

The type section of the Ledyard Shale Member, described by Cooper (1930), is on Paines Creek both at, and upstream from, Moonshine Falls; at this locality the lower 10 to 13 meters (33 to 43 feet) is a distinctly dark grey to black fissile shale which yields a sparse fauna consisting of the planktonic organism Styliolina, the rhynchonellid Leiorhynchus, diminutive nuculid bivalves, and flattened orthoconic nautiloids. At Moonshine Falls the Ledyard Member, above the Centerfield, resembles a typical black shale, although it may, in reality, represent a dark grey, dysaerobic deposit. Shale coloration in this region is deceptive; Hamilton shale deposits, west of the Rochester meridian all display a proportionately lighter color. The eastward darkening and hardening of shales as well as muddy limestones and siltstones is pervasive throughout the Middle Devonian section; one explanation, which we favor and which is a subject of ongoing study by the present authors, is that shales in this region experience higher late diagenetic burial temperatures. This effect makes it harder to prepare fossils collected in this region and further east, and it can fool inexperienced geologists into mismatching various units in correlation and misinterpreting paleoenvironments.

However, some regional color changes in the shales at this level reflect real paleoenvironmental controls; the Ledyard of this region

grades both to the west and east into various lighter colored, richly fossiliferous facies (Baird and Brett, 1985: Fig. 5). This pattern mirrors that in the overlying King Ferry and higher Ludlowville and Moscow units in this region, as well as within the underlying Centerfield Member; a region of differential subsidence and deeper-water conditions (Finger Lakes trough) was centered in this region during the deposition of these beds (see Baird, 1981; Baird and Brett, 1981; Brett and Baird, 1985). This feature explains the differentially darker color of the Ledyard and certain higher King Ferry shale units in the Cayuga Valley as well as the differentially thick and muddy character of the Centerfield Member in this area (see Brett and Baird, 1985).

The Centerfield-Ledyard contact in the Genesee Valley-Batavia area appears conformable and gradational, but in the Central Finger Lakes Region, the boundary is distinctly abrupt (Gray, 1984; Brett and Baird, 1985). Beginning to the west, at Wilson Creek and Kashong Glen near Geneva, this boundary is expressed as a layer of brachiopod and coral debris (Moonshine Falls Bed), suggesting minor scour or a period of nondeposition. East of Seneca Lake, the boundary is clearly erosional, but, in many sections, the horizon is cryptic; bioturbation activity timed with initial burial of the discontinuity at the beginning of Ledyard time, has essentially scrambled together top-Centerfield and basal Ledyard muds to the point that the discontinuity surface has been erased (Baird and Brett, in press). Such contacts, termed stratomictic (see Baird, 1981), typically cannot be found in a uniform shaly sequence by cursory examination. Depositional breaks of this type are frequently located when a biofacies discontinuity is first identified (see Baird, 1978) or when the stratomictic contact is correlative with a more readily identifiable unconformity such as is developed along the top of the Centerfield at Moonshine Falls (Stop 6). Hence, the minor Centerfield-Ledyard discontinuity (Moonshine Falls Bed) west of Cayuga Lake becomes more readily recognizable in the Aurora area, both because the magnitude of the erosion is greater and because the basal Ledyard shale above this break is distinctly darker than that observed above the contact at other localities.

In actuality, the Moonshine Falls section is quite anomalous when compared to outcrops in the immediate vicinity; virtually the entire upper transgressive part of the Centerfield section has been removed by erosion at this locality (Fig 4). In three nearby sections, respectively to the west, south, and north of Moonshine Falls, the 3 to 4 meter-thick resistant, medial Centerfield calcareous mudstone interval is succeeded by a comparable thickness of less calcareous, shaley mudstone (Gray, 1984). This sequence contains a biofacies succession recording transgression up to the relatively minor discontinuity horizon (Moonshine Falls Bed) below the Ledyard. At Moonshine Falls, this entire upper Centerfield transitional shale sequence is absent (Fig. 4); Ledyard black shale deposits are juxtaposed directly on the medial Centerfield, allowing a rare opportunity to study the spectacularly fossil-rich facies of the main Centerfield calcareous mudstone on a large clean bedding plane surface (see Road Log: Stop 6).

The cause for this local downcutting is problematic. Even in sections in the Owasco Valley further east, several meters of transgressive Centerfield shaley mudstone are always observed above the medial Centerfield sequence. However, we suspect that the Moonshine Falls area was very certainly not a single closed erosional depression into upper Centerfield beds. More likely, this erosional cut-out is the expression of an erosional channel or runnel into the Centerfield deposit which happened to coincide with this locality, or, conceivably, the differential bevelling of a local area of upfolded Centerfield beds. Although the explanation for this local anomaly remains enigmatic, the character of the erosion process can be deduced from the character of the Moonshine Falls Bed discussed below.

At all sections, from the vicinity of Aurora to the Chenango Valley, in central New York, the Moonshine Falls Bed is an easily identifiable horizon, marked by abundant phosphatic pebbles and fossil steinkerns, as well as by abundant reworked and highly corroded fossil fragments derived from Centerfield beds. Moonshine Falls is the best locality for sampling the phosphatic material, but Ensenore and Seward ravines on Owasco Lake also yield excellent pebbles and steinkerns at this level. On the west side of Cayuga Lake at Hicks Gully, Big Hollow Ravine, and in the unnamed brook adjacent to and downstream from the Poplar Beach shale pit (Optional Stop), broken pyritic borrow tubes, as well as phosphatic debris, occur within the Moonshine Falls Bed; these pre-cemented, pyrite-impregnated burrow castings show evidence of secondary disturbance by burrowing infauna which penetrated the mud-floored discontinuity surface (Baird and Brett, in press).

East of Skaneateles Lake the Moonshine Falls Bed thickens and splays into several discrete, shell-nodule debris layers which are most likely multiple-event storm (tempestite) beds. In this valley and eastward to the vicinity of Hamilton, New York, the Moonshine Falls Bed becomes more of a condensed sedimentary sequence than a simple erosion surface *per se*; this unit, both in this area, and also west of Seneca Lake, is thus believed to record little or no submarine erosion and minimal sediment accumulation.

The nature of the erosion process which produced the top-Centerfield discontinuity appears to be closely linked to the major transgression that began during deposition of the upper Centerfield; it seems likely that reduced sediment supply, associated with transgression-induced migration of coasts and sediment sources away from western New York, may have produced a situation in which erosion, not deposition, became the predominant process. During such a time, the burrowing activity of numerous organisms in surface muds would have served to soften and liquify these same muds, hence increasing their erodability under storm conditions; over a long period of time, repeated scour of these muds in shallow settings could have produced significant erosion and the conspicuous hydraulic concentration of phosphatic debris and shell fragments (see Baird and Brett, 1981: Stratomatic Erosion Model). In deeper areas within the Finger Lakes Trough, a greater proportion of the bottom erosion may have been produced by secondary, storm-induced bottom currents.

CORRASIONAL DISCONTINUITIES IN THE GENESEE FORMATION:
THE PROBLEM OF THE FIR TREE AND LODI LIMESTONE BEDS

Recent stratigraphic study of the lower and middle parts of the Genesee Formation (uppermost Middle- to lowermost Upper Devonian stages) by the present authors has produced important results and surprises. Much of this work has been a detailed stratigraphic study of the basal Genesee Leicester Pyrite Member, a well-known pyritic bone bed flooring the formation (see Brett and Baird, 1982; introduction to the present paper). However, our recent studies have been directed toward a microstratigraphic study of that part of the Genesee Formation (uppermost Genesee Member and lowermost Penn Yan and Sherburne Members) which contain the newly-described Fir Tree and the slightly younger Lodi Submembers (Figs. 1, 12, 13). These thin, fossiliferous, impure carbonate layers are closely associated with discontinuities which are produced by combined processes of abrasion and dissolution within a largely anoxic basinal environment; both of these limestones are clearly bevelled by corrasion in gently sloped settings. The chemical end products of this erosion, including reworked pyritic grains and bone fragments, serve as telltale markers of the resulting cryptic discontinuities developed within black shale facies where the limestone deposits have been destroyed.

Recent study of the Lodi interval by the present authors shows that two different carbonate-rich units had been lumped under the name Lodi in earlier work (see work of deWitt and Colton, 1978); these units, including a lower layer (Fir Tree Limestone Submember) within the upper Genesee Member and the true Lodi, which occurs slightly higher stratigraphically at the base of the Penn Yan and Sherburne Members, can be traced over large areas in the Finger Lakes Region (Figs. 1, 5-10).

The Fir Tree Submember, named for excellent cliff exposures at Fir Tree Point on Seneca Lake, is separated from the younger Lodi Bed by four to thirty meters of black to dark grey, silty shale which we designate the Hubbard Quarry Shale Submember for the complete and accessible exposure at Hubbard Quarry in Interlaken Township in Seneca County (see Stop 3: Fig. 13). This shale unit, equivalent to hard, black shale facies of the uppermost Genesee Member in western New York, thickens southeastward from 4.3 meters at Hubbard Quarry to 18 meters near Genoa, Cayuga County and to approximately 30 meters north of Renwick, Tompkins County. The lower part of the Hubbard Quarry submember is composed of black, laminar, silty shale usually rich in the inarticulate brachiopod Orbiculoidea lodiensis and the articulate brachiopod Leiorhynchus quadricostatum; the black shale facies grades both upward and southeastward to olive grey, silty mudstone and flaggy siltstone facies which contain, in addition to the above taxa, local bedding plane concentrations of auloporidae corals.

The Fir Tree Submember is a zero to ten-meter-thick, wedge-shaped deposit of interbedded concretionary, grey to chocolate brown, micritic limestone and shale which expands conspicuously southward in the central Finger Lakes Region (Figs. 5-7). The Fir Tree sequence becomes increasingly fossil-rich, calcareous, and condensed as its northern

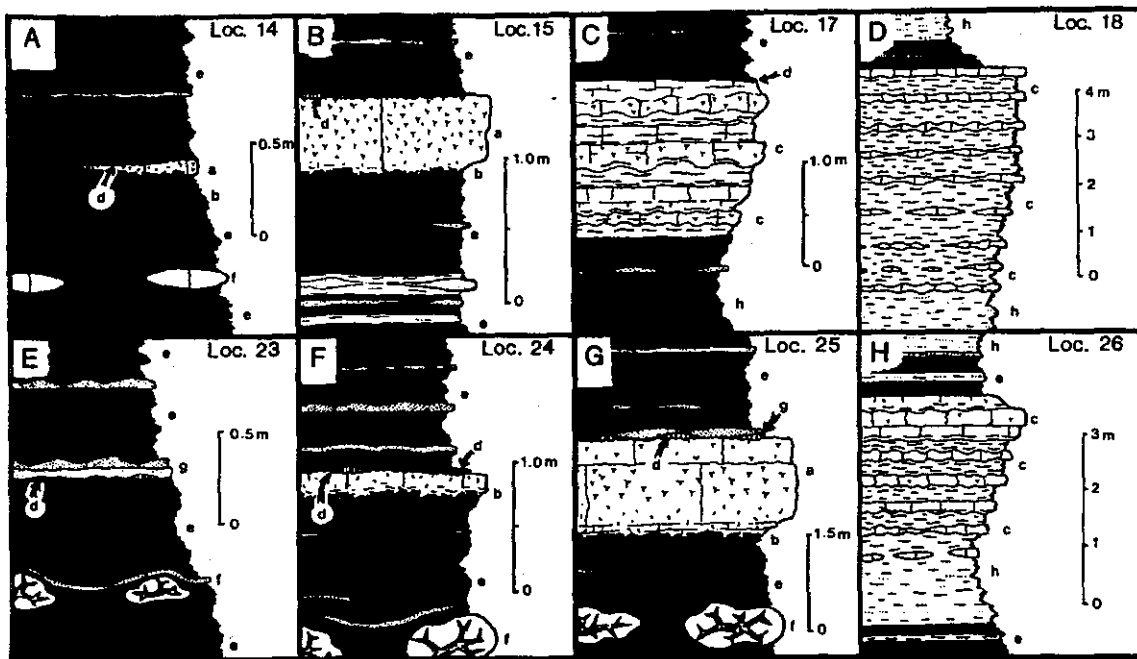


FIGURE 5.--Fir Tree Submember; selected sections along two transects normal to depositional facies strike. A-D is a transect from Hubbard Quarry southeastward to Trumansburg Creek (also see Fig. 6). E-H is a transect from Venice Center southward to Lansingville. Inferred upslope direction to the right. Note concurrent downslope condensation of Fir Tree beds and enrichment of these beds with auloporid corals; this paradoxical pattern is discussed in text; also note corrasional truncation of condensed Fir Tree section at the north end of both transects; a) condensed auloporid wackestone; b) thin, orbiculoid-rich, basal grey mudstone layer of Fir Tree Bed; c) thick, marginal facies of Fir Tree unit; sequence is mostly grey mudstone with sparsely-fossiferous concretionary lime mudstone layers; d) lenses of reworked detrital pyrite on top Fir Tree discontinuity surface; e) black shale facies with minor siltstone and grey mudstone interbeds; f) septarial concretions; g) siltstone bed draped over discontinuity and detrital pyrite; h) grey silty mudstone below thick Fir Tree deposits.

erosional margin is approached (Figs. 5-7: see stop sequence 1 to 4 in road log); thin (≤ 1 meter-thick), northernmost, Fir Tree sections consist of a dense, falls-forming ledge of impure limestone which contains bedding-plane mats and thickets of auloporid corals, small *Devonochonetes* sp., a sparse molluscan fauna, and numerous pyrite-suffused burrows (Fig. 5). This fauna disappears southward as the Fir Tree thickens exponentially, and the sequence becomes almost devoid of fossils as it changes laterally to a sequence of interbedded dark grey mudstone, concretionary "ribbon" limestones, and turbiditic siltstones (Figs. 5- 7).

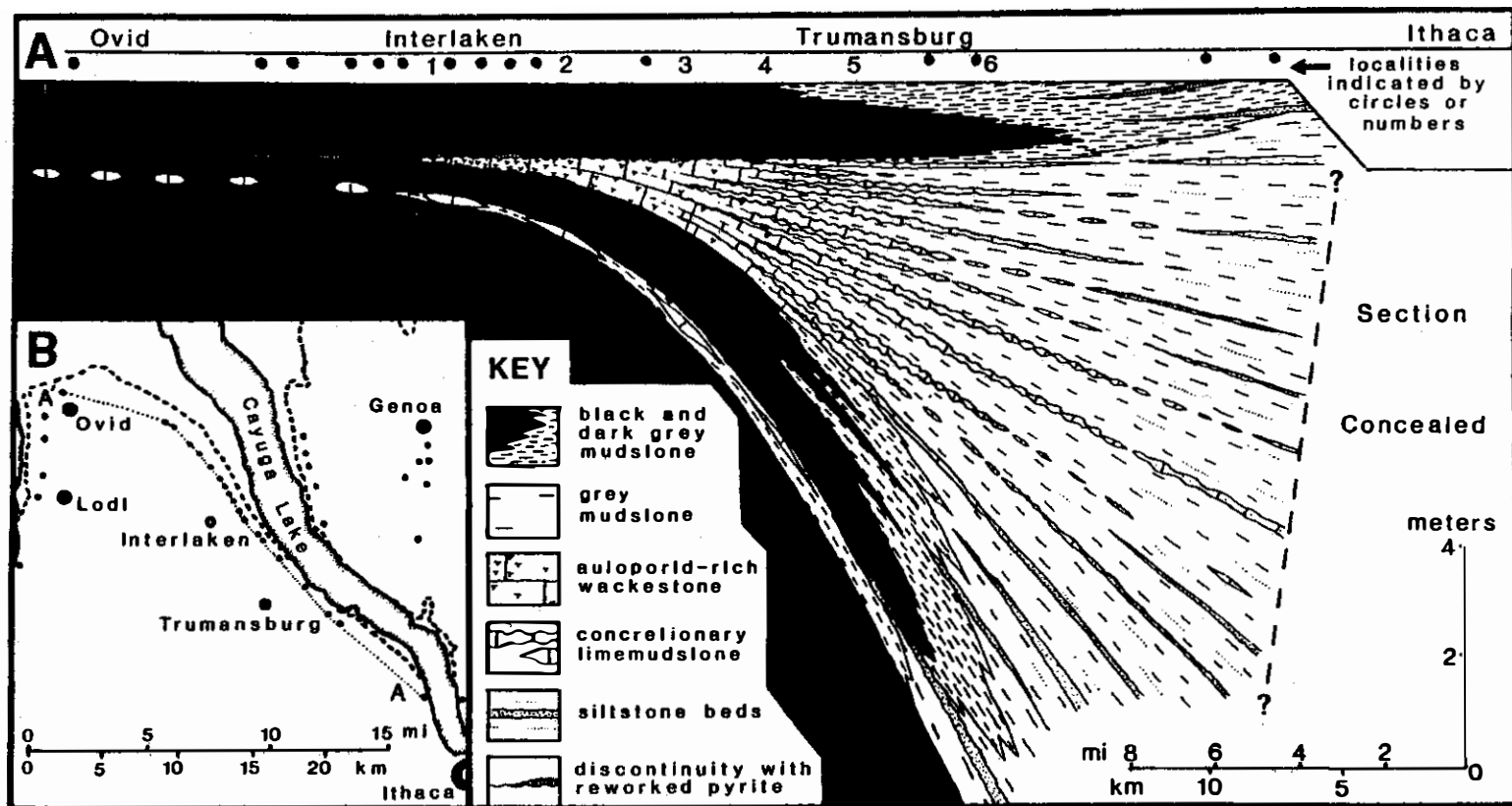


FIGURE 6.--Stratigraphic transect of the Fir Tree Submember approximately normal to depositional strike; note very rapid southwestward thickening and splaying of bed; presumed basal direction toward northwest.

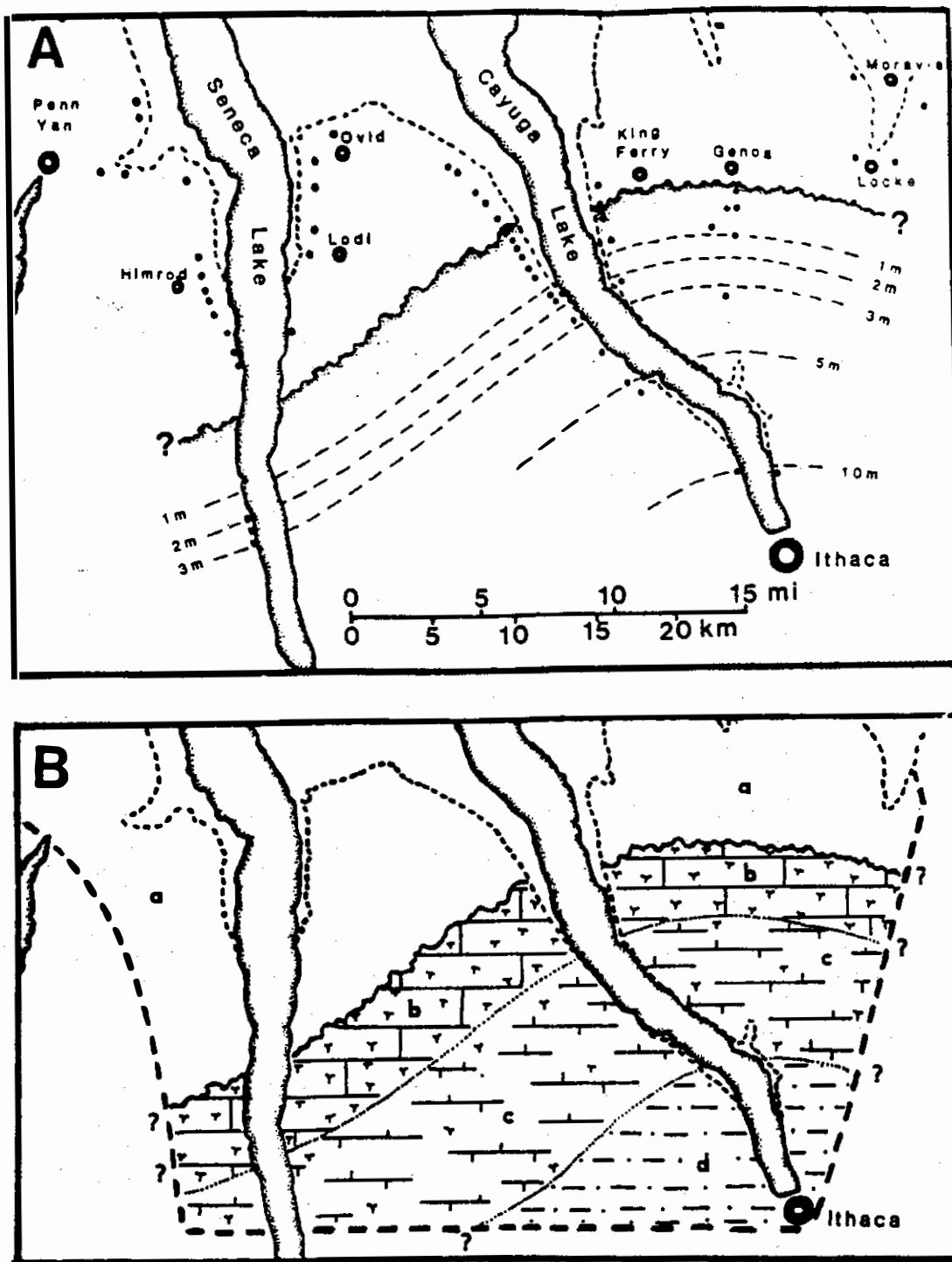


FIGURE 7.--A) Isopach map of Fir Tree Submember in the Seneca-Cayuga Valley region; note abrupt northward thinning due in part, to erosional truncation at top of bed; B) Facies map of Fir Tree Submember in the same region; symbols: a) black shale; b) auloprid-bearing limestone; c) sparsely fossiliferous calcareous siltstone.

As the Fir Tree Submember thins northward in the Cayuga and Salmon Creek valleys, its upper contact with the Hubbard Quarry Shale Submember changes progressively from gradational to abrupt and erosional along three south-to-north transects: from Little Point to Interlaken (west side of Cayuga Lake), Lake Point to King Ferry (east side of Cayuga Lake), and Lansingville to Genoa (east of Cayuga Lake), the Fir Tree is progressively cut out by a discontinuity that originates at the Fir Tree-Hubbard Quarry shale contact (Figs. 5, 6). North of the Fir Tree erosional limit this discontinuity persists within undifferentiated black shale facies of the Genesee Member (Figs. 5, 6).

The Lodi Limestone Submember (or Bed), named for a particularly good exposure on Mill Creek, Seneca County (see Lincoln, 1897; Kirchgasser, 1985), has now been found to be widely distributed throughout the Cayuga and Salmon Creek Vallies and is now known to extend continuously from the vicinity of Himrod, Yates County to the Skaneateles Valley (Figs. 8-10). This unit, like the Fir Tree Submember, is an impure, concretionary carbonate deposit that contains conspicuous bedding plane and thicket accumulations of aulopodid corals (Fig. 9). Also, like the Fir Tree, the Lodi displays lateral sedimentary condensation towards an erosionally bevelled margin in the Seneca Lake Valley, except that this concurrent condensation and truncation is in a northwestward rather than northward direction (Figs. 9, 10).

As with the Fir Tree, Lodi carbonate content and fossil diversity increases notably as the Lodi erosional limit is approached near Himrod, Yates County and Ovid, Seneca County (Figs. 8-10); at these places, the Lodi yields a moderate diversity fauna dominated by aulopodid corals, but also yielding Orbiculoidea lodiensis, the articulate brachiopods Pseudoatrypa cf. P. devoniana, Devonoconetes sp., Leiorhynchus sp., orthoconic nautiloids, the goniatite Ponticeras perlatum, pelmatozoan debris, and pyrite suffused burrows. This fauna becomes less diverse as the Lodi grades southeastward to silty mudstone and muddy siltstone lithofacies near Renwick, Tompkins County and at Locke, Cayuga County. Only further east, at New Hope Mills, Onondaga County, does a conspicuously diverse, shelly fauna appear in Lodi-equivalent siltstone and fine sandstone lithofacies (Fig. 8).

As with the Fir Tree deposit, the Lodi thickens away from its northeastward erosional margin, but unlike the Fir Tree, it reaches a maximum thickness of only one to two meters in the Cayuga and Salmon Creek valleys. The Lodi is closely associated with two to three recurrent beds of similar character which occur above the Lodi Bed proper (Fig. 9).

Both the Fir Tree and Lodi beds are missing in the northern Seneca Valley, in the vicinity of Penn Yan, Yates County, and in the Canandaigua and Bristol Valleys; in this region, their positions are marked by cryptic discontinuities within black and dark grey shale facies which have been identified at only a few localities (Fig. 9). In the Genesee and Honeoye Valleys, an aulopodid-rich concretionary limestone unit corresponding to the Lodi Bed is observed; this unit displays a southeastward erosional margin in the Honeoye Valley which closely resembles the northwestern

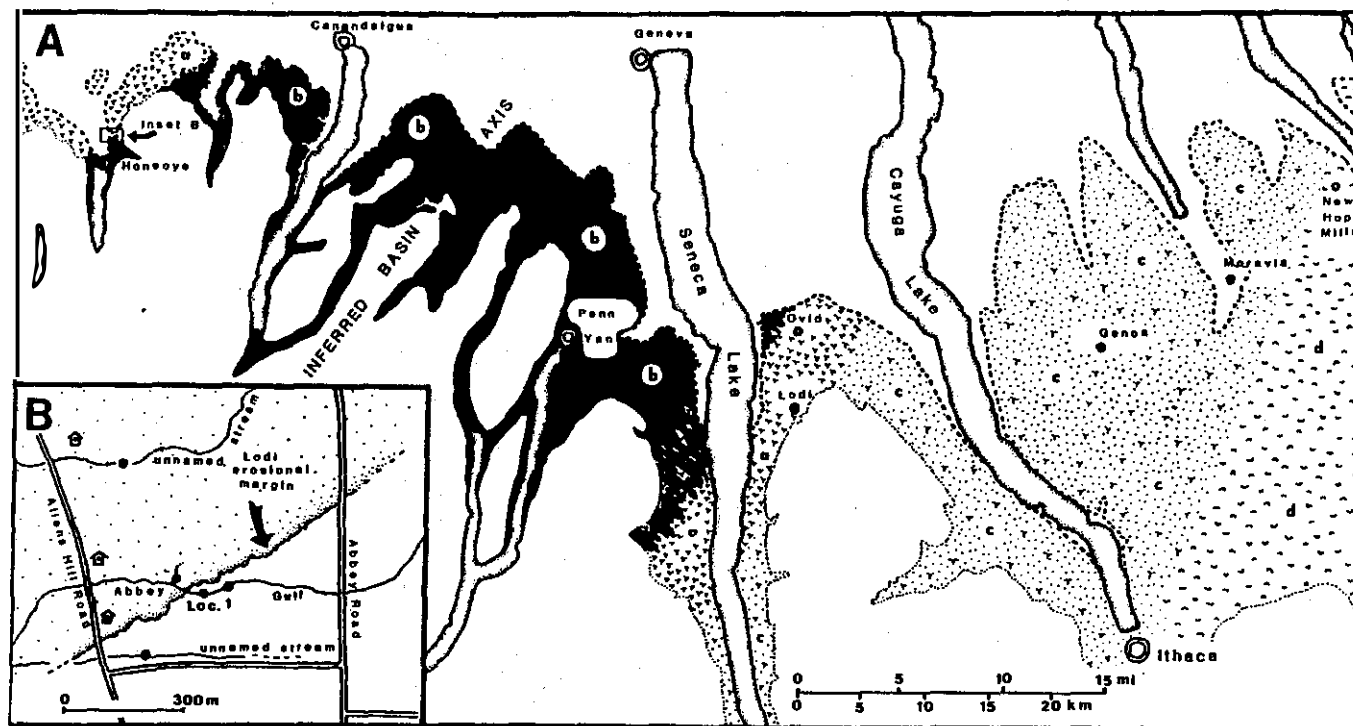


FIGURE 8.--A) Lodi facies belts and inferred basin axis. B) inset shows details of the eastern erosional limit of the Lodi Submember west of the basin axis; symbols: a) aulopod-bearing, concretionary wackestone; b) dark grey to black shale with minor bed of pyritic, bone-rich remanic sediment; area of basin axis; c) sparsely fossiliferous, aulopod-bearing, calcareous siltstone; d) fossiliferous, brachiopod-rich siltstone.

terminus of the Lodi (Fig. 8). Similarly a lower Genesee carbonate interval is recognized in the Genesee Valley; this unit may correspond to the Fir Tree level, but the correlation is only tentative at this time.

The northern Seneca Valley-Bristol Valley region is believed to mark the axis of a depositional basin during the time of lower Genesee deposition (Baird and Brett, 1985); the preponderance of monotonous black shale lithofacies and the absence of several grey shale sequences and fossiliferous horizons observable in the lower Genesee Formation, both to the east and west, suggest that the basin axis was probably centered in the Canandaigua-Geneva meridian. Paradoxically, the region of maximal erosional bevelling of the Fir Tree and Lodi beds is also in this area or bordering it; both of the limestone beds, and, at least, one additional limestone unit, are truncated in the basinward (downslope) direction such that black shale deposits overlying each limestone are juxtaposed on underlying black shales where complete bevelling of the carbonate units has occurred (Figs. 8-10).

Not surprisingly, the character of these discontinuities and of the associated lag debris upon them is distinctive, providing several clues as to the process of erosion which may have produced these breaks. Firstly, the top-Fir Tree and top-Lodi discontinuities, where developed, are overlain by laminated black shale deposits. These unconformities are only two of many similar submarine erosion surfaces which are overlain by black shale deposits (see Brett and Baird, 1982; Baird and Brett, in press). Where the Fir Tree and Lodi are overlain by grey mudstone, the upper contacts appear conformable, but where the grey facies grades laterally and downslope to black sediments, a discontinuity appears at the contact.

Secondly, both the Fir Tree and Lodi discontinuity surfaces, as well as a more minor intervening diastemic contact in the Bergen Beach ravine (Stop 2) near Interlaken, are littered with reworked detrital pyrite debris along with lesser quantities of bone, quartz sand, and orbiculoid fragments (Baird and Brett, 1986; this report). Much of this material consists of reworked, fragmental, pyrite-suffused castings of burrows derived from the bevelled, burrow-rich, limestone unit (Fig. 2). In a paper now in press, we present several criteria for the recognition of reworked pyrite; very important among these is observed one-to-one similarity of detrital pyrite grains, such as burrow tube fragments on the Fir Tree and Lodi, with in-situ pyritized tubes within these beds.

However, the one conclusive line of evidence for pyrite exhumation, best developed in the top-Lodi pyrite layer near Himrod, Yates County, is the aforementioned evidence of reorientation of geopetal pyritic stalactites (Fig. 2). Where early diagenetic, microbial sulfur-reduction occurs in voids such as in cephalopod chambers or in unfilled burrows, stalactites of framboidal pyrite often develop as projections from void ceilings (Hudson, 1982). Because no "stalagmites" develop from this process, the stalactites serve as a geopetal indicator. However, if pyrite grains have been exhumed and hydraulically concentrated on the sea floor, one would predict a random orientation for stalactites within these

grains. This prediction was fulfilled when cross-sections of top-Lodi pyrite lag grains were examined; stalactites developed in void spaces within in-situ Lodi burrows were always downwardly-oriented, while reworked tubes on the Lodi upper surfaces displayed stalactites oriented in all directions (Fig. 2).

The third distinctive feature of post-Fir Tree and post-Lodi discontinuity debris is the absence or rarity of reworked calcareous fossil fragments associated with the detrital pyrite. There is an abundance of tabulate coral, brachiopod, and pelmatozoan material within the underlying Fir Tree and Lodi source beds but almost none of this occurs with the detrital pyrite. This pattern is hardly an anomaly where black shales unconformably overlie fossil-rich strata; both the aforementioned Leicester Pyrite Member below the Genesee Member and the Skinner Run Bone Bed below the black Cleveland Shale Member are accumulations of detrital pyrite which contain little or no associated carbonate allochems (Brett and Baird, 1982; Mausser, 1982).

We believe that the absence of carbonate allochems is integrally related to the abundance of reworked pyrite. In essence, it is a direct consequence of the special basinal conditions within which this submarine erosion had occurred; under the predominantly anaerobic conditions recorded by the black shale deposits, exposed calcareous grains would have been unstable for the very reason that the pyrite was stable. Low pH conditions on the anaerobic or minimally dysaerobic basin floor would have produced conditions favoring the dissolution of exposed carbonate, particularly grains exposed for long periods of time on an erosion surface. Anaerobic conditions which could have produced a pitted corrosion surface on the Cherry Valley or Onondaga limestones in western New York localities, would have also produced distilled lag accumulations of pyrite and bone debris on discontinuities such as those in the Genesee Formation.

The last feature of the Genesee discontinuities discussed here is the fact that the pyrite lag accumulations always occur as laterally separated lenses when seen in outcrop profiles. This is also characteristic of the Leicester and Skinner Run pyrite accumulations, as well as other more minor deposits (see Brett and Baird, 1982; Mausser, 1982). This lensing is the two-dimensional perspective view of either tractional wave-bedforms of pyrite gravel or profiles of debris-fillings of erosional channels or runnels on the sea bed. In the instance of the Leicester pyrite study, we concluded that Leicester lenses were fillings of erosional runnels similar to erosional furrows on certain modern marine substrates (see Flood, 1983); we based our conclusion on the absence of foreset bedding in lenses, the overall width-dimensions of lenses, and the fact that the azimuths of current-aligned pyrite tubes usually paralleled the long axes of Leicester lenses. Whatever the correct interpretation of the Leicester lenses may be, it should apply similarly to the younger Fir Tree and Lodi pyrite lags since these display similar patterns of grain-current alignment and gross lens geometry.

When the distinctive features of the post-Fir Tree and post-Lodi discontinuities are reviewed collectively, a paradoxical pattern emerges - the submarine erosion appears to be intimately associated with the

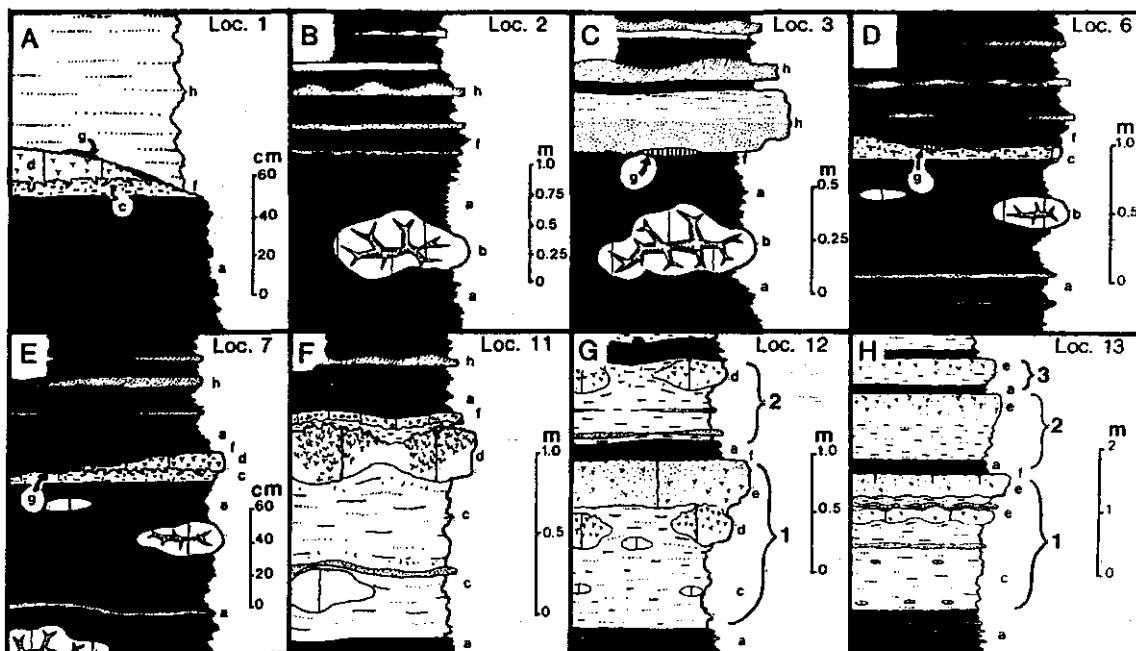


FIGURE 9.--Lodi Limestone Submember; selected sections along transect from the Honeoye Valley to Cayuga Lake. Sections F, G, and H are also shown on Figure 10. Note that section A (Abbey Gulf) displays the eastern erosional limit of Lodi deposits west of the basin and sections D and E show the northwest (downslope) limit of this unit southeast of the basin. Numbers in sections G and H denote Lodi (1) and two younger and similar sedimentary cycles which are developed concurrently at localities between Ovid and Interlaken. Key: a = black shale; b = septarial concretions; c = basal Lodi orbiculoid- and pyrite burrow-tube-rich grey mudstone; d = auloporida-rich concretionary limestone (good Lodi development); e = auloporida-bearing, calcareous, burrowed siltstone bench; f = Lodi discontinuity; g = lenses of reworked detrital pyrite on Lodi discontinuity surface; h = silty mudstone and current-rippled siltstone beds.

presence of overlying black shale deposits, and absent where the black shale tongues are absent. This intimate association of laminated black shale facies with hydraulically concentrated detrital pyrite and bone debris is similarly documented for the Leicester and it is most assuredly not a coincidence. The transgressive, upslope migration of the anaerobic lower water zone appears to have been closely associated with the onset of bottom scour and corrosion of the sea floor.

Two possible explanations for this erosion, one rather intuitive and the other much more theoretical and controversial, are provided here. The first is an oft-cited, but reasonable explanation - that the observed submarine erosion is related to the reduction in sediment supply during

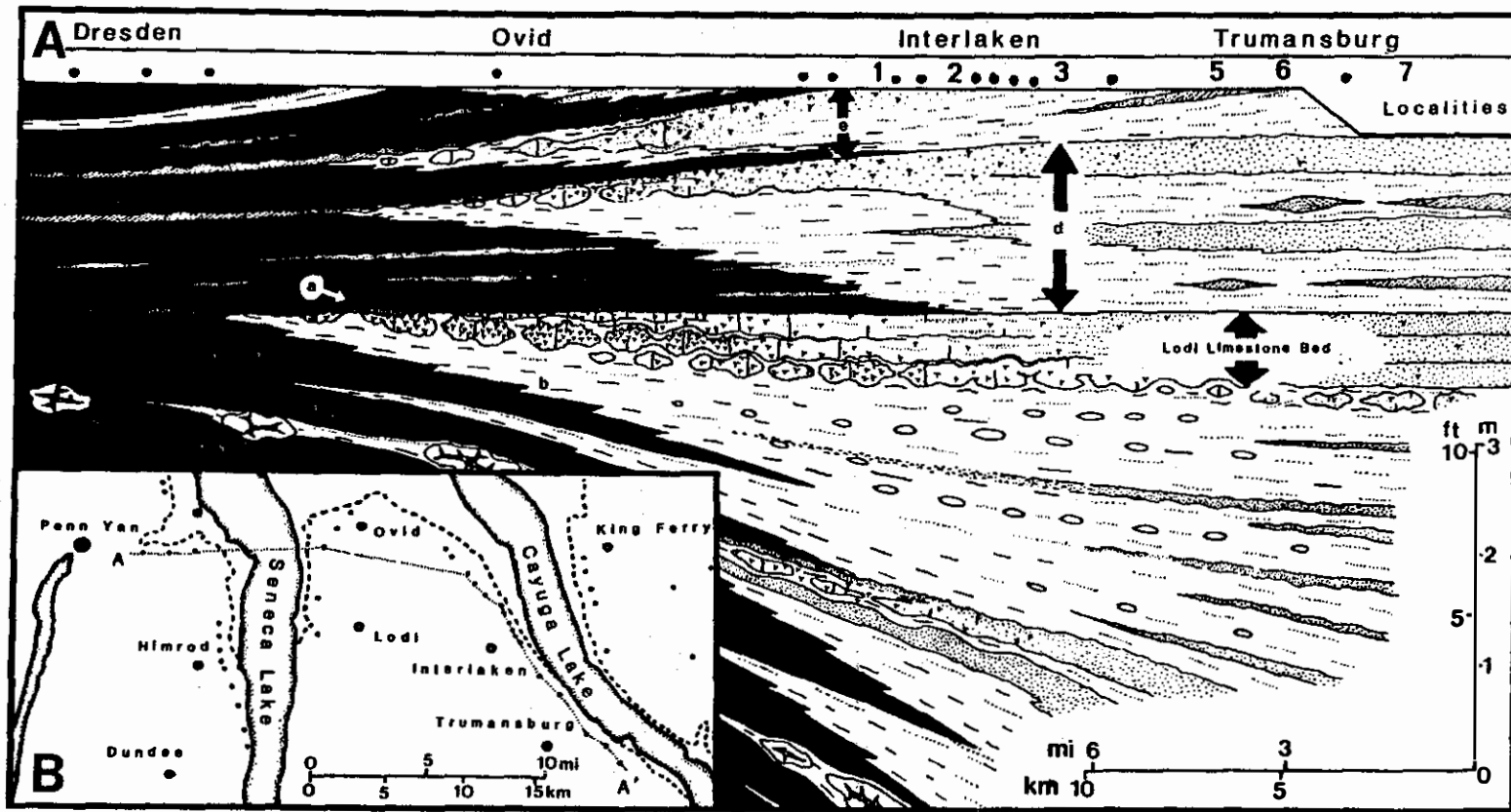


FIGURE 10.--Stratigraphic transect of the Lodi bed normal to depositional strike; symbols: a, pyritic lag deposit along truncated edge of Lodi Submember; b, d, e minor shallowing upward cycles which begin with grey bioturbated mudstone and culminate in silty limestone with auloporphids.

transgression; a net rise of sea level would presumably cause an eastward migration of coasts away from the site of deposition resulting in nearshore alluviation of sediments which would have normally been deposited in the offshore Genesee basin (see McCave, 1973; Heckel, 1973 for application of nearshore alluviation model to explain episodes of Devonian carbonate deposition). In such an instance, a reduction in terrigenous mud supply to the Fir Tree or Lodi substrate during deposition could have shifted the sediment budget balance from net accumulation to nondeposition or erosion.

A more difficult and important question, posed by our data, can be stated as follows: what causal mechanism can account for the concentration of gravel-grade pyrite placers within a black shale environment? In the absence of a known modern analog we have to resort to specific known current processes which could partly explain deposits of this type. Three processes discussed here include: 1, bottom currents and turbulence generated by deep storm-generated gravity waves; 2, latent current activity in the deeper part of the basin, which is originally storm-generated, but which persists at depth long after the storm has passed; and 3, internal wave activity within the pycnocline which impinges against the basin margin slope.

The first two of these processes, storm-generated turbulence and latent storm-generated deep-water currents, are known to be important on the continental shelves and slopes as well as in large lakes; these are assumed to have been of some importance in particle transport in the Devonian shelf and basin settings. Deep storm wave turbulence is known to transport sediment to depths of 200 meters or more (Liebau, 1980) and storms can also generate boundary currents which can produce erosional furrows to depths of 200 meters as has been observed on the floor of Lake Superior (Johnson, et al., 1984). Furthermore, latent, storm-generated, cyclone-like disturbances are known to be important in scouring the sea floor and transporting sediment at bathyal depths (Hollister, et al., 1984). In particular, the episodic character of storm currents and latent, storm-generated current disturbances could explain the numerous, recurring levels of current-winnowed silt and sand which are characteristic of Devonian black shale deposits.

However, the erosional bases of the black shale units plus the occurrence of heavy, gravel-grade pyritic allochems on these discontinuities suggest that some process-specific erosion mechanism is associated with the transgressive upslope migration of the anaerobic water mass up the basin margin slope. The fact that both the Fir Tree and Lodi beds as well as their probable equivalents in western New York are differentially bevelled in the basinward (downslope) direction would seem to indicate that erosion may have been associated with the water mass boundary layer.

We believe, as do Ettensohn and Elam (1985) that a dynamic water mass boundary layer or pycnocline was developed in the Devonian slope and basin settings during periods of black shale deposition. Since pycnoclines are boundaries between layers of different temperature- and/or, salinity-related water density, they may behave, to a limited extent, like

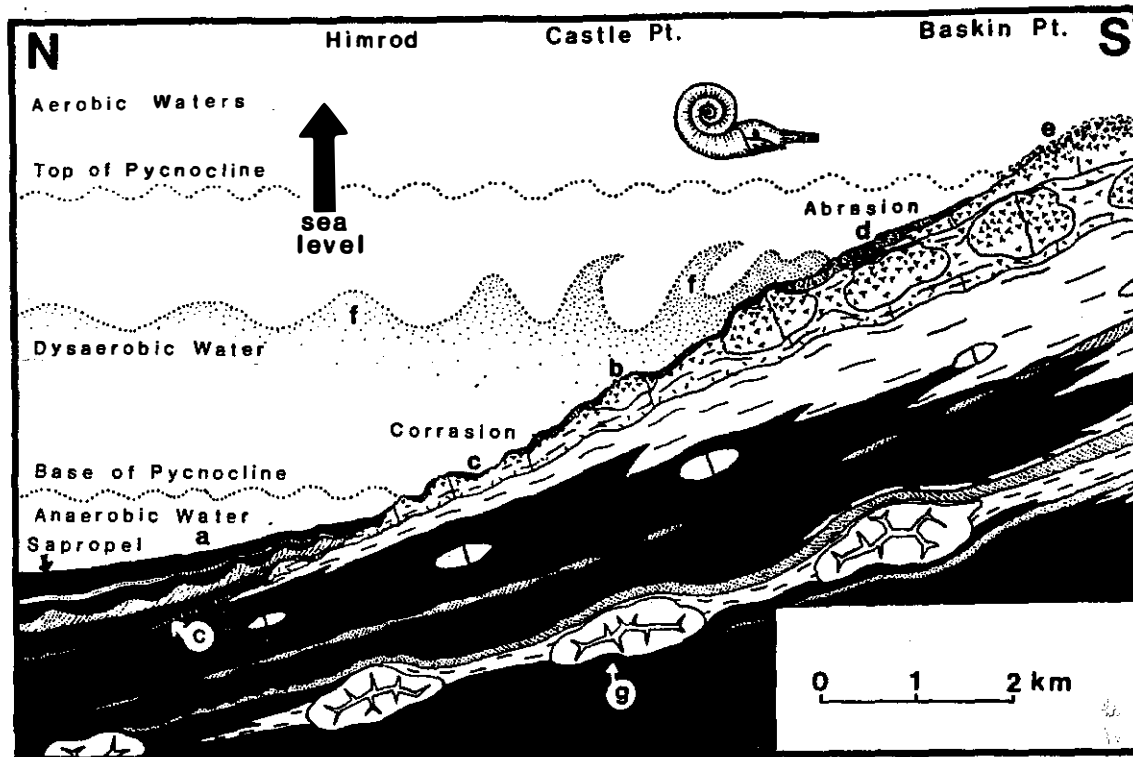


FIGURE 11.--Erosion model. Schematic shows process of mechanical erosion and corrosion of Lodi carbonate deposits during a transgression event. Internal wave-impingement on the paleoslope is believed to have been a significant factor in the submarine erosion process. As shown here these waves generate turbulence on slope, removing fines, and exposing carbonate material for corrosional attack in dysaerobic water. Residual early diagenetic pyrite and phosphatic fossil debris is concentrated as a tractional placer in lenses. Upslope migration of the pycnocline produces a widespread, gently-inclined unconformity which is progressively buried by overlapping black muds and silts as water deepens. Key: a) black muds; b) Lodi concretionary, auloporid-rich wackestone which is truncated by corrasion; c) reworked detrital pyrite (corrasion residue); d) top-Lodi veneer of mixed pyrite-calcareous debris; packstone; e) living auloporid thicket; f) Internal wave-train impinging paleoslope; g) septarial concretions marking stratigraphic position of similar Lodi-type cycle ("Bergen Beach") developed further upslope to southeast.

the air/water surface when the boundary is abrupt; energy transferred to the pycnocline will propagate along that layer in the form of waves; these internal waves are believed to be of some significance in transporting sediment on the outer continental shelf (Karl, et al., 1983).

In two recent papers, Woodrow (1983, 1985) has applied the internal wave concept in explaining a distinctive facies sequence within regressive, prodeltaic facies of the upper Genesee Formation in the Seneca Lake Valley. He attributes the formation of an interval of thick siltstone beds, which overlie a thick unfossiliferous turbiditic sequence and underlie fossil-rich, silty delta platform deposits to the impingement of internal waves against the upper prodelta slope during delta progradation.

In a tentative model presented here, we envision the impingement of internal waves against the basin-margin slope, during a period of net sea level rise and during a period of transgression-related sediment starvation on the basin slope as opposed to the progradation setting discussed by Woodrow. In this model, the zone of internal wave impingement on slope would migrate upslope during the transgression producing a continuous erosion surface which would extend to the maximum transgressive upslope limit reached by the lower water layer (Fig. 11). As erosion in the impingement belt proceeded, sediment fines would have been swept downslope (bypassed) onto the anaerobic lower slope to be deposited as laminated black mud and thin, winnowed siltstone layers above the discontinuity surface (Fig. 11). Exposed shells and concretionary carbonate would be destroyed by a combined process of abrasion and corrosion (corrasion) downslope from the belt of maximal internal wave impingement, but upslope from the low energy zone of black mud overlap on the discontinuity surface (Fig. 10).

In summing this process discussion, we believe that the Fir Tree and Lodi discontinuities, plus numerous other similar breaks below black shale deposits, may be a time-lapse record of the pycnocline position on the ancient basin slope. This explanation is still only a model, and it will probably remain so until a modern analog of these erosion surfaces is found. Such a discovery is anticipated, given the great increase in studies of modern basinal processes, and we believe that it will significantly change some of our views concerning the origin of black shale deposits.

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ROAD LOG FOR MIDDLE DEVONIAN CORRASIONAL DISCONTINUITIES
IN THE CAYUGA LAKE AREA

Cumulative Mileage	Miles from Last Point	Route Description
0.0	0.0	Begin trip at junction of Route 13-34 and Routes 79, 89-96 in town of Ithaca. Turn right (north) on Route 79-89-96.
0.2	0.2	Junction of New York Routes 89 and 96. Turn right (north) on Route 89.
0.2	0.8	Leave town of Ithaca, proceed north along west side of Cayuga Lake.
1.4	0.4	Tremain State Marine Park
1.5	0.1	Outcrops of Sherburne siltstone (Genesee Formation) on left.
3.5	2.0	Town of Ulysses.
4.6	1.1	Glenwood Creek; falls over Fir Tree beds (lower Genesee Formation) just east of road.
7.1	2.5	Cross Willow Point Creek; good exposure of Tully Limestone and Windom Shale.
9.2	2.1	Cross mouth of Taughannock Creek at Taughannock State Park. To the left (upstream) from the road bridge a low waterfall is visible; this section includes the Tully Formation (limestone) overlying the uppermost Hamilton Group (Windom shale); we will examine the gorge section upstream from this site.
9.4	0.2	Junction with road to Taughannock Falls overlook; turn left (west).
10.2	0.8	Parking area for Taughannock Falls overlook (Stop 1); pull off and walk down steps to viewing area.

STOP 1. TAUGHANNOCK FALLS

Location: Taughannock Falls Park overlook on north side of Taughannock Creek about 1 km southwest of N.Y. Route 89 town of Ulysses, Tompkins Co., N.Y. (Ludlowville 7.5' Quadrangle).

References: Cornell University Department of Geology (1959); deWitt and Colton (1978); Kirchgasser (1981, this volume).

Description: This overlook provides a spectacular view of the 66 m (215') high Taughannock Falls (purportedly the highest falls east of the Mississippi), a hanging valley at the end of about a mile long

post-glacial gorge (Fig. 12). The configuration of the falls is controlled by prominent joints; it is held up by resistant siltstone beds in the Sherburne Member.

In the early days of the silver screen, this locality and nearby glens were used as backdrops for early westerns, owing to the spectacular canyon-like walls in these gorges. Another interesting side light of this outcrop is the occurrence of thin (1-2 cm) dikes of Mesozoic kimberlite-like (actually alnoite) intrusive rock that crop out in the stream bed at the base of the falls (see Kay and Foster, this volume).

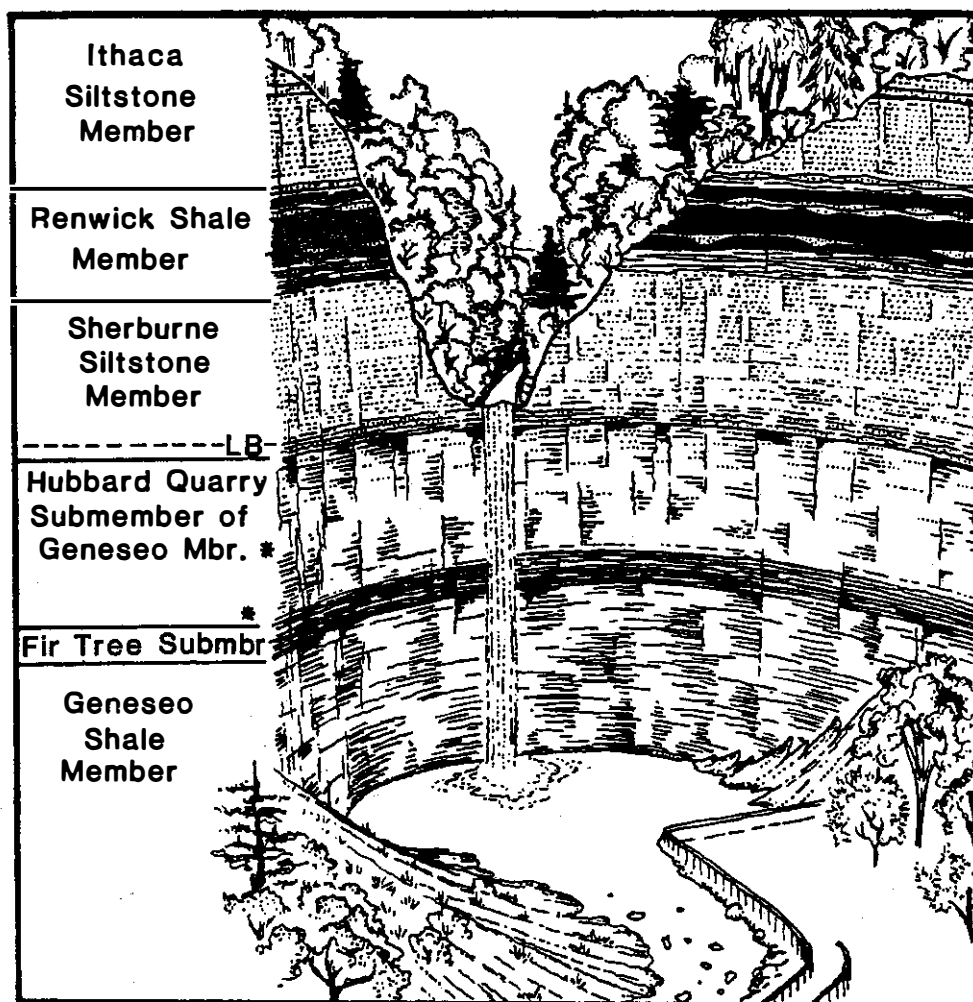


FIGURE 12.-- Great falls on Taughannock Creek showing key stratigraphic divisions and the positions of the Fir Tree and Lodi Submembers. Terminology used in figure and in text is modified from existing usage (see asterisks); the new designations (Fir Tree and Hubbard Quarry Submembers of the Geneseo Member) for the middle-upper part of the falls face correspond, respectively, to the Penn Yan Shale Member of deWitt and Colton (1978), Patchen and Dugolinski (1979), and to the lower part of the Sherburne Siltstone Member (see Kirchgasser, 1981). In this figure, the Renwick Member is shown as a black band to accentuate the lenticular siltstone beds distinctive to this interval.

Gorge walls, up to 120 m (200-400') high, expose shales and siltstones of the lower Genesee Formation of late Givetian to early Frasnian (latest Middle to earliest Late Devonian age). Depending upon the source, three or four members of the Genesee Formation are recognized at this locality. Black shales of the Genesee Member form the lowest exposed unit; the plunge pool level is about 3.8 m above the base of this member, the gradational lower contact with the Tully Formation being exposed downstream. The upper contact is about halfway up the face of the falls, 27 m (90') above the plunge pool and can be recognized by a change in coloration and bedding. The Genesee is nearly barren, platy to fissile black shale; it is largely covered by talus at the base of the gorge walls.

The overlying 26 m (80') unit in the falls face is readily recognizable as a lighter grey-weathering interval with several prominent bands (of silty limestone) near the base and a major system of smooth, vertical joints within the middle and upper parts of this division. This interval has been termed Penn Yan shale (deWitt and Colton, 1978; Patchen and Dugolinski, 1979) or simply lower Sherburne siltstone (Kirchgasser, 1981). In actuality, neither term is entirely appropriate as these shales are correlative with upper Genesee beds below the base of the Penn Yan or Sherburne elsewhere which has been placed at the Lodi Limestone bed. Recent study shows that the lower, 6-7 m (19-22') thick part of this unit, characterized by dark brown, rhythmically-bedded, silty limestone layers corresponds to a usually thin interval of auloporid coral-bearing carbonates termed the Fir Tree submember of the Genesee Member (Baird and Brett, 1986). In this particular area, however, the Fir Tree Submember is very thick and the limestone beds are very sparsely fossiliferous and seemingly turbiditic; this unit is the expanded equivalent of much thinner, condensed, and auloporid-rich, limestone facies to the north (see Figs. 6, 7, 13: stops 2 and 3). Above the Fir Tree interval is approximately 22 m (70') of silty shale which is largely black in the lower portion but which becomes progressively more silty towards the top. This sequence (Hubbard Quarry shale submember of the Genesee Member), like the Fir Tree below it, thins markedly to 4 m (13') at the type section, Hubbard Quarry (Stop 2) near Interlaken. Recently, conodonts of the disparalis zone (upper Givetian) have been obtained from the top of the Hubbard Quarry shale interval at the Lodi type section on Mill Creek (W.T. Kirchgasser pers. comm.). Ancyrodella rotundiloba, indicative of the lower asymmetricus zone, now accepted as the base of the Frasnian stage, has been observed from shales immediately overlying the Lodi horizon in westernmost Ontario County, N.Y. (Kirchgasser, pers. comm.). This indicates that the Middle/Upper Devonian boundary closely overlies the Lodi bed within the lowermost Penn Yan or Sherburne members. At Taughannock Falls the Middle/Upper Devonian boundary is probably at or near the lip of these falls.

Above the Hubbard Quarry submember is the 26 m (80') Sherburne siltstone member; the base of this unit can be seen as an abrupt change from the smooth joint-faced Hubbard Quarry interval to an irregularly-bedded and more resistant siltstone sequence about 9 m (30') below the falls lip. Although the lowermost Sherburne is obviously inaccessible at this locality, the basal 3 meters (10') of the member includes the silty, auloporid coral-rich interval of the Lodi unit, as can more easily be seen

beneath the Route 89 overpass bridge at a small, nearby creek (Loc. GEN 4 of deWitt and Colton, 1978) 0.65 miles northwest of Taughannock Point. In this region, the Lodi is composed of bioturbated calcareous siltstone which yields auloporids, small bivalves, and the diagnostic goniatite Ponticeras perlatum.

In the upper cliffs (about 11 m (35') above the level of the falls lip) the Renwick Shale Member forms a distinctive marker. It consists of dark grey-weathering shales with thick, light grey, highly convoluted turbidite siltstones. The Renwick Member, about 10 m (32') thick at this location, is considered to be an important stratigraphic marker. It consists of interbedded silty, black shale which is distinctly interbedded with lenticular and often channeloid sandstone bodies which appear to be of turbiditic origin.

The highest beds, overlying the Renwick to the top of the gorge are flaggy, brownish grey-weathering siltstones of the Ithaca Formation.

Interpretation: The visible Genesee section (lower middle Genesee-to-lower Ithaca) stratigraphic succession at this falls is equivalent to approximately 2 meters (6.5') of Genesee-Penn Yan section at Cayuga Creek in eastern Erie County; the tremendous southeastward expansion of this sequence shows the influence of Catskill Delta progradation in this basin setting.

The Genesee represents a major deepening event associated with the Taghanic onlap event. This transgressive event, occurring near the end of the Middle Devonian is recognizable nearly worldwide and is almost certainly eustatic in nature. In New York it is associated with major black shale deposition indicating anoxic bottom waters. The Fir Tree and Lodi intervals appear to record brief interruptions in the anoxia allowing colonization of the sea floor by a low diversity benthic fauna including aulopodid corals and small brachiopods and mollusks, whereas the Hubbard Quarry, parts of the Sherburne and Renwick reflect a return to more dysaerobic, perhaps deeper, water conditions. Superimposed upon these fluctuations of the aerobic/anaerobic water masses was a general increase in the impact of silty turbidites, associated with progradation of the Catskill Delta.

At this locality the 6 to 7 meter (19-22') thick Fir Tree section and the overlying 20 to 22 meter (65-70') thick Hubbard Quarry shale interval are greatly expanded equivalents of far thinner deposits some 6 miles to the north (see text) and stops 2 and 3. The Fir Tree submember thickens exponentially southward from the latitude of Interlaken, King Ferry, and Genoa and undergoes a profound facies change from a compact, aulopodid-rich, concretionary limestone bed in the north to a thick splayed succession of turbiditic concretionary "ribbon" limestone beds at Taughannock Falls (Figs. 5-7, 13). South of this locality, the Fir Tree interval continues to thicken, and this unit loses its identity into Sherburne facies before this interval dips below surface view, just north of Ithaca. The Lodi interval remains relatively thin through the Seneca, Cayuga, Salmon Creek, and Owasco Valleys, though it becomes coarser and, eventually, quite shell-rich east of the Owasco Valley.

Cumulative	Miles	
11.0	0.8	Return to Route 89, turn left (north).
11.75	0.75	Bridge over Trumansburg Creek; a large waterfalls east of the bridge exposes about 4 m (13') of interbedded limestone and shale at the Fir Tree position.
	2.4	Pull-off on right overlook on Cayuga Lake, note Lake Ridge Point and Milliken Power Station on east side of lake.
14.15	1.1	Little Point Road. Little Point Creek shows about 2.4 m (7') of Fir Tree interval (slabby limestones with shale interbeds, 8 m (26') of Hubbard Quarry shale, and 3 m (10') of Lodi aulopoid bearing calcareous siltstone.
15.25	0.85	Entrance to Rocky Dock Campsite on right. Rocky Dock Creek near campsite ground shows high waterfalls capped by Sherburne siltstone. Fir Tree is 1.6 m (5.5') thick.
16.1	0.95	Bergen Beach Road on right; turn right (east) for Stop 2.
17.15	0.1	Outcrop of Sherburne siltstone in road ditch.
17.43	0.28	Pull off for Stop 2, on left side of road.

STOP 2 UNNAMED RAVINE NEAR BERGEN BEACH POINT

Location: Unnamed ravine south of Bergen Beach, 0.5 mi. west (upstream) of Bergen Beach Point, town of Covert, Seneca Co. (Trumansburg 7.5' Quadrangle).

Description: The waterfalls section on this small stream displays strata within the upper Genesee Member and the lowermost Sherburne Member of the Genesee Formation. Lower rock divisions (Windom Shale Member and Tully Formation) exposed both downstream and along the adjacent lake shore will not be examined here. This section is a greatly thinned lateral equivalent of the section exposed in the Taughannock Falls face; approximately 70 meters (215') of that section below the Lodi bed has thinned to less than half of that thickness here with most of the northward thinning taking place in the Fir Tree and Hubbard Quarry intervals (compare figures 12 and 13 of road log).

In the gently sloped floor of this creek, hard black, fissile to platy, Genesee black shale is exposed. The first prominent lip or ledge in the lower falls face, however, marks the position of the Fir Tree Limestone submember which is separated from the overlying, falls-capping basal Sherburne (Lodi) section by 8.5 meters (27') of the Hubbard Quarry Shale Submember. At this locality, the upper part of the Hubbard Quarry

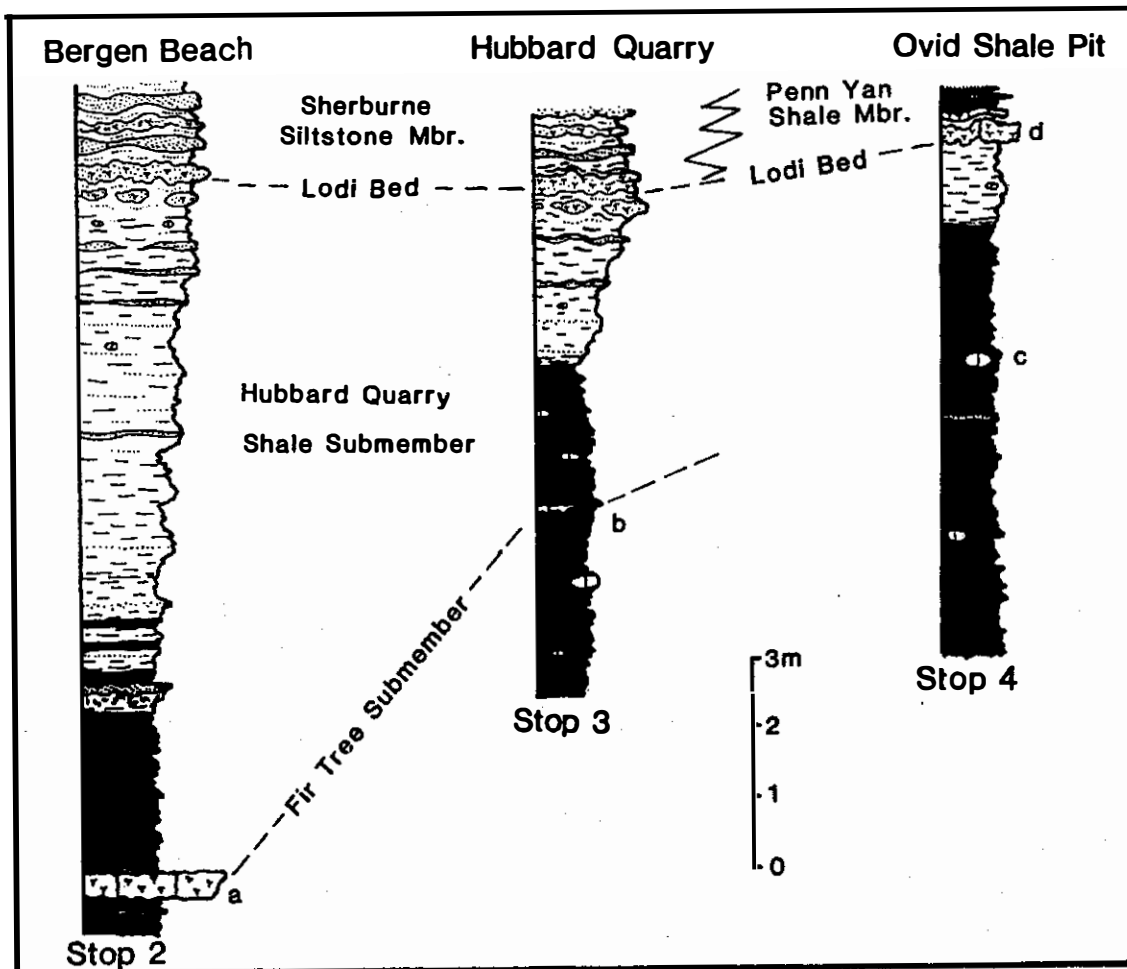


FIGURE 13.--Field trip stops 2-4: lateral changes within the Fir Tree-Lodi stratigraphic interval. Note northward truncation of Fir Tree Submember by discontinuity flooring Hubbard Quarry Submember. Units include: a, 0.5 meter-thick, aulopoid coral-rich bed of Fir Tree Limestone Submember; b, horizon of detrital pyrite and nearly completely truncated Fir Tree layers; c, concretionary horizon within undifferentiated Genesee black shale which is probably correlations with the Fir Tree Submember; d, calcareous Lodi bed, which resembles closely the Fir Tree layer at stop 2 (see discussion in text: Figs. 5-7).

submember and silty, hard, auloporid-rich strata of the Lodi Bed are inaccessible; we will observe this part of the section at STOP 3.

The Fir Tree Submember is markedly different at this section as compared with Stop 1; at Taughannock Falls and adjacent creeks the Fir Tree exceeds 5 meters (16') in thickness and consists of an alternation of dark shale and brown, concretionary to distinctly turbiditic, silty limestone beds. These sparsely fossiliferous "ribbon" limestones converge exponentially northward to form a thin, massive auloporid coral-rich impure limestone bed in this region (Figs. 6, 13). The allodapic limestone facies of the Trumansburg-Lansing area is replaced northward by increasingly fossil-rich, and bioturbated facies which is essentially identical to that of the Lodi Bed in the Seneca Valley region. The Fir Tree biota, through conspicuously auloporid coral-rich, is not really very diverse; we believe that the 0.6 meter (1.7')-thick Fir Tree Bed at this locality records an episode of reduced sediment supply and turbidity in a dysaerobic or minimally aerobic bottom setting. This would explain the absence of stereolasmatid corals, large brachiopods, and bryozoans in condensed Fir Tree deposits.

One feature of the Fir Tree Submember, however, stands out; its upper contact with black, laminated, silty shale beds of the lower Hubbard Quarry shale interval is sharp and planar. This marks the position of a submarine discontinuity, which first appears about 2.5 miles south of this locality. This break increases in magnitude northward along three separate southeast-northwest transects in the Seneca, Cayuga and Salmon Creek valleys, eventually overstepping the condensed Fir Tree deposit in each valley (see text, Stop 3: Figs. 5, 6, 13). This pattern of erosional overstep is strikingly similar to that observed at the Lodi level further to the northwest in the Seneca Lake Valley. This discontinuity, overlain by black Hubbard Quarry shale is distinctive for thin (0-0.8 cm-thick) accumulations of reworked pyrite, orbiculoide fragments, and occasional pieces of bone; this material can be skimmed up with a chisel from discontinuous debris lenses at the top of the Fir Tree Falls bench.

Above the Fir Tree Bench is a 2.5 meter (8')-thick sequence of lower Hubbard Quarry black shale facies which contains numerous Leiorhynchus quadricostata? and Orbiculoidea lodiensis on many bedding planes; above this level, the Hubbard Quarry Submember displays an upward-coarsening and upward-shallowing facies succession to the base of the Lodi Bed. Upper Hubbard Quarry beds are distinctly silty and ledge; these strata are usually variably bioturbated and diminutive molluscan material as well as occasional auloporid corals can be found at some levels.

At the top of the lower Hubbard Quarry black shale sequence, 2.5 meters (8') above the Fir Tree bench, is another erosional discontinuity which is marked by reworked pyrite. This is associated with a current-rippled siltstone layer which we informally term the Bergen Beach bed; this unit can be traced into surrounding gullies, and it may work the approximate position in this section of still a third Lodi-like unit which is developed within the Hubbard Quarry interval in the Seneca Lake Valley. Sectioned specimens of this siltstone bed at this locality show spectacular density-shape sorting of the pyrite grains into distinctive laterally-segregated lentils on the scour surface beneath the quartz silt.

17.7	0.38	Reverse route and return to NY87, turn right (north) and proceed.
18.2	0.6	Unnamed creek; here Fir Tree is 45 cm (1.5') thick and overlain by a pyrite layer up to 3 cm (1.5") thick.
18.3	0.1	Road to Interlaken on left.
18.5	0.2	South fork Interlaken Creek.
18.6	0.1	North fork Interlaken Creek; Fir Tree bed is compact 6" ledge here.
18.8	0.2	Interlaken Beach Road on right.
18.85	0.05	Lively Run Creek. Exposure of Lodi bed; Fir Tree level is inaccessible; prepare to stop.
18.95	0.1	Road into Hubbard Shale quarry (STOP 3) on right; pulloff on side of NY 89.

STOP 3 HUBBARD SHALE QUARRY

Locality: Small shale pit on east side of N.Y. Rt. 89 and about 0.3 km north of bridge over Lively Run Creek, town of Covert Seneca County, N.Y. (Sheldrake 7.5' Quadrangle).

References: Cornell Univ., 1951; deWitt and Colton, 1978; Kirchgasser, 1981; Van Diver, 1980; Baird and Brett, 1986; Hubbard Quarry Shale type section.

Description: The floor of this shale pit exposes very dark grey to black, platy shale of the Genesee Member (Genesee Formation). A layer of tabular, rusty stained carbonate concretions, containing auloporphid corals is exposed near the lowest (northeast) end of the quarry. This is the northern feather edge of the Fir Tree limestone Submember which is represented at Taughannock Falls by about 5-6 m (16'-19') of alternating tabular limestones and dark shale. Here the unit is condensed and has been almost completely truncated by submarine erosion (Fig. 13). Note the discontinuous sheet accumulation of reworked pyrite debris and orbiculoid fragments that marks the position of the discontinuity that cuts out the Fir Tree deposit. This same discontinuity can be followed locally within the undifferentiated Genesee black shale succession north past the Fir Tree bevelled margin; reworked pyrite and bone debris occurs locally, concentrated at the discontinuity position, usually as a lag concentration at the base of a bed of rippled, winnowed siltstone. Similar, nearly bevelled, Fir Tree deposits can be seen across Cayuga Lake southwest of King Ferry and in the Salmon Creek Valley just south of Genoa. This margin cannot be seen in the Seneca Lake Valley because it is below lake level north of the Fir Tree anticline, but we suspect that its position should be just south of Baskin and Caywood points on that lake.

The overlying black, fissile shale, here about 4.3 m (14') thick, has been designated the Hubbard Quarry submember (Baird and Brett, in prep.),

it contains bedding planes covered with the flattened specimens of Leiorhynchus quadricostatum and Orbiculoidea lodiensis. Two or three horizons of large septarian concretions, containing calcite and barite filled fractures, appear in the upper part of this division. The lower 2-2.5 meters (6.5-8.0') of the Hubbard Quarry Submember is a fissile to platy black shale. The upper part of this unit, below the Lodi beds, consists of grey to olive grey, silty mudstone which contains several thin siltstone beds; this interval, though poor in body fossils, is variably bioturbated suggesting deposition under dysaerobic conditions. The Bergen Beach bed of Stop 2 is not evident here.

The Hubbard Quarry shale is overlain by a nodular calcareous siltstone with very abundant auloporid corals, which is the local representation of the Lodi Submember, marking the base of the Sherburne-Penn Yan Member (Fig. 13). As noted at Taughannock, this bed lies slightly below the Givetian/Frasnian (Middle/Upper Devonian) boundary. The Lodi grades northwestward into a more prominent and more fossil-rich concretionary silty limestone bed in the vicinity of Lodi and Ovid (see STOP 4: Fig. 13) before it, too, is bevelled below a black shale tongue still farther to the northwest.

		Return to vehicles; continue north on NY 89.
19.45	0.5	North Interlaken town line.
20.25	0.8	Kidders Gully north fork. Remnant pods of Fir Tree beds found here. Hubbard Quarry Shale is 4 m (13 ft.) thick.
20.65	0.4	South fork Sheldrake Creek in large water falls east of Rt. 96, Fir Tree bed is absent, Lodi caps falls.
20.95	0.3	North fork Sheldrake Creek
21.40	0.45	Road to Sheldrake on right.
20.6	0.2	Unnamed creek with good Hamilton exposures.
22.1	0.5	Powell Creek
22.55	0.45	Groves Creek. Tully/Window exposure is below road, in small quarry-water falls area.
22.85	0.3	Barnum Creek and shale pit in Window Shale Member on left. Prepare to turn.
22.95	0.1	Junction Route 89 and Co. Rt. 138 on left; turn left.
24.75	1.8	Junction Route 96, turn right (west).

26.75	2.0	Junction Route 96/96A turn right (north) on Rt. 96.
27.75	1.0	Junction Hayt Corners Road/West Blaine Road; turn left (west) onto West Blaine Road for STOP 4.
28.50	0.75	Gravel road leading into shale pit on left (south) side of West Blaine Road; lay down cable gate and drive in to end of road.

STOP 4 OVID SHALE QUARRY

Location: Shale pit 1.2 miles northwest of Ovid, Seneca County. North-south entrance road to pit is on the south side of West Blaine road 0.75 mile west of the Rt. 96-414/west Blaine road intersection in the town of Ovid. Ovid 7.5' Quadrangle.

References: This is locality OV-16: deWitt and Colton (1978).

Description: This borrow pit is excavated mainly into hard, fissile to platy, black shale facies of the Geneseo Member; the uppermost 10 meters (30') of this unit is visible here as is the basal contact with the overlying Penn Yan Shale Member (Fig. 13). This contact corresponds to the base of the 0.4-0.6 meter (1.2-1.8')-thick Lodi limestone bed which is exposed at the top of the south and east walls of the pit. The Lodi bed is easily seen in loose blocks around the quarry; it is composed of nodular concretionary grey limestone masses with associated calcareous, silty, gray mudstone. This unit contains framestone thickets of aulopodid corals which probably partly control the form and distribution of concretions within the bed. The Lodi fauna in this region is richer than at Stop 3; the aulopodid growth is more profuse here and additional taxa, including the brachiopod Pseudoatrypa devoniana, the gastropod Palaeozygopleura as well as numerous small bivalves and pelmatozoan debris, can be collected here. The Lodi here is overlain by rusty-weathering, silty, black shale facies comprising higher Penn Yan beds.

Discussion: This section displays an uncanny resemblance to those at stops 2 and 3; at those places an aulopodid rich unit was developed within a black shale sequence much like what we see at this stop. The deception here is that the aulopodid-rich beds at these stops seem to be all one and the same. In reality, we are looking at a repetition of the Fir Tree facies motif at the horizon of the Lodi; at this locality no Fir Tree beds are present and the Lodi has changed northwestward from a non-descript, rubbly siltstone deposit at Stop 3 to a more impressive fossil-rich ledge resembling the Fir Tree layer near its northern bevelled margin (Fig. 13).

This similarity becomes even more impressive due to the fact that the Lodi bed at this stop is close to its northwestward erosional margin; the Lodi is completely absent directly across Seneca Lake in localities near Dresden. An imaginary projection of the strike of the Lodi erosional margin (Fig. 8), based on its northeast-southwest alignment near Himrod west of Seneca Lake, would place the Lodi erosional limit less than a mile northwest of this pit.

As with the post-Fir Tree discontinuity, the erosion surface above the Lodi is marked by a discontinuous lag layer of reworked material which underlies black to dark gray, transgressive facies. At this locality a one to three centimeter (0.5"-1.5")-thick layer of transported aulopoid coral debris, pelmatozoan fragments, and subordinate detrital pyrite marks this contact at the top of the Lodi. In the vicinity of Himrod, this surface is marked by a discontinuous lag accumulation of reworked pyrite with little or no associated carbonate allochems.

Three meters (10') below the Lodi Bed in this section is a band of septarial concretions near the top of a meter-thick interval of dark gray chippy shale which is not as black as Geneseo beds above and below. We believe that the sharp upper contact of this interval with laminated black shale facies corresponds to the post-Fir Tree discontinuity within the Geneseo Member proper (Fig. 13). This contact shows that what may, at first, appear to be a homogenous and continuous black shale sequence may, in actuality, be a stacked, discontinuous succession of black shale packages bracketed by numerous internal erosion surfaces.

Return to vehicles and retrace route to

29.5	1.0	Junction Route 96-414. (<u>To continue on main route:</u> turn left (north) onto Route 96-414 and proceed for 1.6 miles to the junction of Route 96/Route 414 and turn right (north) on Route 414; then continue north for 0.45 miles to junction of Bromka Road where log for Optional Stop ends; route continues from that point. <u>To go to the optional stop:</u> follow the log below).
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LOG FOR OPTIONAL STOP

0.0	0.0	Go straight across intersection of 96-414 staying on West Blaine Rd. which becomes Hayt Corners Road.
4.60	1.6	Hayt Corners; continue straight eastward.
2.60	1.0	Hayt Corners Road bends sharply to the left (north); follow.
3.75	0.75	Intersection Stout Road; continue straight (Hayt Corners Road changes name to Iron Bridge Road.)
3.60	0.25	Road bends sharply to the left (west).
3.7	0.1	Road bends sharply to the right (north).

- | | | |
|-----|-----|--|
| 3.8 | 0.1 | Cross first unnamed creek. |
| 4.4 | 0.6 | Cross second unnamed creek; prepare to turn right into small gravel road to shale pit adjacent to creek. |

- OPTIONAL STOP. POPLAR BEACH SHALE PIT

Location: Shale borrow pit bordering the east side of Iron Bridge Road 0.7 mile southwest of Poplar Beach, Town of Romulus, Seneca County; pit also borders north branch of an unnamed creek which flows into Poplar Beach. Ovid 7.5' Quadrangle.

References: Baird, 1981.

Description: This shale pit is developed in medium to dark gray, chippy mudstone facies of the Wanakah (=King Ferry) Shale Member (Ludlowville Formation: Hamilton Group). This shale interval is sparsely fossiliferous and nearly homogeneous, with the sole exception of a layer of reworked calcareous hiatus-concretions which can be easily sampled on the service road leading into the pit and from the upper parts of the shale banks within the pit.

Discussion: This stop is included to show the great difference between submarine discontinuities developed under anaerobic or minimally dysaerobic conditions and those produced in aerobic settings. This discontinuity is one of two very similar horizons which were mapped within the King Ferry Member as part of a study of mechanical and bioerosional processes on a regionally sloped, mud-floored, Devonian sea floor (for details see Baird, 1981; Baird and Brett, 1981: N.Y.S.G.A. Binghamton Meeting: Discussion on road log). At this locality, we merely wish to emphasize that infaunal organisms play a disproportionately greater role in the erosion process on oxygenated substrates owing mainly to their burrowing and boring activity. It should be noted also that burrowing animals may erase sedimentary discontinuities in unconsolidated or minimally consolidated muds; this commingling or thorough mixing of the older and younger sediments changes the discontinuity from a discrete contact into a stratomictic interval of scrambled mud, shells prefossilized debris, and reworked concretions or small nodules.

At this locality, the level of hiatus-concretions and associated shell hash (Barnum Creek Bed) marks the position of a stratomictic erosional discontinuity; there is no obvious erosion surface associated with these nodules-rather the hiatus-concretions almost "float" in a slush of burrowed shell debris and mud. This is a good collecting stop for the bored and bioencrusted hiatus-concretions. Epizoans on the nodules include auloporid corals, the rugose coral Stereolasma, ctenostome bryozoans, and occasional pelmatozoan holdfasts. Distinctive flask-shaped borings are abundant on some nodules. Scratch marks on nodule exteriors, produced by burrowers colliding with still-buried concretions, are commonly found at this locality.

Return to Iron Bridge Road and turn right (north).

4.75	0.35	T-junction with Swick Road, turn left (west).
5.70	0.95	T-junction with Marsh Corner Road, turn right (north).
5.75	0.05	Cross Big Hollow Creek.
5.95	0.2	Junction Bromka Road, on left, turn left (west).
Total		
7.60	1.65	Junction Route 414; turn right (north) and continue on main route.
		END OF OPTIONAL STOP ROUTE
		(Add 7.6 miles to route for extra stop)

30.55 (36.1)	2.05	Junction Bromka Road (end of optional stop route).
35.7	5.15	Junction Rt. 336; Peter Witmer historic farm.
36.1	0.4	Village of Fayette, cross Poorman Road; (a turn to left leads to Fayette town shale pit, a well known Centerfield/Levanna section 0.3 miles west of this junction). Continue north on Rt. 414.
38.5	2.4	Junction Yellow Tavern Road (Co. Rt. 121) turn right.
38.9	0.4	Note stone farm house on left.
39.1	0.2	Road bears right.
39.8	0.7	Entrance to Warren Brothers, Seneca Stone Quarry (STOP 5), on left; turn left into yard by office to obtain clearance to enter.
40.3	0.5	Proceed into quarry on gravel road, turn left on side spur; park near dump piles on south berm of quarry.

STOP 5 SENECA STONE QUARRY

Locality: Large quarry of Warren Brothers Co., 4 km SSE of Seneca Falls, just north of Yellow Tavern Road 1.5 km west of Canoga Springs, Seneca Co., N.Y. (Romulus 7.5' Quad.).

References: Oliver (1983); Rickard (1984); Conkin & Conkin (1984).

Description of Units: This active quarry displays an outstandingly complete sequence from the base of the Oriskany sandstone, through the complete Onondaga Formation and to the base of the overlying Marcellus Formation of the Hamilton Group. Oriskany Sandstone and its unconformable

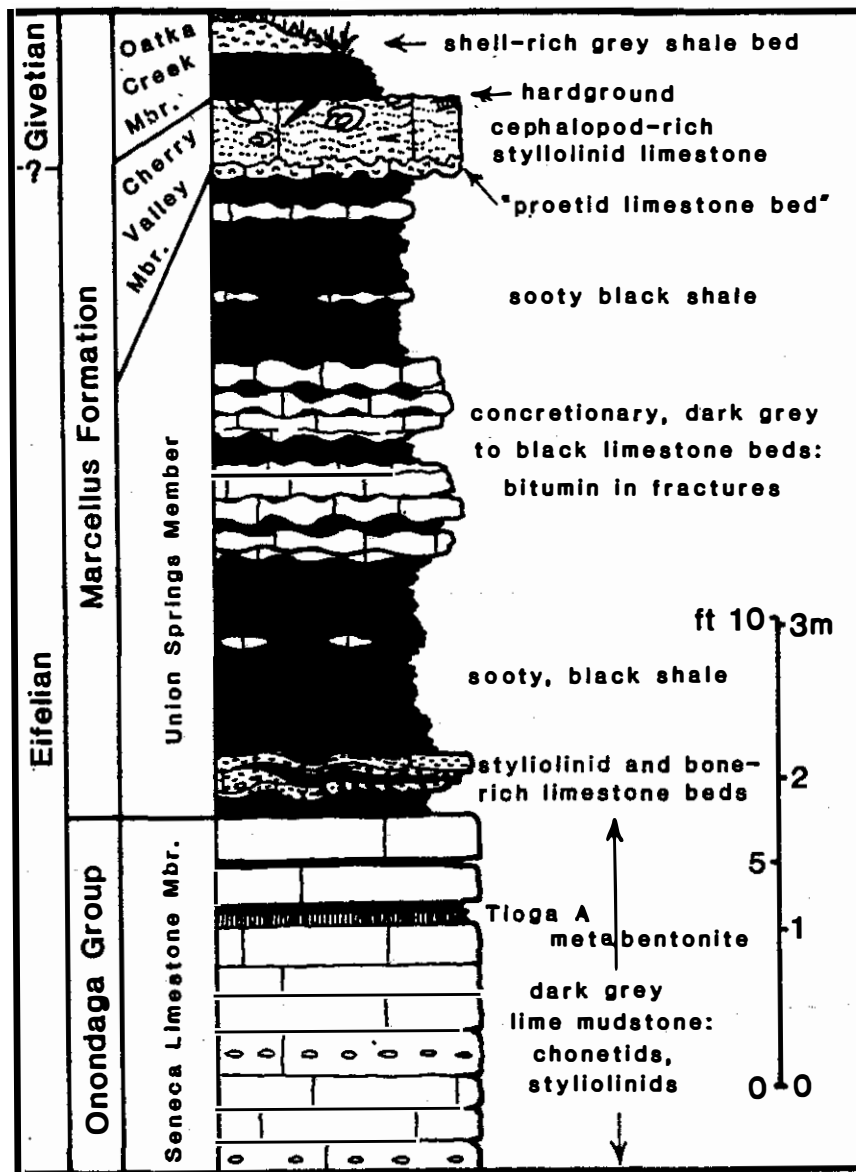


FIGURE 14.--Uppermost Onondaga and lower Marcellus strata at the Warren Brothers Seneca Stone Quarry. Key units and beds shown in figure are discussed in text. Note prominent bone-rich bed ("Bone Bed 7" of Conkin and Conkin, 1984) in the basal Union Springs Shale and conspicuous corrasional hardground on the top of the Cherry Valley Limestone Member.

contacts can be observed in the floor of the quarry; four members of the Onondaga, including ash layers are variably well exposed in the east wall of the quarry along an access road ramp. The lower Hamilton sequence including (in 1986) an excellent dip slope exposure of the top of the Cherry Valley Limestone, is well exposed on the south rim of the quarry. Considerable structure is visible in the walls of the quarry including minor gentle folding and a north-directed thrust fault which places Seneca Member over Marcellus black shales on the southeast side of the quarry. Our attention here is directed to the uppermost beds in the quarry, though many features within lower quarry units are worth the effort of a return trip.

Uppermost Onondaga and Marcellus Stratigraphic Units

Onondaga Group: Seneca Limestone Member

The highest division of the Onondaga Limestone includes 8.5 m (26') of dark gray to nearly black, micritic limestone which is separated from the underlying Moorehouse Member by a distinctive, 20 cm (8")-thick, cream-colored soft clay layer which is the expression of the middle Tioga ash bed which is termed the Tioga B layer (Rickard, 1984) or the Onondaga Indian Nation Metabentonite using terminology of Conkin and Conkin, 1984. The Seneca Member is a moderately fossiliferous wackestone with some chert nodules, chonetid brachiopods, dalmanitid trilobites, and gastropods in its lower part, but it is distinctly, darker, more argillaceous, and sparser in macrofossils towards the top; in the uppermost beds only a meager dysaerobic biota of Styliolina shells, diminutive chonetids, and the burrows Planolites and Chondrites is observed. The Seneca Member appears to grade upward almost continuously into the overlying Union Springs Member of the Marcellus Formation. The overall Moorehouse to Union Springs stratigraphic succession records a major transgression in this region; Moorehouse and lowermost Seneca aerobic facies is succeeded by higher dysaerobic Seneca beds which, in turn, pass upward into minimally dysaerobic to anaerobic Union Springs deposits (Figs. 3, 14). This transgression was probably eustatic in nature (see Johnson et al., 1985) and timed with the introduction of a considerable amount of siliciclastic mud (although much of the Union Springs is still carbonate rich and technically a very argillaceous, black limestone). In most areas of New York both east and west of the Cayuga Lake meridian, the Onondaga-Marcellus contact is unconformable and probably erosional, but here, near the presumed basin center it appears conformable. The overlying Union Springs muds reflect major uplift of Acadian source terranes, and the tectonism may have been propagated westward to form minor diastrophic uplift of the Onondaga, prior to subsidence.

About 62 cm (2') below the top of the Seneca facies succession is a second, yellow-weathering ash layer which is designated as the Tioga A bed (Rickard, 1984) or Tioga "sensu stricto" (Conkin and Conkin, 1984); on fresh, backpiled slab heaps along the quarry rim, one can often find crumbly, micaceous pieces of this ash mixed in with the other darker rock fragments. Both the Tioga A and B ash beds produce distinctive rust-staining on the quarry walls due to weathering and their levels can be seen easily from the quarry rim.

Hamilton GroupMarcellus FormationUnion Springs Shale Member

The Seneca Member is overlain by about 4.3 meters (13') of sooty, black, fissile shale which contains numerous brown-black, concretionary limestone beds, as well as thin Styliolina packstone layers near the top and base of the unit (Fig. 14). The biota of this unit is of very low diversity; the 1mm-long conical shells of Styliolina form the main component of tractional hash layers, and several other taxa including the brachiopods Ambocoelia and Leiorhynchus, the bivalve Pterochaenia, and cephalopods occur only sparingly in this interval. The Union Springs is distinctly sooty and fractures, along with the numerous joint and fracture surfaces within limestone beds, are filled with pyrobitumen. Thin bone-rich beds occur within a 15-20 cm (6"-8")-thick interval of Styliolina packstone layers just above the base of the Union Springs. This layer, designated "Bone bed no. 7" by Conkin and Conkin (1984) is a widespread key marker in the Union Springs Member; we have traced this unit from the Jamesville Quarry near Syracuse to the General Crushed Stone ("Five Points") quarry northwest of Lima, Livingston County (Fig. 3). Locally, this bed displays spectacular concentrations of onychodid teeth, placoderm armor, and spines, as can be seen at the Jamesville quarry near Nedrow. Bone beds such as this mark diastems and possibly even larger discontinuities within the black shale; this pattern is very similar to the style of submarine erosion which we observed in the Genesee Formation. Back-piled slabs of this bone-rich layer can be observed at this stop. The seemingly gradational nature of the Seneca-Union Springs contact in this region presents some difficulties in defining the top-Onondaga contact; Rickard (1984), places the Seneca-Union Springs boundary at the base of the upper ash bed, while Conkin and Conkin (1984), place it at a bone bed slightly below the ash. Neither of these layers truly coincide with the level where sooty Union Springs shale first predominates, further adding to the complexity of this problem.

Proetid Limestone Bed and Cherry Valley Limestone Member

Above the Union Springs is a distinctive, highly condensed limestone unit which can be traced in outcrop from the vicinity of Lima, Livingston County all the way into the Hudson Valley. At this locality this unit is a compact, 65 cm (2.2')-thick ledge which is actually composed of three amalgamated limestone subdivisions, two of which ("Proetid bed" and Cherry Valley Member, proper) are extraordinarily widespread. The basal few centimeters of the amalgamated ledge consists of a concretionary underbed which is actually a part of the Union Springs Member. This layer consists of large, brown, septarial concretions, which can be seen on large overturned blocks. Debris from the overlying "Proetid Bed" fills numerous cracks and solution pits which locally penetrate through the underbed.

"Proetid Limestone Bed": The next amalgamated division within the limestone ledge is a highly irregular, 5-15 cm (2-7")-thick

white-weathering, light grey, micritic unit which is sandwiched between the brown concretionary layer and the true Cherry Valley Limestone (Fig. 14). This white horizon contains scattered fossils, particularly, a few atrypid and rhynchonellid brachiopods, comminuted crinoid debris and calyces of the minute crinoid Haplocrinites clio, rare small rugose corals and proetid trilobites. Overall, this bed closely resembles a layer, informally termed the "Proetid bed" (Rickard, 1952). In the east the proetid bed is separated from the overlying Cherry Valley Limestone by 0.1 to 3 m of black shale and thus it represents a distinct carbonate unit. It appears to merge with the base of the Cherry Valley Limestone near the Chenango Valley due to pinch out and/or erosion of the intervening shale. At Seneca Quarry the top of the proetid bed is an irregularly sculptured hardground with a relief or nearly 10 cm. Darker grey matrix of Cherry Valley lithology infills pockets including undercut cavities on this limestone, indicating a period of induration and erosion of the surface prior to Cherry Valley deposition.

Cherry Valley Member: The upper two thirds of the carbonate ledge comprises the true Cherry Valley limestone, it is a dark grey, slightly pyritic, styliolinid and crinoidal packstone with abundant orthoconic nautiloids and large goniatites (Agoniatites). These cephalopods are particularly well displayed on the upper surface of the bed. Along the southwest rim of the quarry this upper surface has been glacially polished and striated; however on the southern edge it was covered by a thin veneer of overlying Oatka Creek shale. This shale has been bulldozed off on a large, gently-dipping surface on the south edge of the quarry, exposing the unweathered upper contact of the Cherry Valley bed. This surface is nearly planar and displays a large number of erosionally truncated nautiloids, goniatites and a few large fish bones. In some cases the bevelled edges of the phragmocones and the chambers on this submarine hardground are encrusted with drusy crystalline pyrite.

This appears to represent a regionally traceable cephalopod bed. Nautiloids are well preserved and show a prominent and highly significant northwest to southeast orientation. This seems to indicate that the cephalopods died shortly before original burial and were oriented by basinward-flowing currents on the sea floor. Very few specimens are vertically oriented.

Beveling of the fossils almost certainly took place on the Devonian sea floor after induration of the limestone. Examination of nearby sections indicates that a few centimeters of overlying Cherry Valley Limestone were removed from this surface exposing the prefossilized cephalopods. Erosional mechanisms probably included mechanical scour by traction sheets of residual debris and shells moved by currents (abrasion) and some component of submarine dissolution under anaerobic conditions prior to burial of this surface by Oatka Creek black muds (Baird and Brett, in prep.); evidence of this corrasion process becomes even more pronounced on this surface, further west at Flint Creek near Phelps, Ontario County and at LeRoy, Genesee County, where the discontinuity surface becomes distinctly irregular and pitted (see Fig. 3).

As noted for other submarine erosion surfaces (see discussion of Moonshine Falls), this erosive beveling apparently accompanied sediment

starvation associated with rapid transgression. The Cherry Valley is sharply overlain by black shale of the Oatka Creek-Chittenango members which records anaerobic conditions following Cherry Valley deposition.

		Return to office area, turn left (east) and proceed on Co. Rt. 121.
41.1	0.8	Sharp turn in road.
41.4	0.3	Bend in road at Canoga Springs.
42.4	1.0	Junction NY 89 Canoga Village; turn left (north) onto NY 89.
42.6	0.2	Leave Canoga Village
45.5	2.9	Entrance to Cayuga Lake State Park on right.
46.5	1.0	Defunct Eisenhower College on left.
47.9	1.4	Note raised stretch of road which follows a conspicuous narrow ridge.
51.85	3.95	Junction Rt. 89 and U.S. 20, turn right, (east) onto U.S. 20.
53.5	1.65	Entrance to Montezuma Wildlife Refuge; bridge over Cayuga Lake outlet.
53.9	0.4	Junction U.S. 20/Route 90; turn right (south) onto Route 90.
54.0	0.1	Slumped old roadcut in Camillus Shale (Upper Silurian).
57.4	3.4	Town of Cayuga.
58.3	0.9	Leave town of Cayuga.
63.9	5.6	Enter town of Union Springs
65.6	1.6	Leave town of Union Springs.
65.65	0.15	Woods Quarry, behind large garage building on left; has exposures of the Onondaga-Marcellus (Union Springs) contact; the same upper Tioga ash layer and fish bone bed seen in Seneca Quarry are present here.
67.25	1.56	Bridge over Great Gully; exposure of Mottville Limestone (Marcellus/Skanateles contact) visible in low falls east of road.

67.65	0.4	Bridge over Criss Creek; more Mottville exposure (visible from road during cold months)
68.95	1.3	Village of Levanna
69.5	.25-.55	Bluffs of Levanna along Cayuga Lake shore.
70.5	.55	Mouth of Dean Creek, base of hill going up into Aurora.
70.15	0.1	Enter town of Aurora.
71.45	1.3	Wells College on left.
71.8	.35	Junction NY 90/Poplar Ridge Road, turn left (east) onto Poplar Ridge Road.
72.55	0.75	Junction Fry Road turn right (south).
73.1	0.55	Junction Moonshine Road (gravel road) on right; turn right (south).
73.2	0.20	Bridge over Paines Creek (STOP 6). Park on left side of road just past bridge.

STOP 6 PAINES CREEK; MOONSHINE FALLS

Locality: Exposures on the bed of Paines Creek above Moonshine Falls 0.3 km west of Moonshine Road, south of Aurora, Cayuga Co., NY.

References: Cooper (1930), Grasso (1970); Kramers (1979); Gray (1984).

Description of Units: Paines Creek flows north westward on the dip slope between the bridge at Moonshine Road and the 12 meter (40')-high waterfalls, exposing extensive bedding planes of the upper Centerfield and the basal Ledyard shale. Strongly jointed, dark, fissile Ledyard shale crops out beneath the bridge and for 100 m above the waterfall. This contact and most of the Centerfield Member are re-exposed still further 400 m upstream from the bridge at the crest of a small anticline. The face of the waterfalls (largely inaccessible) exposes the upper 2 meters (6.5') of the black Butternut-equivalent part of the Levanna Shale Member, a disconformity at the top of the Levanna, and 10.7 m (35') of the lower Centerfield. This sequence can be examined by descending into the gorge along a small tributary gully north of the crest of Moonshine Falls.

Centerfield Member: A fossil hash bed containing scattered hiatus concretions, encrusted by bryozoans and aulopodid corals, forms the base of the Centerfield and rests sharply on black, fissile Leiorhynchus-bearing shales of the Levanna. Overlying Centerfield gray shales contain small brachiopods (Ambocoelia, chonetids) and, at two levels, biostromes of aulopodid corals. Higher beds show a transition to gray, calcareous mudstone with larger brachiopods, particularly Tropidoleptus, and a few larger rugose corals.

The highest beds of the Centerfield, exposed in the creek floor above

the lip of the waterfalls, are hard, light blue-grey, argillaceous limestones or very calcareous mudstones which are thoroughly bioturbated and display well-preserved large Zoophycos spreiten, typically with limonitic (originally pyritic) marginal tubes. Body fossils are abundant and very diverse but patchy in distribution. On the strongly jointed creek bed just above the waterfall are excellently exposed patches (biostromes) of ramose and foliose (fistuliporoid) bryozoans, some of which have associated brachiopods (rhynchonellids, Nucleospira, Elita, Vitulina as well as proetid and Phacops trilobites. Excellently preserved blastoids and crinoids are common in some bryozoan thickets at this level. Scattered large corals including solitary forms such as (Heliophyllum) and colonial taxa (Eridophyllum and Favosites) are also present.

Many localities show some 3 to 4 m of highly fossiliferous softer gray shales overlying the calcareous mudstone capping unit of Moonshine Falls which mark a transgressive facies transition into post-Centerfield deposits; however at this one locality these transitional beds are absent, apparently because of localized erosional truncation (Fig. 4). The upper contact of the Centerfield is disconformable here, as at adjacent localities, but the contact is very striking here owing to the lack of the sharply overlying black Ledyard shales. At the contact is a thin (1-2 cm) lag of crinoid debris with abundant fossils, derived from the underlying Centerfield; many of these are distinctly broken and corroded. This bed also contains abundant small phosphatic nodules, some of which are reworked fossil steinkerns (e.g. of conulariids and enrolled trilobites), fish bones and hiatus-concretions. Reworked, pyritized burrow tubes have been exhumed locally on this surface at some localities. However, where this has been observed, the tubes occur stratomictically commingled with shell hash in burrowed mudstone; these tubes, thus, show evidence of breakage and reorientation by infauna but no evidence of lateral current transport.

This lag bed, termed the Moonshine Falls Bed by Gray (1984) is an important widespread marker, which has yielded the conodont Polygnathus timorensis, diagnostic of the lower P. varcus zone of the Givetian stage; it is on the basis of these conodonts from this locality that Klapper (1981) has assigned the Centerfield to the base of the varcus zone; in actuality, no diagnostic conodonts have been obtained from the Centerfield proper, or from the underlying Skaneateles Formation.

Ledyard Shale: The basal 4 m of the Ledyard Member are exposed here, and this creek is the type section of that member. The lower beds are very dark gray to black, fissile shales with a Leiorhynchus-Styliolina fauna. A few thin stringers of crinoidal and phosphatic debris have been observed in the basal 10 cm of the Ledyard, these apparently have been reworked from lag debris overlying the erosional top of the Centerfield.

Interpretation: The Levanna-to-Centerfield sequence represents an abrupt change from deeper water black Leiorhynchus facies to fossil-rich calcareous mudrock, probably deposited in shallow water close to normal wave base, and within storm wave base. Although the sequence is gradational east of Syracuse meridian, here it is broken by a discontinuity that marks the sharp base of the Centerfield, Gray (1984) has interpreted this interruption in sedimentation to be the result of

local tectonic uplift of the sea floor near the Tully Valley. This updoming could have cut off sediment influx to the west, leading to slight erosion of the sea floor and condensation of shelly debris. Above this basal bed the Centerfield shows a gradual shallowing-upward sequence to the beds that cap Moonshine Falls; above this hard mudstone, in other localities there is a reversal and gradual transition back to deeper water facies (Fig. 4). However, again, the sequence was interrupted in this area by an abrupt jump to dark gray and black Leiorhynchus facies. As noted above, the upper Centerfield transitional shales have been removed, apparently by erosion associated with this interruption in sedimentation. Detailed mapping of the upper contact (Gray, 1984) proved that the transitional shales are present within 2 miles on either side of Moonshine Falls (i.e. both northeast and southwest of here). This further suggests that the erosion was localized (Fig. 4); the area of enhanced erosion may be in a channel-like depression no more than a mile across and perhaps oriented NW-SE, normal to the regional paleoslope. This erosion probably took place during an interval of general sediment starvation associated with rapid transgression at the onset of Ledyard deposition. We have observed similar erosional furrowing beneath blackshales at several other stratigraphic levels. That relatively strong currents persisted, episodically, even after the onset of black mud deposition, is indicated by the stringers of reworked Centerfield debris that occur just above the base of the black shale. Evidently, some dysaerobic settings were not entirely characterized by low energy.

Continue southwest on Moonshine Road.

73.6	0.3	Junction NY 90; turn left (south).
74.75	1.15	Junction Lake Road (to Stony Point campground) (stay on 90).
75.3	0.55	Junction Ledyard/Black Rock Road (stay on 90)
78.9	3.6	Triangle Diner; Stay on Route 90 going east. Junction Route 90/Route 34B. Turn right (south).
82.4	3.5	Enter Lake Ridge.
82.8	0.4	Cross Lake Ridge Point Ravine; here the Fir Tree Limestone is about 7' thick.
84.65	1.85	Pass Lansing Fire Station.
87.55	2.9	Begin descent into Salmon Creek Valley
87.95	0.4	Road to Ludlowville on left (stay on 34B).
88.75	0.8	Cross Salmon Creek; excellent upper Hamilton exposures are present in this creek.
88.95	0.2	Road to Meyer Point at base of long hill (stay on 34B).

89.75	0.8	Enter town of South Lansing.
89.95	0.2	Portland Point Road on right (stay on 34).
90.65	0.67	Junction Route 34B/Route 34, village of South Lansing. Turn right (south) following 34.
91.65	1.0	Upper end of Shurger Glen.
93.9	2.25	Start descent into Cayuga Valley (long down grade).
94.4	0.5	Exposures of upper Sherburne turbiditic siltstone in intermittent road cuts on left; long grade is nearly on dip slope of south limb of Fir Tree anticline; large lake bank section of upper Geneseo, Fir Tree equivalent siltstones, and Hubbard Quarry turbidites and shales is below west of road.
95.4	1.0	Base of long grade at lakeshore; twin glens expose Lodi beds near road level.
95.9	0.5	Junction Route 34/Route 13; large cuts on 13 are into lower medial divisions (Renwick-Ithaca mbrs.) of the upper Genesee Formation; city of Ithaca. (END OF ROAD LOG).

SEDIMENTARY CYCLES AND LATERAL FACIES GRADIENTS ACROSS
A MIDDLE DEVONIAN SHELF-TO-BASIN RAMP
LUDLOWVILLE FORMATION, CAYUGA BASIN

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INTRODUCTION

The Middle Devonian (Givetian) Hamilton Group of western and central New York State constitutes an eastwardly-thickening wedge of black and grey shales, mudstones and siltstones with thin, but widespread, carbonate and sandstone units. These sediments were deposited in the northern end of the stratified Appalachian Basin following early phases of the Acadian Orogeny (Ettensohn, 1985; Kent, 1985). Hamilton facies occur in distinct cyclic motifs of two sorts: subsymmetrical cycles centered on maximally-regressive, limestone facies in western New York, and upward-coarsening, dark shale-siltstone-sandstone hemicycles capped by maximally regressive storm-winnowed deposits in the eastern Finger Lakes region. Recent detailed study of Hamilton stratigraphy and facies distribution has demonstrated that the two types of cycles are correlative and intergradational, and that the carbonate rich beds in the subsymmetrical cycles are the direct lateral equivalent of the tops of the coarsening-upward hemicycles (Brett and Baird, 1985; Gray, 1984; Grasso, 1986); both motifs appear to be manifestations of large scale regressive-transgressive cycles and both carbonates and sandstones record relatively shallow water near wave base facies. This work further permits the development of models explaining the distribution of recurrent fossils associations - or biofacies along environmental gradients.

The Cayuga Lake Valley presents a unique opportunity to examine facies relationships within the upper Hamilton Group. This 42 mile-long valley produces a major southward deflection in the Hamilton outcrop belt allowing examination of strata well south of the prevailing east-west outcrop trend limit. Moreover, the effect is enhanced by the exposure of upper Hamilton beds along the crest of the Fir Tree anticline near the southern end of the lake. This structure, in effect, provides a "window" into the facies of the Upper Ludlowville and Moscow Formations in the Southern Tier, some 30 miles south of the main outcrop belt. Still more important is the fact that the present day Cayuga Valley appears to obliquely cross cut northeast/southwest-trending facies belts in the upper

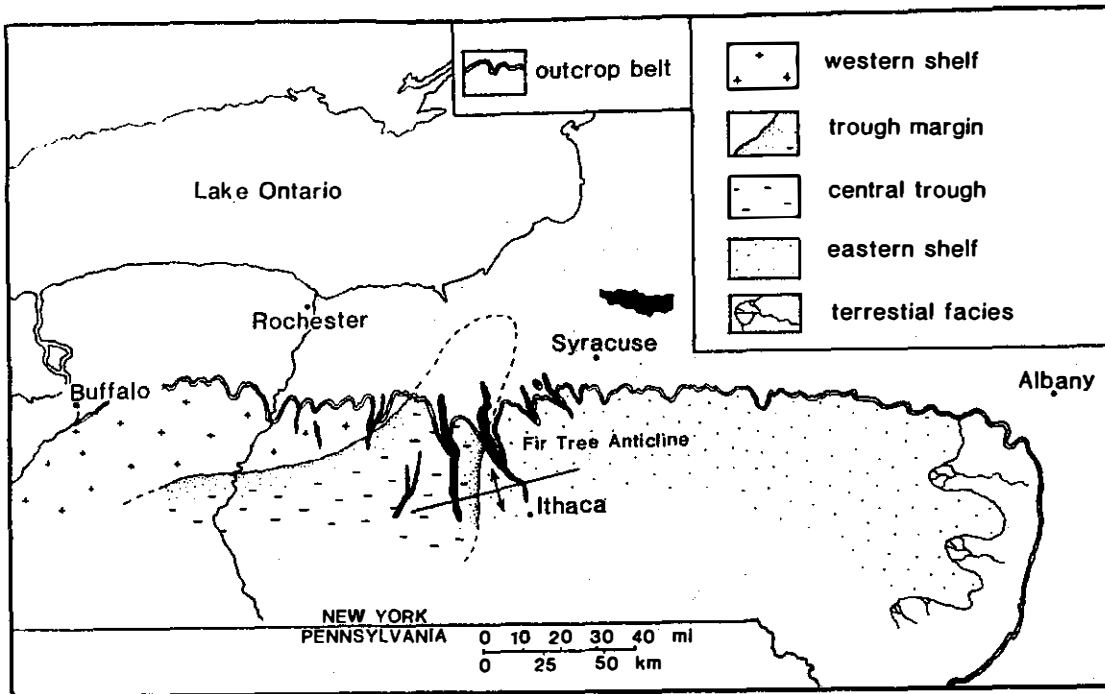


FIGURE 1.--Inferred paleogeography of the northern Appalachian Basin during Middle Devonian time showing northeast/southwest trending margin of the subsiding basin trough in the Cayuga Valley; note position of Fir Tree Anticline. Modified from Brett and Baird (1985).

Hamilton Group, which parallel the southeastern margin of the Appalachian Basin as it existed in Middle Devonian (Givetian) time (Fig. 1). Hence, unlike the exposures in the Seneca Valley, which display relatively little facies change along another extensive north-south transect (probably because this valley more nearly coincides with depositional strike), those of the Cayuga region exhibit marked changes in litho- and biofacies from north to south. Along the east side of the Cayuga Valley from Aurora, south to Lansing, one observes a progression from facies and stratigraphic successions resembling those of western New York into coarser clastic equivalents identical to those of the central New York region. Across a ten mile distance southward along the valley changes occur that parallel those seen 20 to 30 miles east of the Cayuga Lake meridian on the main (northern) east-west outcrop belt. This pattern indicates that facies strike in this area was approximately north-northeast to south-southwest, essentially transverse to the trend of Cayuga Lake.

The facies pattern in the Ludlowville Formation of the Cayuga and adjacent Owasco Valleys indicates that these environmental belts were aligned along a gentle northwest-dipping paleoslope or ramp that bordered a trough-like area of more active subsidence near the present-day Seneca Lake (Fig. 1). Lines of evidence supporting this inference include consistent northwestward changes of a) medium, brownish grey, silty mudstone facies into dark grey or black, fossiliferous shales; b) packstones or grainstone lithologies in carbonates to argillaceous

wackestones; and c) differential northwestward thickening of several units. This suggests that the area of maximum subsidence and/or depocenter of the northern end of the Appalachian Basin lay in the Cayuga to Seneca region during much of Ludlowville deposition. However, facies and isopach patterns also imply that the precise area of deepest water (axis of subsidence) and the area of maximum thickness (depocenter) were not always exactly coincident; typically the depocenter lay to the southeast of the axis of subsidence. Moreover, neither depocenter nor deepest basin axis were fixed but, rather, both tended to shift northwestward through deposition of the Ludlowville and Moscow formations. This basin migration strongly influenced the lateral facies transitions observed at different levels. For example, along the northern parts of the outcrop belt, lower Ludlowville deposits display a major black to grey shale facies change between Owasco and Skaneateles Lakes, while in the middle Ludlowville (primary focus of this field trip) most abrupt facies and thickness change occurs between Cayuga and Owasco Lakes and, for the higher Ludlowville, this area of rapid change is displaced to the Cayuga-Seneca region.

This pattern persists throughout deposition of the overlying Moscow Formation, in which depocenters and deepest-water facies of successive members are displaced westward. This observation is consistent with a model proposed by Ettensohn (1985) for westward migration of the Devonian basin axis due primarily to Acadian tectonic thrust-loading and secondarily due to sediment wedge loading near the basin center. Superimposed on this pattern is a separate effect relating to sediment distribution. It appears that during transgressions the depocenter shifted southeastward away from the axis of subsidence due to the increased distance to source areas (flooding of shoreline areas) and the overall deepening of water which allowed more sediment to accumulate over more slowly subsiding shelf areas. This reduction in sediment supply resulted in accumulation of thin dark shale sequences in the sediment-starved basin center area and thicker, silty mudstones on the southeastern shelf. Conversely, during regressions, sediment tended to prograde northwestward toward the basin center and fine grained sediments tended to bypass the now-shallow shelf regions and to accumulate rapidly in the low energy trough, leaving winnowed sand or carbonate deposits on the shelves. The net result of these processes is that the basin axis and depocenter tended to nearly coincide during times of overall regression but to be offset from one another during deepening episodes; this effect is most directly the result of an energy threshold controlled by storm and/or normal wave base but it is also controlled by changes in sediment supply. Understandably, this complex pattern of facies change has led to nomenclatural problems, as noted below.

The primary focus of this fieldtrip is the examination of patterns of facies change expressed both vertically in the form of transgressive-regressive cycles and laterally within cycles as viewed north-to-south in the Cayuga-Owasco interlake region. To this end, we will proceed sequentially through an outcrop sequence up the inferred Ludlowville paleoslope; we will examine a regional facies spectrum, starting with basinal dark grey sequences and ending with shallow sublittoral

fossil-rich sandstones. This will allow qualitative discussion of numerous biological and sedimentological changes that can be observed along this Paleozoic submarine ramp.

STRATIGRAPHY OF THE LUDLOWVILLE FORMATION IN
THE CAYUGA LAKE REGION

General Background

Both the type section and the later proposed reference section of the Ludlowville Formation lie on the east side of Cayuga Lake. Hall (1839, p. 298) proposed the name Ludlowville for the shaley sequence between the base of what is now named Centerfield Member and "the Encrinal limestone" (now termed Tichenor or basal Portland Point Member, see Baird, 1979). Hall designated Salmon Creek in the town of Ludlowville, near the southern end of Cayuga Lake as the type section for largely aesthetic reasons. At that time he believed, incorrectly, that these beds were coeval with the Silurian Ludlow Formation and evidently felt it was an interesting and fortuitous twist that the supposedly equivalent strata cropped out at Ludlowville, New York. Ironically, however, the Salmon Creek section is very incomplete and atypical as only the upper 50' (~25%) of the Ludlowville Formation is exposed there. Consequently, Cooper (1930) designated a reference section on Paines Creek in the town of Aurora about 15 miles north of Ludlowville. This is also a poor type section in that structural complications give rise to uncertainties in measurements, particularly in the Ledyard Member. However, at least this creek exposes the entire sequence which in this area is about 80 m (250 ft) thick.

Cooper, 1930		Baird, 1979		Smith, 1935
MOSCOW FM.	Windom Shale	MOSCOW FM.	Windom Shale	
	Kaahong Shale		Kashong Shale (RC Bed)	
	RC Bed Menteth Ls.		Portland Point Ls.	
LUDLOWVILLE FM.	Deep Run Sh.	LUDLOWVILLE	Menteth Ls.	Portland Point Ls.
	Tichenor Ls.		Portland Point Ls.	
	Wanakah Shale		Tichenor Ls.	King Ferry
			Jaycox Sh.	King Ferry Shale
	Ledyard Shale		Wanakah Shale	Ivy Point Siltstone
Centerfield Ls.	Ledyard Shale	Otisco Shale		
SKAN.	Levanna Shale	SKAN.	Centerfield Ls.	Centerfield
			Levanna Shale	

FIGURE 2.--Earlier stratigraphic terminology for the Ludlowville and Moscow Formations in the Cayuga to Skaneateles Lake area.

Detailed subdivision of the Ludlowville Formation into Members was formalized by Cooper (1930), who included, in western New York, five Members, in ascending order: Centerfield limestone and calcareous shale (Clarke, 1903) Ledyard black and dark grey shale, Wanakah grey shale (Grabau, 1917), Tichenor Limestone (Clarke, 1903) and Deep Run Shale. Cooper designated the upper contact of the Ludlowville at the base of the Menteth Limestone which he believed to be stratigraphically equivalent to the base of the Portland Point Member (Hall's "encrinal" bed) at the type area near Ludlowville in the Cayuga Valley (Fig.2). Cooper was unable to recognize the distinct Wanakah, Tichenor and Deep Run Members at Cayuga Lake and he proposed the name King Ferry shale for the supposedly equivalent interval between the Ledyard and Portland Point Members. Later work by Baird (1979), however, demonstrated that the Portland Point Member is, in fact, a condensation of the Tichenor-Deep Run-Menteth interval (see also Baird and Brett, 1981) and reset the base of the Moscow Formation to the base of the Tichenor Limestone (or basal "Portland Point limestone") throughout western and central New York. Baird further proposed a new unit, the Jaycox Member, for a calcareous, fossiliferous, shale at the top of the Ludlowville Formation, between a unique limestone (subsequently designated the Hills Gulch bed by Kloc (1983), and the erosional basal contact of the Tichenor Limestone, in western New York.

In this revision, one unit, the King Ferry Member was more or less left in limbo. Cooper's King Ferry Member was demonstrated to be equivalent to the Wanakah and Jaycox Members of western New York, but not to the Tichenor and Deep Run; further, the approximate contact of the Wanakah and Jaycox Members can be determined at Cayuga Lake. This seems to eliminate the need for the term King Ferry and it will not be used in the subsequent discussion. The presently accepted subdivision of the Ludlowville in western New York (see Rickard, 1975, 1981; Baird, 1979) is shown in Figure 3.

Working in the Skaneateles Quadrangle Smith (1935) independently provided a separate subdivision of the Ludlowville Formation into five members; in upward progression these are the Centerfield Member, Otisco Shale, Ivy Point siltstone (with upper and lower siltstone tongues separated by a middle shale), Spafford Shale and Owasco Siltstone, all below the Portland Point Limestone. The terminology of Smith has not been substantially modified except that Gray (1984) extended the name Chenango Sandstone to the Skaneateles meridian to replace the term Centerfield; Gray noted that this interval is lithologically a coarse siltstone and silty shale, not a limestone and is, thus, more similar to the Chenango Member than to the Centerfield although the two members are closely related and intergradational.

The duality of terminology for the Ludlowville between the Cayuga and Skaneateles areas reflects a persistent problem. Until recently no precise correlation existed between units west of Cayuga Lake and those of the Skaneateles area just 20 miles to the east. Initially there seemed little similarity between the Otisco-Ivy Point-Spafford-Owasco and the Ledyard-Wanakah-Jaycox sequence. Indeed, there is not a one-to-one relationship among them with the exception that the Otisco Member is the

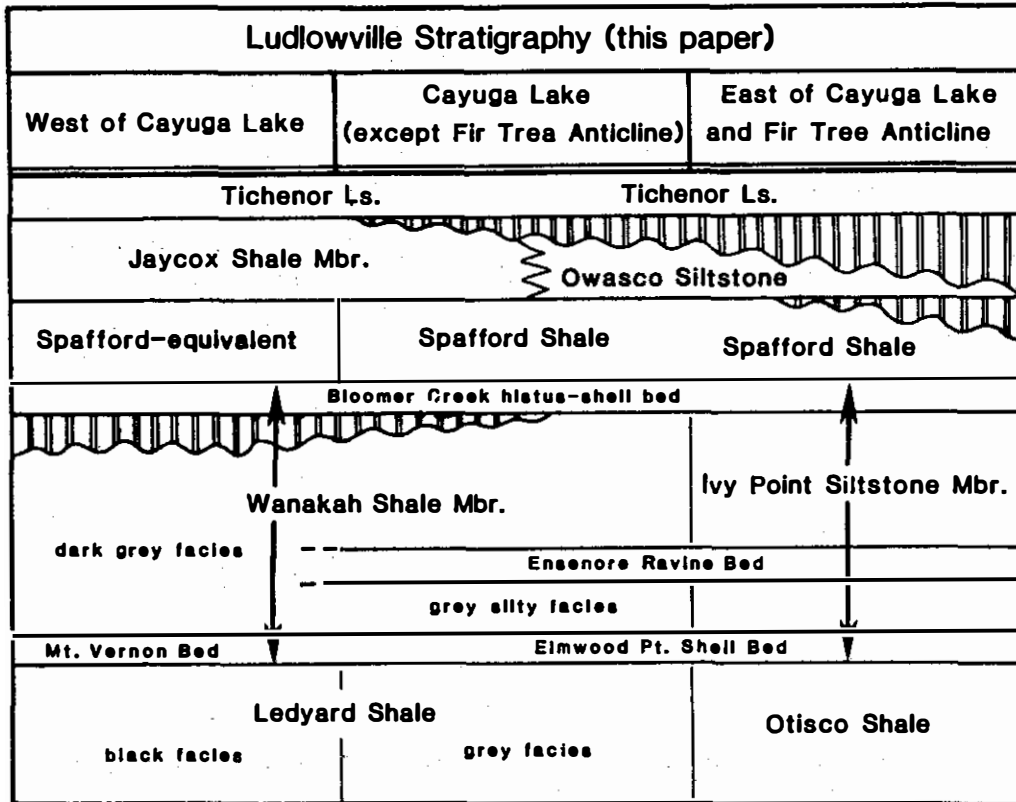


FIGURE 3.--Proposed stratigraphic terminology for Ludlowville Formations in Cayuga Lake and adjacent areas.

sole eastward correlate of the Ledyard Member. One thrust of our recent field studies of the Ludlowville Formation has been to establish detailed correlations between the persistent, easily recognized, and generally, more calcareous western Members and Smith's coarser clastic Members of the central New York region. The Seneca-Owasco Lake region has provided the most challenge in our correlation effort due to the complexity of these facies changes.

Detailed Correlation of the Ludlowville Members

Figures 3 and 4 illustrate our proposed correlations of the Ludlowville units. Our approach has been to locate thin key-beds in this region, both for correlation of surrounding deposits, and also as base lines for assessing lateral facies changes in these same units. These bounding layers are primarily condensed, complex shelly beds, some of which contain reworked fossils, hiatus concretions and phosphatic pebbles. Several of the beds display lateral faunal gradients, indicating that they cross-cut biofacies, despite their consistent thickness and position with respect to other such beds. These observations indicate that the beds formed during

periods of widespread sediment starvation and/or erosion that affected major tracts of the sea floor. In most cases these beds also coincide with the tops or bases of small regressive cycles. The persistence of certain units is remarkable, for example, a thin (20-50 cm-thick) shell bed (Mt. Vernon-Elmwood Point bed) marking the base of the Wanakah Shale and equivalent Ivy Point Member, can be traced from Lake Erie nearly to the Tully Chenango Valley, a distance of over 150 miles; moreover, this bed is underlain everywhere, from Lake Erie to Owasco Lake, by a unique interval containing abundant Phacops trilobites, frequently as molt ensembles. Less persistent beds, apparently recording single sedimentation events (e.g. tempestites, distal turbidites), can be used for precise, isochronous correlation within limited areas; for example, Baird (1981) used the Mack Creek turbidite as a regional control marker in correlating outcrops of the medial (Wanakah-equivalent) part of the Ludlowville Formation around Cayuga Lake. A second correlation technique involves the matching of facies reversal points within small-scale cycles; as noted above, certain of these regressive maxima coincide with facies which are condensed and conspicuous. Using a network of such event beds and cycles we have made more refined (within-member) correlations across the critical Cayuga area. These are summarized as follows.

Centerfield-Chenango

As a result of detailed field studies, Gray (1984) established a direct connection between the Centerfield Member of western New York, and the Chenango Sandstone of the type Hamilton area. The interval is bracketed by a condensed shell-rich bed (Peppermill Gulf bed) below, which rests sharply on the underlying black shale of the upper Levanna or Butternut Members, and above by a phosphatic pebble-rich, shelly layer (Moonshine Falls bed) which locally overlies an erosion surface on the upper Centerfield or Chenango sandstone. A third unit that can be tentatively correlated is the Stone Mill Limestone which overlies the main sandstone bench of the Chenango Member (Earlville submember of Gray, 1984) this encrinite and coral-bearing unit can be traced into a coral bed that overlies the top of Centerfield calcareous mudstone in the Finger Lakes area.

Ledyard-Otisco Members. -- The base of the Ledyard Shale (and equivalent Otisco Member) is sharply demarcated from the Centerfield and Chenango beds by the Moonshine Falls bed; this shell and phosphatic pebble-rich layer can be traced from the Seneca Lake Valley eastward to the Chenango Valley. A series of higher connections can be made locally between the Skaneateles and Owasco Valleys; notable among these is a phosphatic pebble bed, associated with the Staghorn coral biostrome (Oliver, 1951), which persists a short distance across the transition from Otisco to Ledyard facies; we believe that a thin grey shale band in the lower black Ledyard at the type (Paines Creek) section marks this same horizon.

Toward the top of the Ledyard an interval of small carbonate nodules with well preserved cephalopods (Sheldrake beds) is traceable from western New York into the transition with the upper Otisco shales near Owasco Lake. Finally, the top of the Ledyard and Otisco shale is uniformly

marked by a condensed shell-rich silty layer (formerly called the Strophalosia Bed; Cooper (1930), which is now termed the Mt. Vernon-Elmwood Point bed.

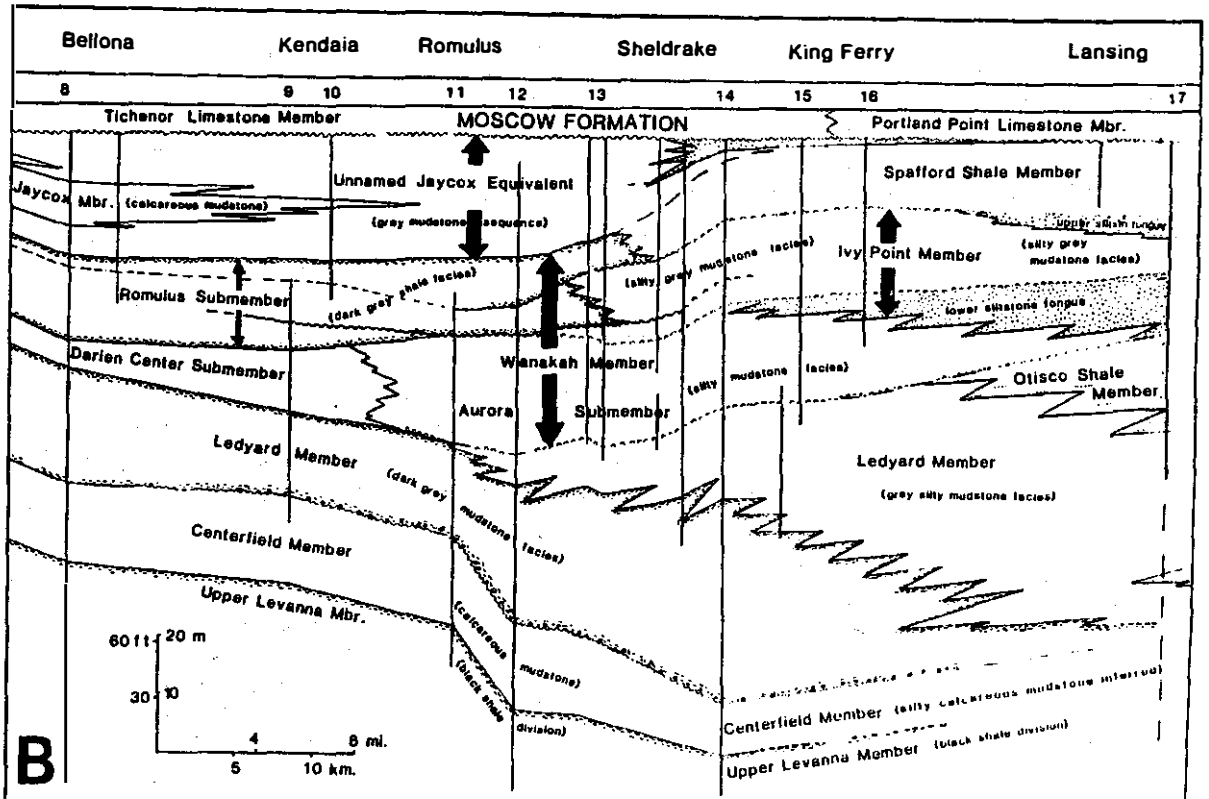
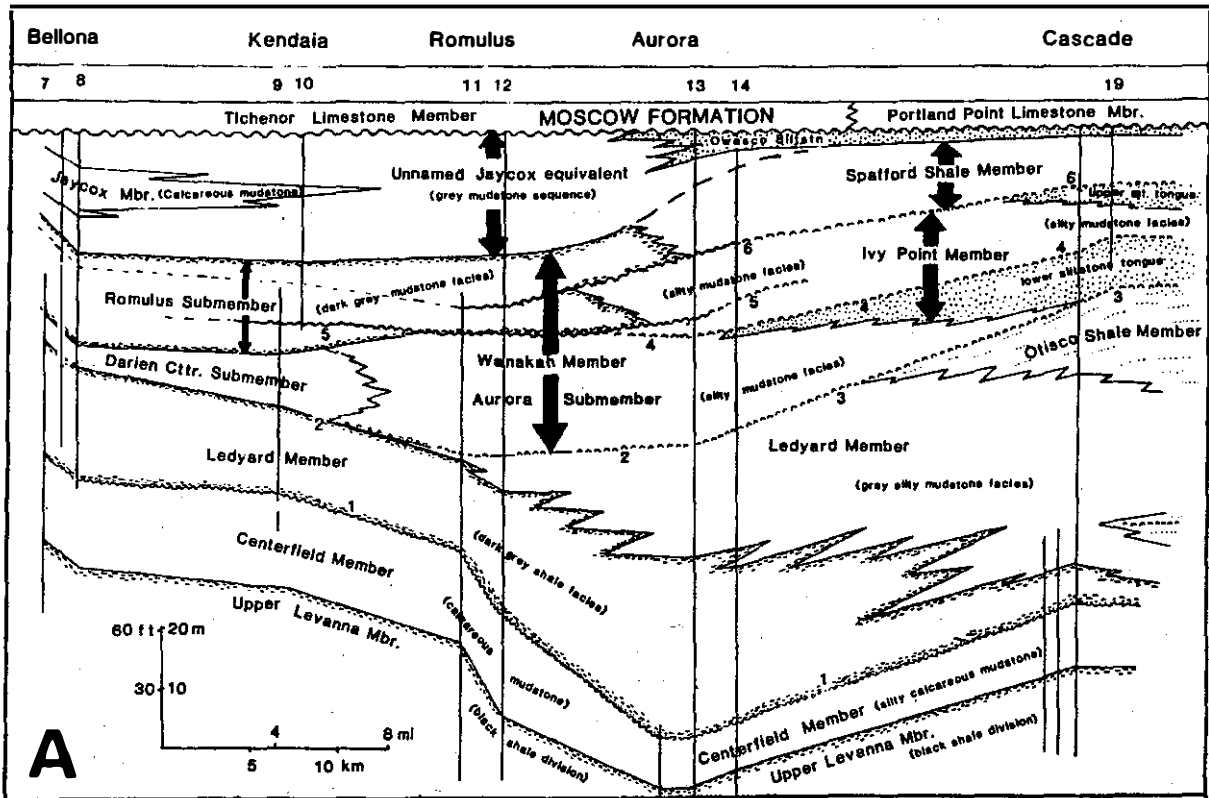
Wanakah Shale Ivy Point.-- The Elmwood Point shell bed establishes an important reference line between the basal Wanakah shale and Ivy Point Members. The Ivy Point Member, defined by Smith (1935), is a tripartite unit with lower and upper siltstone tongues, separated by a middle shale unit. Detailed correlation of the lower Ivy Point siltstone submember with the lower Wanakah beds (Aurora and Darien Center submembers) has been established by Miller (in prep.); he has noted three subcycles within the lower Ivy Point; these upward-coarsening hemicycles correspond respectively to the "Pleurodictyum beds," Darien Coral bed, and Murder Creek ("trilobite") Bed of the lower Wanakah Member. The middle Ivy Point shale contains a minor silty subcycle the top of which appears to correlate with Baird's (1981) Barnum Creek hiatus bed; overall, the middle shales correlate with a middle black shale lentil of the Wanakah in the Seneca Lake meridian.

Precise stratigraphic relationships of the upper siltstone submember of the Ivy Point with the upper Wanakah have been established because of the continuity of the Bloomer Creek hiatus concretion bed (Baird, 1981) with a rich shell-phosphatic bed at the abrupt upper contact of this higher upward-coarsening cycle. Minor subcycles within the upper siltstone submember have not yet been correlated westward.

Spafford Shale - Uppermost Wanakah black lentil.-- The Spafford Shale is rather clearly demarcated at its base by the Bloomer Creek-equivalent shell bed. On the basis of its position the Spafford can be correlated with a thin (1-2 m) dark grey to black shale lentil (Romulus submember) that overlies the Bloomer Creek bed westward from Cayuga Lake. About 10 m of soft grey shales exposed below an erosional contact with the Tichenor limestone at Salmon Creek (Ludlowville type section) and at Portland Point can be designated as Spafford shale as these shales overlie the Bloomer Creek bed.

FIGURE 4.-- A. Refined stratigraphy of Ludlowville Formation in an east-west direction across the central Finger Lakes region; this transect crosses depositional strike such that basinal facies in the western half of the section grade into shallow subtidal and possibly even lower shoreface deposits on the extreme right. Marker beds, indicated by numbers, include: 1) Moonshine Falls bed, 2) Mt. Vernon bed, 3) Elmwood Point bed (eastern equivalent of Mt. Vernon bed), 4) Ensenore Ravine bed, 5) Barnum Creek bed, and 6) Bloomer Creek bed.

B. Stratigraphy of the Ludlowville Formation in a northwest-to-southwest direction across the central Finger Lakes Region. Note the striking similarity of this profile with Figure 4A. Most of this lateral "basin-to-silty platform" facies change is visible in the Cayuga Valley because that lake trends essentially normal to Middle Devonian facies strike.



Owasco Siltstone - Jaycox Member.-- The Owasco Member is a thin bed (0.5 m) of coarse siltstone to fine sandstone which in some areas appears to have a gradational lower boundary with the underlying Spafford shale. Elsewhere the base is sharp and erosional. The top of the Owasco is sharply and erosionaly overlain by basal Portland Point (= Tichenor) crinoidal grainstone. Locally, as in the area of Ludlowville and Portland Point, the Owasco is absent, presumably due to pre-Tichenor erosional truncation. Despite this bracketing, the exact western correlation of the Owasco is uncertain at this time. A calcareous, silty bed with a few corals occurs near the base of the Jaycox Member at King Ferry, New York. This suggests mutual correlation with the Owasco to the east and the basal Jaycox (Hill's Gulch) bed to the west. If so, the main body of Jaycox silty shales must have been erosionaly truncated by pre-Tichenor erosion everywhere east of Cayuga Lake. This situation is reminiscent of that seen in western Erie where the Jaycox is erosionaly truncated and the Tichenor limestone rests on Spafford- equivalent upper Wanakah shales (Baird, 1979).

In summarizing these correlations, it can be noted that nearly all of the conspicuous, fossil-rich, calcareous intervals of western New York can now be matched with upward-coarsing cycles in the east (Figs. 4, 6). Condensed capping shell pebble beds (i.e. Moonshine Falls, Staghorn, Mt. Vernon-Elmwood Point, Darien-Coral, Murder Creek-Ensenore, Barnum, Bloomer, and Hills Gulch) can be tentatively traced across major facies transitions (Fig. 4). These correlations establish a series of about six to seven major and minor transgressive/regressive cycles in the Ludlowville Formation which seem to cross-cut facies and are, therefore, probably allocyclic in nature (see below).

LUDLOWVILLE LITHOFACIES

Several distinctive facies recur in predictable sequences in the Ludlowville Formation of the Cayuga Valley area. These lithologies are described and interpreted in the following sections, arranged approximately in order of appearance of lithofacies within shallowing cycles.

Lithofacies Descriptions

A) Dark-gray to black shales.-- Dark, sometimes rusty-weathering, fissile to platy shales with millimeter-scale lamination, minor disseminated pyrite and rare carbonate concretions; fossils sparse to very common, poorly preserved, of low diversity, and dominated by small pelagic organisms.

These sediments are typical of the central Finger Lakes area; they grade laterally, into facies B and C mudstone. They appear to represent relatively deep, poorly-oxygenated environments of the basin center. Laminae represent slight differences in grain size, with thin, light colored silt and darker clay laminae; preservation of fine laminae is due to absence of bioturbation in minimally dysaerobic to anaerobic sediment (see Byers, 1977; Cluff, 1980; Aigner, 1980, for discussion of similar facies).

B) Medium to dark gray bioturbated mudstone.-- Thinly-bedded and fissile to massive and homogeneous, slightly calcareous to silty mudstones; with interspersed thin shell-rich horizons; and commonly contain scattered small brachiopods, nuculid clams, trilobites and nautiloids. Fine, threadlike to cylindrical pyritic burrow linings abundant; calcareous concretions ranging from 10 to 50 cm in diameter occur at numerous horizons within these deposits.

Detailed examination of Hamilton mudstones indicates that they are composed almost exclusively of pelleted mud, ranging upward to medium silt size (see Wygant, 1986; Brett, et al., 1986). The homogeneous, pelletal nature of the mudstone reflects persistent, though generally shallow, bioturbation that has destroyed most of the primary fabric. The occurrence of thin shelly layers may have resulted from minor storm winnowing. Pyrite aggregates and steinkerns indicate the presence of local, sulfidic microenvironments inside shells and burrows within the reducing, nonsulfidic muds; such microenvironments induced concentration gradients due to bacterial sulfate reduction and anaerobic decay of the contained organic matter (Hudson 1982; Dick and Brett, 1986). Rapid burial of organic matter may also have been a catalyst in producing carbonate precipitation (see Weeks 1957; Berner, 1968; Raiswell, 1971, 1976).

The color of these mudstones ranges from a very dark slate grey to medium grey in the central Finger Lakes region; although local variations in shale color reflect differences in organic content between beds, a regional, east-west color gradient appears to have a late diagenetic origin.

C) Bioturbated silty-calcareous mudstones.-- Massive 0.5 to 5 meter-thick calcareous to silty mudstones; conspicuously structureless except for abundant spreiten of Zoophycos and occasional shell beds or relicts of thin (2-5 cm), laminated siltstone layers; fossils rare and scattered, but typically well preserved.

The mudstone intervals record numerous rapid burial events; these siliciclastic-carbonate mud layers, presumably winnowed from nearby shallow shelf areas during storm events, were deposited in adjacent lower energy regions. However, pervasive and deeply-penetrating bioturbation, mainly by Zoophycos-producers, destroyed most primary lamination, leading to homogeneous, blocky mudstones. This facies forms the lower and middle parts of upward-coarsening cycles in New York State. These intervals tend to maintain prominent vertical joint planes, forming massive wall-like bluffs. Weathered joint facies display a distinctive pattern ("fretwork" surfaces) which reflect differential weathering of Zoophycos spreiten exposed in cross-section.

D) Cross-stratified, coarse-grained siltstones and sandstones.-- Thick-bedded to massive protoquartzite and subgraywacke units, ranging from one to five meters thick including amalgamated layers of hummocky cross-stratified, coarse siltstone to medium-grained sandstone. Surfaces typically display coquinites of disarticulated, convex-upward brachiopod

and bivalve shells, as well as symmetrical and asymmetrical ripples. Sandstones are carbonate-cemented and contain horizons of large (up to 30 cm-across), spherical carbonate concretions. Persistent bands of soft-sediment deformation (ball and pillow structures) occur at certain levels. Bioturbation is minor where HCS bedding predominates but adjacent beds are usually intensely burrowed by Zoophycos.

The sandstones are interpreted as siliciclastic analogs of the condensed winnowed crinoidal calcarenite and they appear to grade laterally into the latter (Gray 1983, 1984). Sandstone deposits consist of amalgamated, storm reworked accumulations (cf. Aigner, 1985), usually developed within otherwise winnowed shelf sand. They form the upper beds ("roof beds" sensu Bayer et al. 1985) of coarsening-upward hemicycles in the southern Cayuga Lake area.

E) Shell-coral beds.-- Indurated, shell-rich (1-10 cm-thick) wackestone layers alternating with sparsely fossiliferous mudstones (Facies B,C). Thicker units typically display basal lag concentrations of finely-comminuted and corroded shell debris with well-preserved, articulated and unbroken skeletons concentrated along the upper surfaces. Shells on bed tops are preferentially oriented convex-upward and may display mud-sheltering. Certain shell beds pinch and swell along outcrops at wavelengths in the range of one to six meters. Intervening mudstone beds are sparsely fossiliferous, but may contain extraordinarily well-preserved fossils imbedded in, or attached to, the upper surfaces of underlying shell beds (see Brett, et al. 1983).

This evidence suggests that these shell beds are multi-event layers indicating relatively long-term accumulations of skeletal debris during times of minimal sedimentation. Skeletal accumulations were subjected to minor local currents and bioturbation during exposure.

Minor scouring at the bases of certain shell beds suggests brief intervals of erosion associated with the local shifting of skeletal debris across the seafloor during major storms. Shell beds were smothered during episodes of rapid mud accumulation (distal mud tempestites?) which buried groups of living organisms inhabiting the upper surfaces of the shell layers, usually as a single event.

F) Condensed beds; nodular phosphates and hiatus concretions.-- Thin (5-50 cm-thick), extremely widespread beds (traceable in outcrops for 150 to 200 kilometers); typically shelly, mud-supported beds containing a high proportion of disarticulated, fragmented and/or corroded skeletal material (rugose corals, thick-shelled brachiopods, crinoids, bryozoans and others), in addition to well-preserved fossils (Baird, 1978; 1981; Baird and Brett, 1983). Fossil diversity is high, but may reflect additive incorporation of taxa from several distinct communities. Thick-shelled brachiopods and corals commonly display loss of surficial detail and abundant micro- and macroborings, as well as encrustation by one or more generations of epizoans indicative of long-term exposure on the sea bed (Baird and Brett, 1981; 1983). Condensed beds may also contain small (usually less than 1 cm-diameter), rounded, black phosphatic nodules, some

of which represent reworked, prefossilized steinkerns (Baird, 1978), and bored, encrusted and corroded hiatus concretions also occur at several localities (Baird, 1981).

Discontinuities and condensed beds are particularly associated with the tops of upward-coarsening, regressive hemicycles which occur east of the Finger Lakes Trough; condensation and erosion appear to be primarily associated with transgression events. Where condensed beds are sharply overlain by laminated black sediments, they may contain lags of reworked, tubular or steinkern pyrite (Baird and Brett, in press).

These beds clearly reflect widespread episodes of minimal net sedimentation and episodic bottom scour, for periods probably lasting 100's to 1,000's of years. In some cases significant erosion is evident, particularly where paleoslopes are indicated by other lines of evidence (Baird, 1981; Brett and Baird, 1982). The mixture of well-preserved and corroded fossils indicates complex histories for these layers, involving multiple episodes of rapid burial followed by exhumation and reworking. Bioturbation, predominantly by *Zoophycos*, also may have been significant in comingling sediment, fossils, and clasts of different ages, producing stratomictic discontinuities sensu Baird (1978; 1981).

LUDLOWVILLE BIOFACIES

Ludlowville beds have long been famous for their well preserved and diverse marine invertebrate fossils which form rather clearcut recurring associations (Cleland 1903; Cooper 1930, 1933; Smith, 1935). In the past two decades well over 70 fossil "communities" (also termed associations, or biosomes by some authors) have been recognized from the Hamilton Group, either qualitatively or as the result of multivariate analyses (Grasso, 1970, 1973, 1978, 1986; Miller 1986; Brower et al. 1978; Savarese et al. 1986; Gray 1984); however, many of these named communities are approximately synonymous and, until recently there has been little attempt to integrate the various studies. Synthesis of these studies with recent facies mapping of the Hamilton Group has permitted the development of a general biofacies model for these rocks (Brett et al. 1983, Miller, 1986, Gray 1983, 1984). This model incorporates and integrates the previous studies, as well as abundant unpublished data, to delineate generalized biofacies representing groups of closely related fossil associations (or paleocommunities). The model relates biofacies to one another and to inferred paleoenvironmental parameters.

Figure 5, modified from Brett et al. (1983) and Brett and Baird (ms. in preparation) depicts relationships among several recurring Hamilton biofacies. Adjacent biofacies on this diagram are those which most commonly border and intergrade with one another. The relative ordering of biofacies has been established on the basis of consistent vertical and lateral intergradation of fossil associations in five or six relatively complete sedimentary cycles of the Hamilton Group. In a general sense, the left hand side of the chart represents typical ordering of fossil associations in dark shale-carbonate, regressive cycles of the Hamilton Group in western New York, while the right side reflects the typical

vertical sequence of associations in the correlative, upward-coarsening cycles in the eastern Finger Lakes region (see Figure 6). The lateral gradation of various portions of these two types of faunal cycles has been demonstrated in numerous sequences bounded by apparently isochronous beds.

Hence, the vertical axis of the diagram records changes in response to overall regressive (shallowing) episodes which affected both western and central New York. Changes in biofacies along this axis are thought to be the direct or indirect consequence of changing water depth. Factors actually responsible for community replacement may include oxygen levels (which should decrease in the deeper waters of a stratified basin), turbulence, light penetration, temperature, food supply, etc. The deeper-shallower polarity along this axis has been established by several lines of evidence including: 1) facies geometry (e.g., black, *Leiorhynchus*-bearing shales tend to occur nearest the presumed center of the Appalachian Basin and to grade concentrically outward into gray *Ambocoelia*-rich shale, then into *Athyris* or *Mucrospirifer*-dominated gray mudstones, etc.); 2) position within upward-coarsening hemicycles, in which sedimentary structures (e.g. hummocky cross stratification) unequivocally indicate decrease in water depth upward; and; 3) lateral

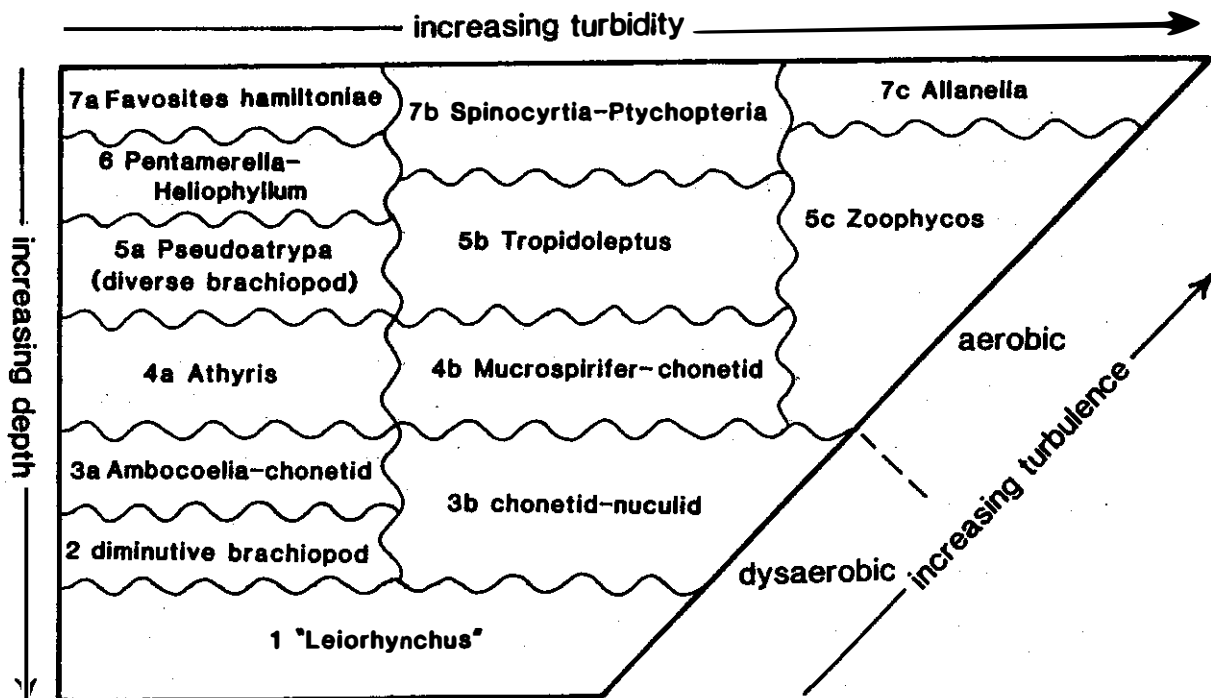


FIGURE 5.--Paleocological model relating Ludlowville biofacies to inferred gradients of depth-related parameters and turbidity and/or sedimentation rates.

gradation within single isochronous condensed beds or intervals, which display upslope biofacies change in the relative order shown on the Figure.

Absolute depths can only be approximated at present. Evidence here is twofold: 1) evidence for position relative to the photic zone (e.g. indications of benthic algae or of visual systems in benthic organisms) and, 2) indications of position of biofacies relative to normal and storm wavebase (see Liebau 1980, for approximations of these depths) utilizing a proximity spectrum for storm generated beds (see Brett et al. 1986).

In general, most Hamilton facies, except perhaps the Leiorhynchus black shales, are thought to have been deposited within the photic zone (probably ~50 m for a muddy epeiric sea), because of the abundance of phacopid trilobites with well developed eyes in most facies and circumstantial evidence for existence of benthic algae (herbivorous gastropods and evidence of possible algal substrates (see Brett et al. 1983). The large, complex eyes of Greenops and Phacops were quite possibly adapted for utilization of very low light levels so that these trilobites, (which are abundant, at least into the Ambocoelia-biofacies, though often lacking in Leiorhynchus associations), may have thrived in dysphotic settings. Finally, study of microendoliths produced by algae indicates that most Hamilton facies were deposited within photic depths; however, there is a marked increase in the abundance and diversity of microborings that correlates with overall faunal diversity increase (Vogel, et al, ms. submitted). This distribution pattern corroborates other evidence for the relative depths of the Hamilton biofacies.

A more refined subdivision of relative depths may be possible using evidence of tempestites (Brett et al. 1986). Amalgamated proximal storm beds are typical of the Favosites hamiltoniae and Allanella biofacies. Similar beds are forming today in shallow shelf settings generally at 10-20 m depths (Reineck and Aigner, 1982). Crudely graded proximal storm layers (coquinities, calcisiltites) are commonly associated with Tropidoleptus and diverse brachiopod biofacies; the lowest evidence for storm wave scouring (minor gutter casts) occurs in the Ambocoelia biofacies which may, therefore, have been deposited near the lower end of storm wave base (about 50-75 m; see Liebau, 1980), while only the most distal storm-generated currents apparently affected Leiorhynchus biofacies. In summary, then, the entire spectrum of Hamilton biofacies probably was developed in relatively shallow epeiric sea environments ranging from perhaps about 10 m downward to about 100 m. However, changes within this range appear to have exerted a strong control on fossil distribution.

The sequence of depth related biofacies is not identical in western and central New York, nor in every successive cycle in a given area. Certain biofacies (especially Leiorhynchus) persist with little change at an analogous position within any cycle from east to west. These are most probably the deepest water facies (see below) nearest the basin center, and, as might be predicted, they show the least effect of local conditions. However, other biofacies, particularly those toward the top

of the diagram, appear to substitute for one another laterally at a given level within the cyclothem. Thus, for example, assemblages of diverse brachiopods and corals (biofacies 5A), in western New York, typically grade eastward into Tropidoleptus-dominated associations (biofacies 5B). The same type of substitution has also been observed vertically between lower and upper halves of a given cycle at a particular location.

Thus, the horizontal axis of the diagram (Fig. 5) records differences in communities, at a given position within the regressive cycle (and thus within a particular range of water depths), which are due to factors other than depth-related parameters; these may include substrate, turbidity, sedimentation rate, biotic factors, etc. However, the rather consistent association of biofacies changes with the eastward thickening and slight coarsening of cycle members strongly suggests that this change is related to sedimentation-controlled features. We believe that turbidity may have been somewhat more critical than absolute rate of sedimentation or substrate. The chemical composition (i.e., whether carbonate or siliciclastic) and grain size of the substrate appear to have exerted relatively little control, at least on the distribution of brachiopods, as we have seen nearly identical associations, at analogous portions of cycles (i.e. similar depths), developed in sandstones, mudstones or limestones. Similarly, although some biofacies changes are, in general, correlated with increasing thickness of a given sedimentary package, the same type of change can sometimes occur without substantial change in thickness of a bed; this tends to rule out net sediment accumulation rate as the critical factor. However, the biofacies substitutions almost always coincide with circumstantial evidence for increased turbidity, such as increased abundance of trace fossils (Zoophycos) and infaunal bivalves, decrease in presumed turbidity intolerant organisms (i.e. corals), and decrease in abundance of algal endoliths (Vogel et al. submitted).

Biofacies Descriptions

In the following sections we briefly characterize the various Hamilton biofacies from offshore (deepest) to onshore (shallowest).

1. Leiorhynchus.-- Low diversity (10-15 species), or even monotypic assemblages of poorly preserved leiorhynchid, chonetid and rare ambocoeliid brachiopods, nuculid bivalves, nautiloids, goniatites and Styliolina. Invariably associated with black to dark gray, laminated shales (Lithofacies A). Probably the deepest water assemblage (50-100 m); quiet dysaerobic water settings, sediments anoxic near to the sediment water interface, turbidity low to moderate; below storm wave base.

2. Diminutive brachiopod.-- Distinctive, low-diversity assemblages of diminutive brachiopods, particularly small (juvenile?) Tropidoleptus, Truncalasia, and Ambocoelia nana; small bivalves, including Cardiola and nuculids; archeogastropods, such as Palaeozygopleura and Retispira, and Phacops trilobites, associated with dark gray to black, fissile but non-laminated shales, with minor pyritic burrow fillings (Lithofacies A, B).

Slightly shallower and/or more aerobic than the Leiorhynchus association with dysaerobic to lower aerobic bottom waters and anoxic, shallowly-burrowed muds; low turbidity settings below storm wavebase; the occurrence of bedding planes covered with monotypic assemblages of small brachiopods may indicate mass mortality of spatfalls.

3A. Ambocoelia-chonetid.-- Low to moderate diversity (20-40 species) fossil assemblages, including abundant small, free-lying brachiopods (e.g. Ambocoelia, Devonochonetes scitulus); nuculid and modiomorphoid bivalves, archeogastropods, phacopid trilobites; rare small inadunate crinoids and blastoids, aulopodid corals. Dark to medium grey, concretionary, pyritic shales and mudstones (Lithofacies B).

Relatively deep (~50 m), quiet water, moderately oxygenated with anoxic or dysoxic, nonsulfidic, shallowly-burrowed substrates; near storm wave base; at least dysphotoc zone.

3B. Chonetid-nuculid.-- Low to moderate diversity assemblages dominated by Devonochonetes scitulus or Longispina, nuculid bivalves, archeogastropods, nautiloids and Greenops trilobites; rare small Zoophycos. Typical occurrence in dark to medium grey, bioturbated mudstones (Lithofacies B, C).

Depths, analogous to those for the Ambocoelia-chonetid association, but substrate probably more thoroughly bioturbated; turbid water conditions near sediment interface.

4A. Athyris. -- Moderate diversity assemblages (30-40) including stereolasmatid corals, fenestellid bryozoans, the brachiopods Athyris, Mucrospirifer, Devonochonetes, medium sized protobranch modiomorphoid and pterioid endobryssate bivalves, archeogastropods, phacopid trilobites and a few flexible and inadunate crinoids. Common in medium grey claystones to silty mudstones (Lithofacies B), sometimes concretionary or with nodular pyrite; commonly with shell-rich beds (Lithofacies F,G).

Moderate depths, within lower storm wave base (30-50 m), but normally rather quiet with low to moderate turbidity; bottom waters fully aerobic but upper sediments dysoxic.

4B. Mucrospirifer-chonetid.-- Moderate diversity assemblages (30-35 species) often largely dominated by Mucrospirifer and Devonochonetes brachiopods and/or bivalves such as Cypricardella, Paleoneilo; also minor trilobites, typical of medium grey, commonly silty mudstones and muddy siltstones (Lithofacies B,C,D), concretions and pyrite nodules uncommon.

Depths similar to the Athyris biofacies, with which it interfingers; fairly high sedimentation and relatively high amounts of silt; abundant shallow burrowing; probably high turbidity and instable substrate.

5A. Pseudoatrypa (diverse brachiopod)-- Moderate to high diversity fossil assemblages (40-60+ species); contains the highest brachiopod diversity of any Hamilton unit, including various atrypids, spiriferids (Mediospirifer,

Cyrtina, Elita, Nucleospira, Spinocyrtia), strophomenids (Strophodonta, Pholidostrophia, Douvillina, Megastrophia), terebratulids (Centronella, and others); small- to medium-sized rugose corals (Stereolasma, Amplexiphyllum), fenestellid, and fistuliporoid bryozoans, various epi- and endobysate bivalves, trilobites, crinoids, blastoids. Primarily in medium to light gray, soft, commonly calcareous (marly) mudstone (Lithofacies B) with abundant thin shell-rich layers (Lithofacies E); minor Zoophycos bioturbation.

Shallower (20-30 m), fully aerobic muddy- to shelly-bottomed settings below wave base but well within storm wave base; upper sediments aerobic; moderate to deep burrowing; turbidity and sedimentation rates generally low.

5B. Tropidoleptus.-- Moderate diversity (30-50 species) assemblages, dominated by brachiopods (Tropidoleptus, Devonochonetes, Longispina, Mucrospirifer, Athyris, Meristella, Spinocyrtia). Pseudoatrypa and strophomenids may be present but usually rare, bivalves diverse and common, including Cypricardella, Modiomorpha, Ptychopteria, and Glyptodesma; rugose corals rare or absent but the tabulate Pleurodictyum may be common and large; varied ramose and fistuliporoid bryozoans; phacopid and proetid trilobites, often large; platyceratid gastropods; high diversity of crinoids and blastoids. Typifies medium to light bluish gray, soft, bioturbated, blocky and moderately to sparsely fossiliferous mudstones (Lithofacies B) or muddy siltstones (Lithofacies C); scattered coquinites, some with grading, gutter casts.

Analogous depths to Pseudoatrypa association, but with distinctly higher sedimentation rates and/or turbidity. Aerobic muddy to silty bottom areas, commonly affected by epeiric storm waves.

5C. Zoophycos. -- Zoophycos burrowing pervasive; very sparse, moderate diversity (20-30 species) but often large and well preserved body fossils, include various large clams (Actinopteria, Modiomorpha) and brachiopods such as Spinocyrtia. Associated with massive, bluff-forming, light gray, buff-weathering, bioturbated, silty mudstone or muddy siltstone (Lithofacies C).

Shallow, but low to moderate energy high turbidity settings, below normal wave base, probably affected by storm waves but most traces of bedding obliterated by Zoophycos; turbid and unstable substrates inhibited colonization by epifauna (trophic group amensalism) but those organisms which did colonize often grew to large size (abundant suspended food).

6A. Pentamerella (Heliophyllum).-- diverse (50 to 60+) coral-dominated fauna with many species of large turbinate and fasciculate rugosans (Heliophyllum, Cystiphyllodes, Eridophyllum, and ramose to massive tabulates (Favosites, Alveolites, Cladopora); fenestrate, fistuliporoid and ramose trepostome bryozoans; brachiopods less common but represented by several genera, some of which (e.g. Pentamerella, Parazyga, Elita, Pentagonia) are largely confined to this facies; crinoids are abundant, but usually highly disarticulated; platyceratid gastropods and large

pteroid clams locally common. May be biostromal; fossils commonly fragmented and heavily corroded.

Shallow, moderate to high energy settings approaching normal wave base (~20 m), well aerated, with abundant food supply; skeletal buildup on seafloor possibly inhibiting burrowing but favoring diverse epibiont communities; sedimentation rates very low. Light gray, commonly soft, crumbly, calcareous (marly) mudstone (Lithofacies B) to nodular argillaceous limestone, or, rarely, calcareous siltstone (Lithofacies C).

6B. Spinocyrtia (Ptychopteria).-- Low to moderate diversity (20-30 species) fauna concentrated in winnowed coquinite layers; heavily dominated by large epibyssate bivalves (Ptychopteria, Actinopteria, Glyptodesma), but with common large and/or robust brachiopods such as Spinocyrtia, Camarotoechia; Protoleptostrophia, Devonochonetes, Tropidoleptus, minor ramose bryozoans and crinoids; Dipleura trilobites; fossils usually preserved as uncompressed molds. Typical of blue gray, buff-weathering massive to laminated or cross laminated, coarse siltstones and fine grained sandstones, (Lithofacies C,D), with or without abundant Zoophycos burrowing.

Shallow (<20 m), high energy, silty to sandy bottomed environments; strongly affected by storm winnowing and approaching normal wave base; deep burrowing may be present or may be inhibited by movement of sand traction sheets.

7A. Favosites hamiltoniae-- A moderately diverse assemblage characterized by large hemispherical favositids, and rugose corals such as Eridophyllum, Heliophyllum, and Cystiphyllodes, trepostome and fistuliporoid bryozoans, robust brachiopods (e.g. Megastrophia, Spinocyrtia, Rhipidomella, large bivalves, and abundant pelmatozoan ossicles; typically in crinoidal pack- and grainstone or coarse siltstone to fine, well-sorted sandstone (Facies D).

Very shallow, moderate to high energy, nonturbid settings near normal wavebase, with well-winnowed, firm substrates.

7B. Allanella.-- Low diversity assemblages, including the brachiopods Allanella tullius, Camarotoechia, Tropidoleptus, Schuchertella and large chonetids, large bivalves such as Ptychopteria and Glyptodesma may also be present; fossils commonly disarticulated, fragmented and/or moldic; associated with cross-bedded, commonly hummocky, rippled, coarse siltstone and sandstone, with intermittent coquinite beds;

Very shallow (10-15 m) high energy settings within, or close to, normal wave base; burrowing typically prevented by physical disturbance of silts and sands (Lithofacies D).

LUDLOWVILLE FACIES CYCLES

The Devonian System in the Appalachian basin, has already been examined for evidence of allocyclic and autocyclic depositional events; Dennison

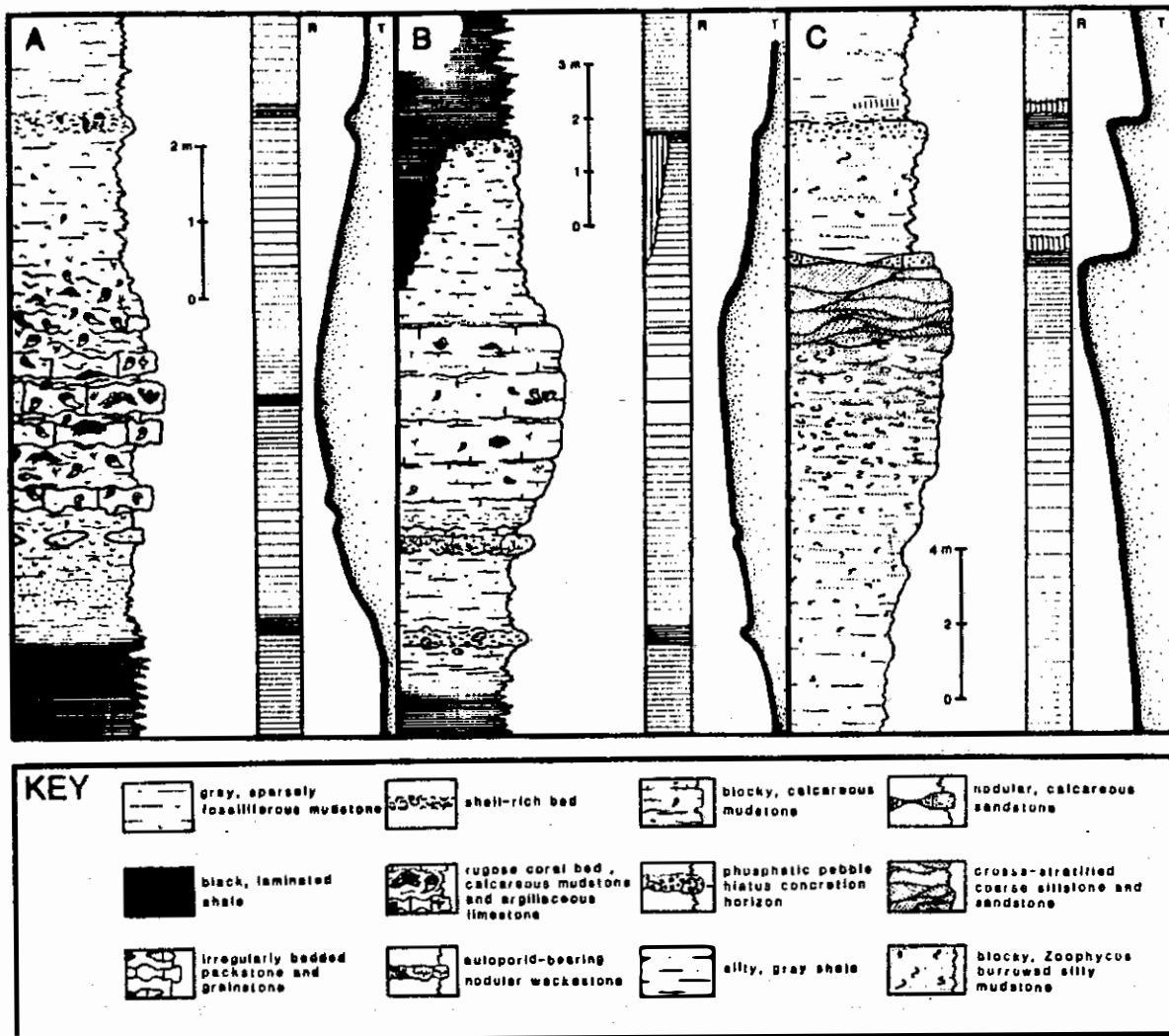


FIGURE 6.--Sections of a representative Ludlowville cycle, the Centerfield Limestone and equivalent Chenango Sandstone from A) Genesee Valley, B) Cayuga Valley, and C) Chenango Valley. Interpreted relative sedimentation rates and regression/transgression curves are shown for each section; within columns, close spacing of horizontal lines indicates slow net sedimentation; vertical ruling indicates hiatuses. Note change from subsymmetrical limestone centered cycle in the west to an asymmetrical, coarsening-upward cycle in the east.

and Head (1975), in summarizing intrabasinal correlations of the Devonian section, identified numerous eustatic fluctuations within the sequence. More recently, House (1983) and Johnson, et al. (1985) have presented intercontinental correlations of the Devonian system including deposits within the Appalachian Basin; these authors recognize the widely correlatable Devonian black shale deposits as marking eustatic transgression maxima, and House (1983) recognizes a three-fold hierarchy of cycle-magnitude (based on vertical thickness) in the Upper Devonian.

We recognize two scales of rather regular facies alternation within the Hamilton Group of the Cayuga Lake region. The two different orders of facies oscillations are differentiated on two bases: a) thickness, and b) magnitudes of lithofacies and, particularly, biofacies variation. Smaller (6th order) cycles, are relatively thin packages (generally <1 m in western New York and <3 m in thicker central New York sequences); moreover they display limited facies variation with a range of faunas reflecting only a portion of the total Hamilton biofacies spectrum; they may or may not be basinwide in extent. Conversely, larger-scale cycles are thicker (3 to 20 m) and display more nearly complete facies spectra (e.g., from black, fissile shales with Leiorhynchus to limestones with diverse faunas); these are invariably widespread.

We argue that these sequences are cycles in that they display a fairly regular, repeated and predictable sequence of litho- and biofacies, i.e. the sequence is not random (see Lukasik, 1984; Zell, 1986; Linsley, 1986; Savarese et al, 1986; Miller, 1986, for detailed study of individual cycles).

Whether or not cycles have similar durations is more difficult to ascertain because of a total lack of absolute dates. At present we can only note that the two levels of cyclicity each show rather similar thicknesses of strata per cycle in a given geographic area. If we equate rock thickness with time (a questionable assumption) then we would conclude that cycles have approximately similar periodicities. There does appear to be some discontinuity between small, 0.5-1.0 m-scale, cycles and larger 3-5 m, cycles. It is difficult or impossible to correlate some smaller cycles across very large areas, whereas all of the larger cycles can be traced regionally.

Finally, there is now strong evidence that the larger cycles are allocyclic in nature; not only can they be traced across facies boundaries within the Hamilton Group of New York (Brett and Baird, 1985), but several also can be tentatively correlated with transgressive-regressive cycles in the rest of the Appalachian Basin, the Cordilleran region and also in Europe, Morocco, and elsewhere, using conodont biostratigraphy for rough dating (see Johnson et al. 1985; House, 1983). This suggests a eustatic signature in these events.

The causes of smaller-scale oscillations are less certain. Some appear to be widespread, at least in western and central New York State; these also cross-cut facies and may be the result of allocyclic processes, probably either sea level or climatic fluctuations. Other small-scale,

coarsening upward cycles appear to be restricted to central New York areas, and, therefore autocyclic mechanisms can not be ruled out. These might record episodic subsidence and tectonic adjustment of the southeastern basin margin or possibly local deltaic progradation (Selleck, 1983).

Large and small-scale, upward-coarsening cycles occur within the Ludlowville Formation in the Cayuga-Owasco area (Figs. 6, 8-10). They comprise distinctly asymmetrical, coarsening-upward hemicycles, ranging from 0.5 to 3 meters-thick. Each cycle begins with a thin (1-10 cm) shell-rich bed (Lithofacies F,G) containing diverse fossil assemblages, commonly dominated by the Athyris brachiopod biofacies (Athyris, Mucrospirifer, Spinocyrtia, various chonetids, the small rugose coral Stereolasma, and crinoidal debris). Most shells are disarticulated and some are fragmented and corroded. Small phosphatic pebbles (0.5 - 1.0 cm in diameter), including reworked phosphatic steinkerns, also occur in association with skeletal debris in some cases. Such roof beds usually abruptly overlie, but may be amalgamated with, the uppermost siltstones of the previous hemicycle.

The lower part of each hemicycle above the transgressive debris layer includes thin intervals of clayshale (Lithofacies B) with diverse brachiopod and bivalve faunas, including many of the elements of the underlying shell bed assemblage. Minor thin shelly layers alternate with Zoophycos-bioturbated claystone layers. Higher portions of the cycle consist of silty, thoroughly Zoophycos-bioturbated, massive mudrock with an (Lithofacies C,D) assemblage dominated by Tropidoleptus, Mucrospirifer (Biofacies III, V) and numerous bivalves. A zone of rusty-weathering, calcareous concretions, formed around the pyritized vertical shafts of Zoophycos, occurs toward the top of certain cycles. Thicker cycles may terminate in beds of coarse, laminated or bioturbated siltstone or fine grained sandstone (Lithofacies E), which contain layers of larger skeletons, particularly robust Spinocyrtia, and scattered, large rugose or tabulate corals. These coarse upper beds are more resistant to weathering than the underlying mudstone and tend to form waterfalls in stream exposures. Cycles commonly terminate with a flat, bench-like upper surface on which lies reworked shell and nodular phosphatic debris as well as succeeding softer mudstone beds.

Each cycle appears to record three major phases: First, an initial sediment-starved interval was associated with minor deepening, during which the basal shelly and phosphatic bed accumulated; this was followed secondly by a relatively rapid aggradation of the sediment surface to slightly shallower levels. Upward faunal changes record both the increased sedimentation rate and gradual shallowing. The third and last phase is recorded in the upper siltstone bed; this involved gradual decreases in net sedimentation rates coupled with multiple physical and biogenic reworking of sediment and bypass of fine-grained material (see Figure 6). Winnowing produced a relatively clean, condensed, coarse silt to medium-grained, sandy substrate and even permitted colonization by relatively turbidity-sensitive organisms such as tabulate corals.

CORRELATION OF LUDLOWVILLE CYCLES IN THE CAYUGA LAKE REGION:

FACIES PRECESSION

The present field trip examines not only the vertical relationship of facies within cycles but also the lateral variation of cycles across a distal-proximality spectrum. Walther's law predicts that the vertically stacked facies within a given cycle should grade laterally into one another across depositional strike. Hence, each analogous part of a given upward-shallowing cycle should display predictable lateral variation in a shoreward direction that is more or less in phase with, but consistently offset from that of the next overlying unit; (e.g. basal black shales of a given cycle should grade laterally upslope into gray mudstones, gray shales should grade to siltstones; siltstones to fine sandstones). We refer to this predictable facies change as facies precession.

In asymmetric cycles, the situation may be complicated in that facies present in the lower, shallowing, hemicycle are either absent or their homologs are quite different, in the upper, deepening, half cycle. In some cases, the upper half cycles typically condensed, undergo substantially less change than do the lower counterparts, especially in terms of biofacies. This may reflect variations in sedimentation rates at any given depth during the course of the cycle. During the regressive phase relative sea level drop apparently caused progradation of clastic sediments from southeastern source areas leading to marked differences in depth homologous members of the cycle from west to east approaching the area of maximum sediment deposition. In contrast, during transgressive phases the deepest basinal areas were more uniformly clastic sediment starved. Hence, differences from east to west were less accentuated and a more nearly uniform, thin package of sediments, with rather consistent biofacies, was deposited.

The Cayuga Lake transect provides excellent examples of facies precession within correlative cycles; the best examples come from the middle Ludlowville units, which undergo sufficient change to be designated as different members from northern to southern outcrops in the Cayuga Valley (See Figure 7 for geographic locations described in the following section).

In the King Ferry area (Stop 1) the Ledyard Member is dark gray to black fissile shale (Facies A), dominated by a diminutive brachiopod or chonetid-nuculid biofacies, with minor layers of chonetids and small Tropidoleptus and Ambocoelia. The Wanakah Member consists mainly of rather uniform silty mudstone and darker gray shales with thin horizons of concretions and condensed shell beds, several of which can be traced westward to the other side of Cayuga Valley and beyond.; there is a hint of coarsening-upward cycles, especially in the lower submember (Pleurodictyum beds of Grabau, 1899), and in the upper, Bloomer Creek interval. Most of the Wanakah Shale bears an Athyris or Mucrospirifer biofacies (4A,B), but the silty mudstones of the lower submember display Tropidoleptus or simply Zoophycos biofacies; the tabulate coral Pleurodictyum is abundant in this interval here, as it is west to Lake

Erie. The Spafford Shale overlying the Bloomer Creek shell bed, is sparsely fossiliferous, dark gray mudstone which passes upward into silty mudstone and minor siltstone in the Jaycox Member; the sequence is erosionally truncated by the Tichenor crinoidal limestone. The lower Spafford here yields a recurrent diminutive fauna with small Tropidoleptus, chonetids and nuculid bivalves, closely resembling that of the upper Ledyard shales, and the Jaycox siltstones are characterized by Zoophycos or Tropidoleptus biofacies.

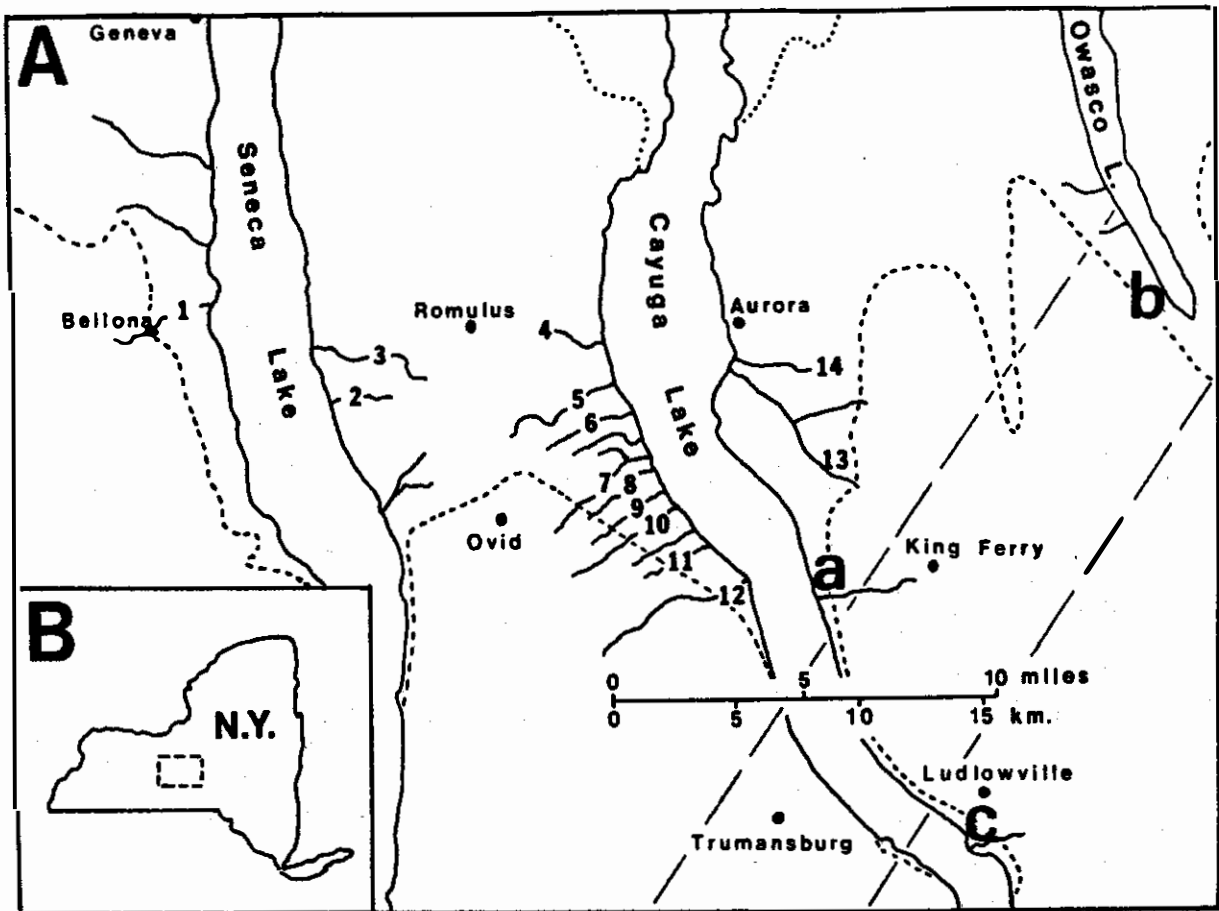


FIGURE 7.--Study area. A. Map of Seneca-Owasco Valley region. Figure shows the outcrop belt of the Hamilton Group (between dotted and dashed lines), including Fir Tree anticline exposures near Ludlowville on Cayuga Lake; large dashed diagonal lines show inferred depositional strike. Key Ludlowville outcrops are numbered, and include: 1) Kashong Glen; 2) Indian Creek; 3) Kendaia Creek; 4) Hicks Gully; 5) Big Hollow Creek; 6) unnamed creek; 7) Mack Creek; 8) Bloomer Creek; 9) Barnum Creek; 10) Powell Creek; 11) unnamed creek, 12) Sheldrake Creek; 13) Paines Creek; 14) Little Creek. Fieldtrip stops include: a) King Ferry Station; b) Cascade, Route 38 roadcut, and c) Portland Point. B. Inset shows position of study area in New York State. Modified from Baird (1981).

At the Route 38 roadcut in Cascade, western Owasco Valley (Stop 2) most Ludlowville units have undergone substantial change, and display a more proximal aspect. The upper Ledyard interval, has changed from a black, fissile shale to a fossiliferous, silty, gray mudstone that can be termed Otisco Member; Ambocoelia and chonetid biofacies have been largely, but not entirely, replaced by Athyris biofacies. Within the Wanakah interval coarsening upward cycles at this locality are well developed, and are distinctly capped by platforms of relatively coarse siltstone. The fossil assemblages of these silty beds are more heavily dominated by Tropidoleptus, Zoophycos, and Spinocyrtia biofacies than those in corresponding beds at King Ferry.

Upper capping shell beds at the tops of cycles display relatively little change but are richer in larger brachiopods (e.g. Spinocyrtia). The Spafford Member has progressed to a relatively fossiliferous, grey, silty shale, with an Athyris to Tropidoleptus biofacies. The basal remnant of the Jaycox-equivalent interval is represented by a coarse, Zoophycos-burrowed to cross-laminated siltstone, of the Owasco Member, which contains coquinites of the low diversity Allanella biofacies.

Finally, at Portland Point (Stop 3) Ludlowville facies are still more proximal in aspect. The Ledyard-equivalent Otisco Member is a grey, silty mudstone with diverse brachiopod faunas resembling those seen in the lower Wanakah at King Ferry. The latter unit, in turn, has graded into burrowed to cross-laminated siltstone of the Ivy Point Member. The capping beds of coarsening upward cycles, which are composed of silty mudstones at King Ferry, are represented at Portland Point by hummocky cross-stratified to massive beds of coarse siltstone or fine sandstone, bearing Allanella coquinites. The lower Spafford Member bears a diverse Tropidoleptus biofacies, while upper beds contain Zoophycos-churned siltstone; the Owasco, if present at all, has been removed by pre-Tichenor erosion. Thus, all evidence points to considerably shallower-water conditions here than at Cascade, which, in turn, represents shallower conditions than at King Ferry. Nonetheless, all three areas were affected by the same shallowing-deepening cycles.

These observations add to the growing evidence for a gently northwestward-dipping ramp or paleoslope existing in the Cayuga Valley during deposition of the Ludlowville Formation (Figs. 1, 7; Baird, 1981; Baird and Brett, 1981). Some qualitative sense of the gradient on this ramp can be obtained by examination of the spectral facies changes within presumed time equivalent packages bounded by distinctive event beds.

One such interval is the Elmwood Point shell bed at the base of the Wanakah Shale (Figs. 8-10). This unit, probably the record of an interval of very low net sedimentation, displays gradational change in biofacies between King Ferry and Portland Point from a small brachiopod and mollusk-dominated assemblage to one containing abundant large spiriferid (Spinocyrtia) and strophomenid brachiopods and large bivalves (Ptychopteria). Thus, the lateral change with this narrow, precisely defined interval corresponds to nearly the full spectrum of biofacies change observed in a major cycle.

The biofacies change is paralleled by a gradation in lithofacies from dark grey slightly silty mud and clay shales to coarse siltstones, with some evidence of winnowing; hence quiet water to near wave base positions are inferred. Based on the relative bathymetric inferences discussed under the biofacies section, we estimate that this might correspond to a depth difference of roughly 30-50 m in about 20 km.

Regional mapping of Ludlowville units demonstrates parallel, upslope gradients in biofacies along the northerly Cayuga to Skaneateles transect as well. The existence of the Fir Tree anticline section at Portland Point permits triangulation of facies belts (Fig. 7). Thus, for example, close resemblance of the Portland Point section to Ludlowville outcrops in the Tully Valley about 45 km (27 miles) to the northeast indicates that the regional depositional strike was approximately northeast-southwest. In contrast, marked facies differences between Portland Point and King Ferry sections only 20 km (12 miles) to the northwest indicate abrupt downslope environmental changes in this direction (Fig. 7). This pattern also explains the intermediate character of the Ludlowville facies at Cascade (Stop 2), relative to King Ferry and Portland Point. In actuality this outcrop is nearly 10 km (6 miles) north of the latitude of King Ferry; however, an imaginary NE/SW strike line projected from Cascade would intercept the Cayuga lake shore some distance southeast of King Ferry Station where the Ludlowville has dipped below the level of the lake, but north of the Fir Tree anticline exposures (Fig. 7). This evidence indicates that, at any given time, biofacies were arranged in elongate parallel belts that trended approximately NE-SW parallel to a gently (~1m/km) northwest dipping ramp.

The precise array of communities was variable depending upon: 1) overall water depth - controlled by transgressive/regressive cycles, and 2) sedimentation rate, as well as other factors. During transgression maxima (e.g. lower Ledyard) the biofacies transect would be as follows (deep to shallow; King Ferry to Portland Point areas): a) Leiorhynchus, b) Ambocoelia, c) Athyris (see Fig. 5).

In contrast, during times of peak sea level lowering the pattern would be either: a) Athyris, b) Pseudoatrypa (diverse brachiopod), c) Spinocyrtia, d) large coral (Pentamerella), if sedimentation rates were low, or: a) Mucrospirifer, b) Tropidoleptus, c) Zoophycos, and d) Allanella, under typical heavier sedimentation. In general, biofacies probably shifted laterally as tracking belts during transgressive-regressive cycles. However, differences in sedimentation between the half-cycles also produced some faunal asymmetries. Thus, the shallowing half-cycles are typified by low diversity Mucrospirifer- or Tropidoleptus-dominated assemblages, whereas the more condensed (sediment-starved) deepening hemicycle may display Athyris, Pseudoatrypa or large rugose coral biofacies, which are more typical of western New York calcareous ("clean-water") facies.

DISCUSSION

In summary, depositional strike within the Ludlowville section trends northeast-southwest nearly normal to the long axes of the Cayuga, Owasco, Skaneateles, and Otisco valleys; there is typically more facies change along these valleys than between them. This perception presents opportunities to examine the key depositional packages and fossil-rich event horizons, both within and between sedimentary cycles, in a new three-dimensional context of inferred Devonian paleobathymetry. It also allows us to predict the lithology and fauna of beds concealed in the subsurface.

A case in point relates to the Ledyard-equivalent Otisco sequence which is nearly entirely below lake level on the Fir Tree Anticline. From our surface mapping work we have found that the Staghorn Coral Bed in the Skaneateles and Otisco Valleys is associated with a spectacular northeast-southwest-trending, northwest-facing submarine escarpment (drop-off) which borders a thick, siltstone platform on which most corals rest (Brett and Baird, 1986). Since this escarpment and coral bed both occur within the lower Otisco Member in those valleys, and since the facies of the upper Ludlowville Formation here at Portland Point has precessed to facies identical to that overlying the Otisco at Staghorn Point, one could predict that the coral bed and possibly the submarine escarpment may be present below Portland Point or below Ithaca. The answer to this must wait because a complete subsurface drill core through the Hamilton section below Portland Point was recently destroyed before we knew of its existence. Future subsurface probes, however, could verify this prediction.

Our recognition of the Hamilton depositional strike trends within the Finger Lakes region, as well as our tentative correlations of numerous sedimentary cycles and event horizons within the Ludlowville and adjacent formations, will enable future students to fine-tune biofacies and taphonomic facies patterns along paleoslope gradients. Gradient analyses of these sequences and coenocorrelation techniques (see Cisne, and Rabe, 1978) will probably be the next steps in paleoenvironmental-paleoecological synthesis of this classic Paleozoic deposit.

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ROAD LOG FOR LUDLOWVILLE FACIES FIELD TRIP

CUMULATIVE MILEAGE	MILES FROM LAST POINT	ROUTE DESCRIPTION
0.0	0.0	Begin road log at junction of NY Route 34 and NY Route 13, at south end of Cayuga Lake in Ithaca. Turn right (north) on Route 34.
1.0	0.5-1.0	Climbing long upgrade on Route 34; note intermittent exposures of latest Middle Devonian Sherburne Siltstone (Genesee Formation); grade is nearly on dip slope of south limb of Fir Tree anticline.
3.25	2.25	Upper end of Shurger Glen.
4.3	1.0	South Lansing; junction Route 34/Route 34B; turn left (northwest) onto Route 34B.
5.0	0.69	Junction Portland Point Road on left (stop 3 is at end of this road); proceed north on Route 34B.
5.2	0.2	Leave town of South Lansing; top of hill
6.0	0.8	Road to Meyers Point at base of long hill.
6.2	0.2	Cross Salmon Creek (original Ludlowville type section).
7.0	0.8	Road to Ludlowville on right.
7.4	0.4	Top of hill above Salmon Creek Valley.
10.3	2.9	Lansing Fire Station.
12.15	1.85	Cross Lake Point ravine.
16.05	3.9	Junction Route 34B/NY Route 90; turn left (west).
17.05	1.0	Jump Corners; junction Route 90 and Clearview Road; turn left onto Clearview Road south of Triangle Diner (90 curves to right).
17.25	0.2	Intersection of Clearview Road with Lake Road; go straight on Clearview.
17.55	0.3	Cross upper King Ferry Creek (which exposes Genesee Formation).

18.25	0.7	Feedlot on left shows exposures of Tully Limestone
18.70	0.45	Sharp bend in Clearview Road.
18.75	0.05	Cayuga Lake shore; junction with lakeshore access road; turn right (north) and proceed (slowly) north. Narrow dirt road follows an old railroad bed along the lakeshore; it is bordered on the shore side by a row of cottages and on the other side by bluffs of Ludlowville shale (and outhouses!).
18.95	0.2	Small gully exposes lower Wanakah (equivalent to type King Ferry Member of Cooper, 1930); from here onward there are intermittent joint face exposures of the lower Wanakah.
19.15	0.2	Small gully; lenses of concretionary, fossil-rich mudstone are visible in wall.
19.25	0.1	More fossil-rich lenses visible in weathered joint face.
19.40	0.15	Stop and park at "Dead End" sign just south of Elmwood Point, for stop 1A.

STOP 1 KING FERRY STATION

Locality: Exposure along lake shore road extending from Elmwood Point, 1.6 miles south to near Cats Elbow Point, King Ferry Station, Cayuga Co., N.Y. (Sheldrake 7.5' Quadrangle).

References: Cooper (1930).

General Description: The south-dipping strata along the Cayuga Lake shore bluffs at King Ferry Station display the complete Wanakah Member (King Ferry Member of Cooper, 1930, in part), which is here somewhat over 30 meters thick (Fig. 8), over a lateral distance of about 1.6 miles. The Spafford and Owasco Members, comprising the uppermost 11 meters of the Ludlowville Formation are exposed at the southern end of this road near Cats Elbow Creek (Stop 1C). The King Ferry locality is the most basinward, and thickest of the sections being examined on this fieldtrip. A progressive northwest transition to an even thicker sequence of poorly fossiliferous, bioturbated, silty shales can be observed along the western margin of Cayuga Lake, as at Big Hollow Creek.

STOP 1A ELMWOOD POINT

Upper Ledyard Member.-- At section 1A we will examine the lower contact of the Wanakah shale at the base of a bed which we here designate the Elmwood Point bed for this locality. The uppermost two meters of the Ledyard Member can be seen below the Elmwood Point bed; the upper contact forms a prominent notch in the bank. These dark grey to nearly black shales

contain common diminutive brachiopods (juvenile? Tropidoleptus, Truncalosis, Ambocoelia), nuculid and Modiomorpha bivalves, orthoconic nautiloids, and Phacops trilobites concentrated on thin bedding plane horizons. This characteristic biofacies, occupying the identical stratigraphic position, can be traced to the west as far as Lake Erie without significant change. A harder, more calcareous bed, about 0.5 meters below the contact, contains a somewhat richer fauna, including the brachiopod Athyris, which is absent to the west, and gives the first hint of facies change which becomes increasingly apparent to the east and southeast, as at Cascade (Stop 2).

Wanakah ("King Ferry") Member.

The base of this member is marked by the very widespread, mollusk-dominated Elmwood Point Bed (equivalent to the Mt. Vernon bed or "Strophalosia" bed; Grabau, 1898-1899; Cooper, 1930), which, at this locality, remains nearly unchanged from its appearance in western New York except for the common occurrence of the brachiopod Mucospirifer and the absence of Truncalosis ("Strophalosia"). A major coarsening-upward cycle, about 18 meters thick, overlying this stratigraphic marker bed, begins with dark grey silty shales and culminates in massive, fretted, Zoophycos-burrowed siltstone. This cycle can be subdivided into three subtle subcycles. The lowest is only about 1.5 meters thick. It consists of dark grey, sparsely fossiliferous shale, at the base, which grades upward into poorly fissile, burrowed, grey silty shales (Fig. 8).

This sequence is overlain by about 3 m of slightly harder silty mudstone, with numerous thin siltstone layers, and a moderately diverse assemblage of brachiopods, including large Spinocyrtia, Cypricardella and Modiomorpha bivalves, trilobites, and the small discoidal tabulate coral Pleurodictyum. Fossils tend to occur in local lenticular (pod-like), somewhat concretionary-aggregates which are visible along a weathered joint face just south of a bridge over a small gully about, 0.1 miles south of the parking area. These pods are interpreted as primary skeletal accumulations which have been enhanced by early diagenesis. Several fossil bands and pods here appear to have been terminated by silty tempestites. The abundance of siltstones in this portion of the section suggests a shallow water position, above storm wave base. These beds are the equivalent of the Pleurodictyum "zone" that can be traced to Lake Erie (Grabau, 1898, 1899; Cooper, 1930).

At Romulus, on the northwest side of Cayuga Lake, this same interval expands to a 4.5 meter-thick, coarsening-upward subcycle of dark grey silty shale with a sparse fauna of nuculids and Cardiola bivalves, orthoconic cephalopods, and small brachiopods and capped by Pleurodictyum-bearing beds.

Return to vehicles and reverse route back to Clearview Road.

20.05 0.65

Junction Clearview Road; continue south on the lakeshore road to STOPS 1B and 1C.

- | | | |
|------|-----|--|
| 20.1 | .05 | Cross King Ferry Creek; first outcrop south of the creek; shale here is about 6 m above the top of STOP 1A; a biostrome of rhomboporid and fenestellid bryozoans occurs here just above a thin siltstone, capping a second subcycle. |
| 20.4 | 0.3 | Pull off on wide area on left side of road adjacent to shale bank for STOP 1B. |

STOP 1B KING FERRY STATION (SOUTH)

This bluff section displays the top of the lower Wanakah (Aurora submember) major cycle, which here forms a distinct bluff of massive siltstone; this unit shows a sharp upper platform because it is abruptly overlain by softer shale. The upper meter of the siltstone contains large spheroidal concretions associated with two lenticular, fossil-rich layers; the upper one, about 70 cm below the upper bench, contains the large rugose corals Heliophyllum and Cystiphyllodes, together with a diverse assemblage of large bivalves, brachiopods, and bryozoans (Fig. 8). This appears to record maximal shallowing within the entire Wanakah. Winneding and bypass of fine-grained-sediments produced a relatively clean, coarse-silt substrate which favored sporadic colonization by large corals.

An irregular, dense coquinite of brachiopods and Stereolasma corals immediately overlies the prominent siltstone bench and shows a hint of a minor coarsening upward cycle about 50 cm above that bench. This interval, here termed the Ensenore Ravine shell bed, is believed to

Figure 8.--King Ferry Station Section. Note subtle development of shallowing subcycles, seen mainly in faunal change and upward coarsening from mudstone to silty mudstone or siltstone. This Ludlowville sequence is divisible into six sedimentary packages, corresponding to key divisions elsewhere (see text); these are: 1, Ledyard Member; 2, the lower Wanakah (Aurora Submember) which is equivalent to the Darien Center Submember (Mt. Vernon Bed to Murder Creek Bed interval; including the "Pleurodictyum zone" and "trilobite beds" of Grabau, 1899) in western New York; 3, short interval in the medial Wanakah capped by the Barnum bed of the central Finger Lakes region (see Baird, 1981); 4, overlying cycle capped by siltstone bed correlative with the Mack Creek bed; 5, upper Wanakah siltstone division; capped by the Bloomer bed; 6, coarsening-upward cycle of Spafford Shale Member. Numbered beds include: a, Elmwood Point shell bed; b, silty interval with lenticular fossil layers; c, Rhombopora-rich beds, overlying coarsening-upward subcycle; d, siltstone division of Aurora Submember with large rugose corals and large iron-stained concretions; e, Barnum bed; f, Mack Creek bed; g, Bloomer Creek bed interval; h, shaley bed with rich Jaycox-like fauna; i, sub-Moscow regional disconformity; j, Portland Point Member.

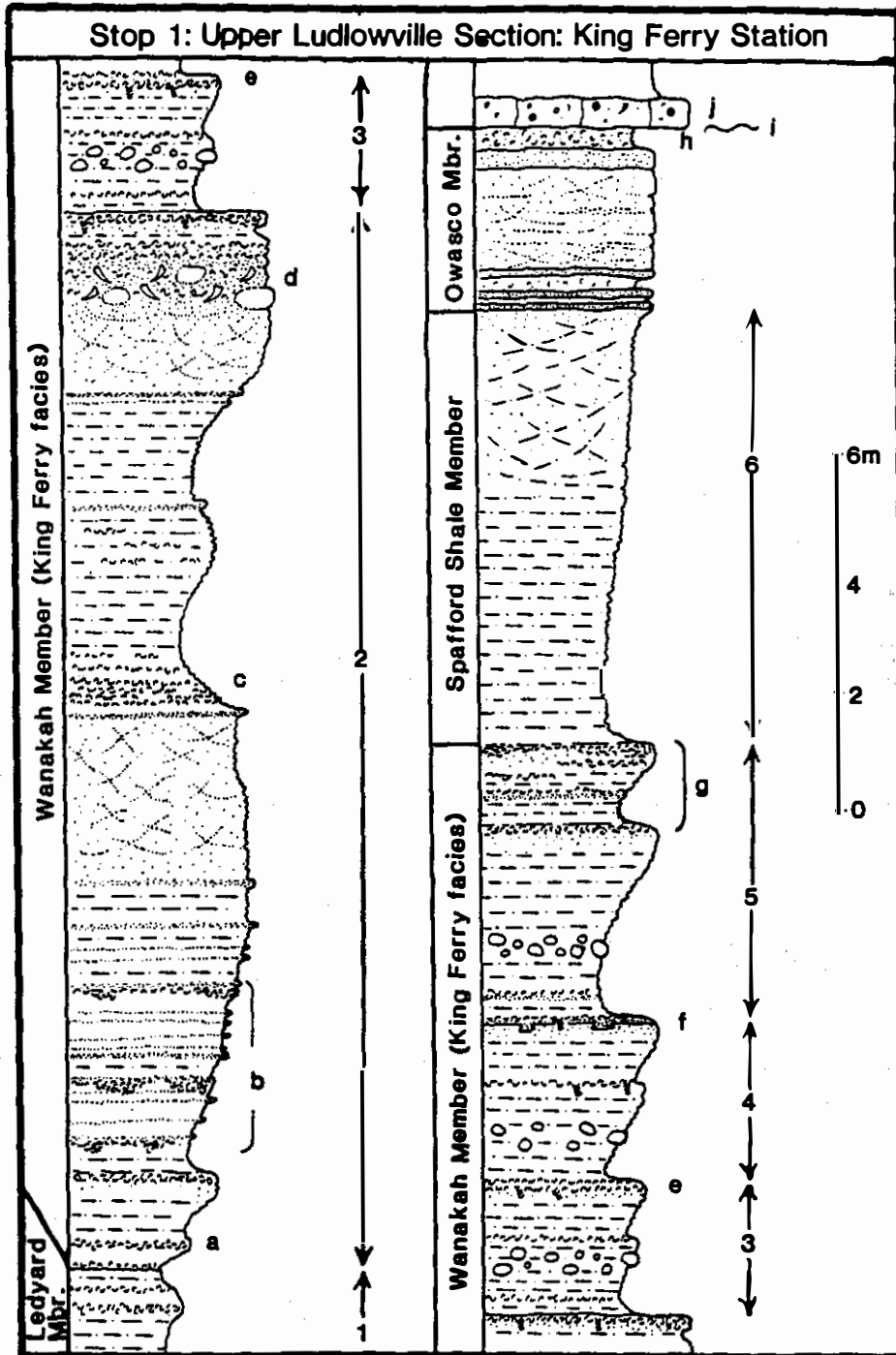


Figure 8.--King Ferry Station Section: See facing page for caption.

correlate with Grabau's (1898-1899) lower trilobite bed (Murder Creek bed of Kloc, 1983) in western New York, with which it bears striking faunal similarity.

Three other major coarsening-upward cycles, each with minor subcycles, occur within the remaining 9.5 m of the Wanakah Member (Fig. 8). The lowest of these sequences will be examined by walking about 0.15 miles south of the parking area. Each subcycle has a well developed interval of irregular, commonly rust-stained concretions near its base and is capped by a thin, silty brachiopod-dominated coquinite. These coquinites display scoured bases with pods or gutter casts and nested stacks of convex-upward shells.

The lower shell cap bed can be correlated with the Barnum Creek bed, the upper thick coquinites with the Bloomer Creek bed, both of which become subtle, but widespread, shale-on-shale, hiatus concretion horizons to the northwest (Baird 1981; Baird and Brett, 1981).

The middle shell bed level is capped by a 2-4 cm thick cross laminated siltstone which is traceable to the west side of Cayuga Lake as the Mack Creek turbidite (Baird, 1981).

Return to vehicles and continue south along the lake road.

20.6	0.2	Cross Cats Elbow Creek, which exposes a section of the Wanakah with overlying units.
20.65	0.05	Yellow 10 mph signpost; note excellent exposure of the Mack Creek turbidite siltstone.
20.85	0.2	Cross unnamed creek.
21.00	0.15	End of road; pull off in parking area and walk to shale bank behind cottages for STOP 1C.

STOP 1C CATS ELBOW POINT (SOUTH)

This bank provides an excellent view of the top of the Wanakah Member, here a sharp contact at the top of a 30 cm-thick, shell-rich argillaceous limestone, the Bloomer Creek bed. This bed caps a second prominent shallowing cycle in the upper Wanakah Shale; mimicking the Ensenore Ravine bed seen at STOP 1B; like the Ensenore bed the Bloomer bed contains a highly diverse brachiopod, bivalve, bryozoan assemblage with scattered large corals (Fig. 8). It correlates with a very shell-rich interval that is also traceable to Lake Erie (the Stictopora - Demissa zone of Grabau, 1898-1899; Blasdell beds of Kloc, 1983).

The Bloomer bed is sharply overlain by sparsely fossiliferous, dark grey shale of the lower Spafford Member. These shales coarsen upward into silty mudstone and siltstone, visible in the upper portion of this cliff

section. Upper beds are 10 to 30 cm-thick, blocky, brownish-weathering, laminated siltstones with large concretions closely resembling the capping beds of the lower Wanakah (Aurora) cycle. This interval appears to be the northwestern equivalent of Smith's (1935) Owasco Member. It is separated from the Tichenor limestone in Cats Elbow Creek by about 40 cm of bluish grey mudstone with a very diverse fauna resembling the Jaycox Member west of Cayuga Lake. This section is critical in establishing a link between the lower Jaycox Member and the Owasco Member.

SYNOPSIS OF STOPS 1A - 1C.

In summary, this sequence of outcrops reveals the beginnings of a change within the upper Ludloville Formation from uniform dark grey to black shale with thin shell beds to a series of asymmetrical, shallowing-upward, muddy siltstone cycles and subcycles. In ascending order these cycles and their capping shelly or silty beds are: A) upper Ledyard: Elmwood Point bed; B) lower Wanakah (Darrien Center- Aurora submember): Ensenore Ravine bed (and including three subtle subcycles); C) lower-middle Wanakah: Barnum Creek bed; D) middle Wanakah: Mack Creek bed; E) upper Wanakah: Bloomer Creek bed; F) Spafford Member: Owasco Siltstone. Most of these cycles are traceable, at least in western and central New York State, and appear to record allocyclic fluctuations in relative sea level.

Reverse route and return to Clearview Road.

22.0	1.0	Junction Clearview Road; turn right (east) and retrace route uphill.
23.55	1.55	Triangle Diner and Jump Corners; proceed straight (gentle right) onto Route 90 east.
24.5	1.0	Town of King Ferry. Junction Route 90/Route 34B; proceed straight on Route 90.
25.8	1.3	Cross Little Salmon Creek.
27.6	1.8	Town of Genoa; cross Salmon Creek.
28.1	0.5	Junction Route 90/Route 34.
30.7	2.6	Junction Pine Hollow Road, Route 90 bends to the right. Go straight (gentle left) onto Pine Hollow Road.
31.9	1.2	Overlook into Owasco Valley.
33.3	1.4	Junction Route 90/Route 38, turn left (north) onto Route 38.
33.8	0.5	Cross Owasco Inlet Creek.
36.0	2.2	Fillmore Glen State Park on right.

36.1	0.1	Town of Moravia.
37.0	0.9	Junction of Route 38/38A; turn left staying on Route 38.
37.4	0.4	Cross Owasco Inlet.
40.3	2.9	Minor Otisco shale outcrop.
40.9	0.6	Town of Cascade; pull off on shoulder opposite road cut at junction of small side road and Route 38. Cautiously cross the highway for STOP 2A.

STOP 2. CASCADE; ROUTE 38 ROADCUT

Locality: Long roadcut along west side of N.Y. Route 38, on the west side of Owasco Lake Valley, town of Cascade, Cayuga Co., N.Y. (Moravia 7.5' Quadrangle).

Reference: Baird (1979).

Description: This roadcut provides an excellent exposure of the entire upper Ludlowville section from the upper Otisco Member through the Tichenor Limestone (Fig. 9). Most striking at this location is the clearly developed repetitive sequence of coarsening-upward cycles, capped by very diverse brachiopod-bivalve coquinites. These easily accessible beds, with extensive bedding plane exposures, are ideal for collecting a nearly complete suite of Hamilton fossils. The hierarchical pattern of subcyclic units is also very well displayed, here as are the iron-stained, concretionary intervals occurring toward the bases of the major coarsening upward cycles.

STOP 2A ROUTE 38 ROADCUT, CASCADE (LOWER)

Otisco Member and Elmwood Point Bed

The upper Ledyard Member-equivalent, here designated as upper Otisco Member, has changed noticeably in both litho- and biofacies, from dark, small brachiopod and mollusk-dominated shales seen at King Ferry to medium grey brachiopod-rich, silty shales here. Abundant Mucrospirifer and Athyris are associated with the bivalves Cypricardella and Modiomorpha, together with the more typical upper Ledyard nuculid bivalve-diminutive Tropidoleptus fauna.

The Elmwood Point bed can be seen at the top of the exposure along a side road intersecting Route 38. It is represented by a silty coquinite layer, up to 10 cm thick, with an irregular base (Fig. 9), which contains a moderately diverse assemblage dominated by the brachiopods Mucrospirifer; Mediospirifer and Athyris with associated strophomenid brachiopods and pterioid bivalves. The striking change from its appearance at King Ferry may be due, in part, to the condensation of the lowest, thin subcycle present at King Ferry onto the Elmwood Point bed continuing the eastward thinning trend from northwestern Cayuga Lake (Big Hollow Creek).

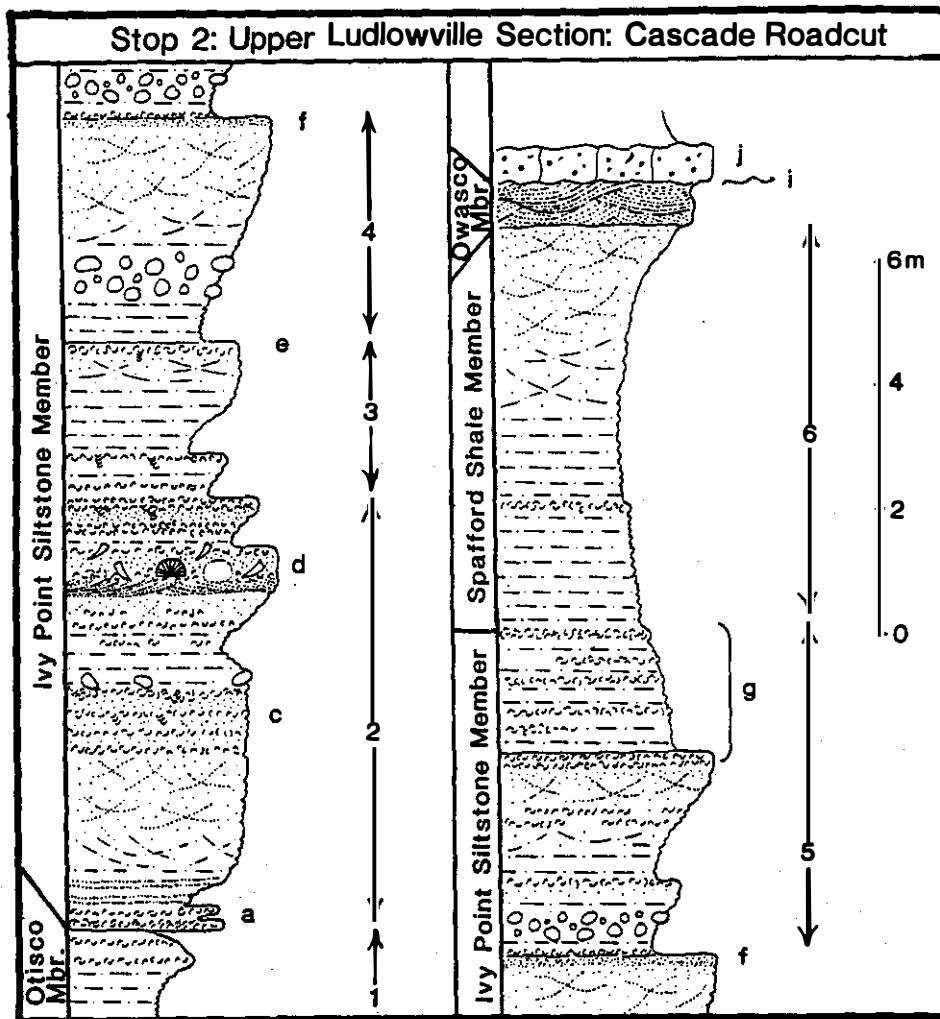


FIGURE 9.--Route 38 roadcut at Cascade. Note distinct upward-coarsening mudstone-to-siltstone cycles, capped by shell beds; also note distinct thinning of the lower Wanakah/Ivy Point interval as compared to that at King Ferry. Numbered intervals include: 1, Otisco Member; 2, lower Ivy Point/Wanakah (Aurora Submember); 3, short interval in the medial Ivy Point/Wanakah capped by the Barnum bed of the Central Finger Lakes region (see Baird, 1981); 4, overlying cycle capped by siltstone bed correlative with the Mack Creek bed; 5, upper Ivy Point/Wanakah siltstone division; capped by the Bloomer bed; 6, coarsening upward cycle of Spafford Shale Member. Numbered beds include: a, Elmwood Point shell bed; c, brachiopod-rich shell layers corresponding to *Rhombopora*-rich beds at King Ferry; d, siltstone division of Aurora Submember with large rugose corals and the hemispherical tabulate *Favosites hamiltoniae*; e, Barnum Creek bed; f, Mack Creek bed; g, Bloomer Creek bed interval; i, sub-Moscow regional disconformity; j, Portland Point Member.

- 41.3 0.4 Reboard vehicles and proceed uphill for 0.4 miles; prepare to make a u-turn across Route 38, at crest of hill; roadcut on west side of road is in upper Ludlowville and overlying Tichenor ("Portland Point") interval; reverse route downhill (south) for 0.2 miles.
- 41.5 0.2 Pull off along Route 38 at point opposite a small creek gully, disembark for STOP 2B.

STOP 2B ROUTE 38 ROADCUT, CASCADE (UPPER)

Wanakah/Ivy Point Member

This interval is dominantly muddy siltstone and is, thus, intermediate between the typical Wanakah and Ivy Point Members. Above the Elmwood Point bed the major lower cycle has both coarsened and thinned markedly to the east, being only about 7.5 meters thick at Cascade compared to over 17 meters at King Ferry (compare Figs. 8 and 9). This illustrates well our observation that sediment packages thicken basinward due to the rapid dumping of finer grained sediments bypassed from more proximal, wave swept environments.

Subcyclicity within the major lower cycle is more pronounced here than at either King Ferry or Portland Point (STOPS 1, 3). The lowest subcycle comprises 4-meters of highly bioturbated and sparsely fossiliferous silty mudstone contains several irregular, poddy fossil layers with a large brachiopod-bivalve fauna (including: Mucrospirifer, Mediospirifer, Athyris, Spinocyrtia, Protoleptostrophia, Tropidoleptus, Modiomorpha, Cypricardella, and Actinopteria). These beds likely correlate with the rhomboporid-rich interval at King Ferry. Above this is a 2-meter thick subcycle which coarsens upward into massive, locally hummocky cross stratified, coarse siltstone, indicating shallowing into normal wave base. Two prominent coraliferous horizons are present within the coarse upper unit which contain Cystiphyllodes, Heliophyllum, and large hemispherical colonies of Favosites hamiltoniae. Certain of the solitary rugose corals show evidence of corrosion and reworking. This is clearly equivalent to the coral-bearing cap of the lower cycle at King Ferry. A thin (<1 meter) subcycle follows which is capped by a thin dense coquinite with an extremely diverse fauna. This bed is believed to be equivalent with the Ensenore bed which forms a prominent cap to the major lower cycle to the west as far as Sheldrake Creek on the west shore of Cayuga Lake.

The cycles capped by the Barnum, Mack, and Bloomer Creek beds are all siltier and more prominent than at King Ferry. Unlike the major cycle below, their thicknesses show little change from King Ferry; and the upper two cycles are even somewhat thicker. Most subcycles continue to be recognizable, and the upper two concretionary intervals are well developed. Abundant pyritized burrow tubes occur below the Barnum bed in a position analogous to the concretionary interval at King Ferry. The most easily identified interval in this cut is the rust-stained concretionary horizon below the Bloomer Creek shell bed. Each of the

capping shell beds contains a diverse fauna, dominated by Athyris or Mucrospirifer.

Spafford Member and Owasco Members

The upper part of the Route 38 roadcut, near the crest of the hill, exposes the upper members of the Ludlowville Formation, and its contact with the overlying Moscow Formation (Fig. 9). The Bloomer bed at the top of the Wanakah/Ivy Point Member is not sharply defined but is represented by a series of closely spaced, discontinuous fossil coquinities with a very rich and diverse fauna, characterized by Athyris, Spinocyrtia, Pseudoatrypa, and Strophodonta demissa; these beds are considered to mark the top of the Ivy Point Member. The overlying the Spafford Shale Member is somewhat thinner than at King Ferry and contains a more diverse fauna, particularly typified by chonetid brachiopods and Tropidoleptus. The overlying Owasco Member displays marked coarsening and thinning due to the erosional truncation of the upper siltstones and Jaycox-like fossiliferous shales. We interpret this as the result of south-eastward downward cutting of the Tichenor Limestone Member which comes to rest directly on the Spafford at Portland Point. The Owasco shows distinctly more proximal facies than at King Ferry with Allanella-rich, sandy layers and hummocky cross-stratification.

The highest unit exposed in this part of the roadcut is the "Portland Point" Member; the Tichenor Limestone-equivalent is a coarse, crinoidal, coral rich packstone or grainstone. This unit is now considered to form the base of the Moscow Formation (Baird, 1979).

		Walk back downhill to vehicles; reboard and continue southward on Route 38.
46.7	5.2	Moravia. Turn right to follow Route 38; continue south.
50.45	3.75	Town of Locke; Junction Route 90 turn left (southwest) onto 90.
51.75	1.3	Junction road to West Groton; stay on Route 90.
51.95	0.4	Y-junction of Route 90 and Lamphier Road, North Lansing; turn left (west) on Lamphier Road; name changes to Lane Road.
54.7	2.8	Junction Route 34, town of North Lansing; turn left (south) on combined Route 34-34B.
61.25	6.65	Route 34 bends sharply to the right (west)
62.25	1.0	Junction Route 34/34B; turn right (north) onto Route 34B.

62.9	0.7	Junction road to Portland Point; turn left (west) onto Portland Point Road.
63.4	0.45	Entrance to Portland Point Quarry (in Tully Limestone) on left.
63.5	0.1	Pulloff on right is next to Minnegar Falls over Tully Limestone.
63.8	0.3	Pit in dark lower Windom Shale on left.
64.0	0.2	Cargill Salt Company mine on right; small Ludlowville outcrop on left.
64.25	0.25	Low outcrops of Ludlowville (upper Ivy Point) on left; pull off in parking area on right, just before white-painted building north of abandoned cement plant, for STOP 6. Proceed on foot from parking area to railroad track just below (west of) the road; walk northwest along railroad track for about 0.1 miles to large roadcuts at STOP 3A.

STOP 3A PORTLAND POINT; NORTHERN RAILROAD CUT

Locality: Exposures on railroad cut along east-shore of Cayuga Lake between Cargill Salt Company and abandoned cement plant, 0.1 to 0.3 km north of Portland Point, town of Lansing, Tompkins Co., N.Y. (Ludlowville 7.5' Quadrangle).

Reference: Patchen and Dugolinski (1979).

Description: This weathered railroad cut exposes about 12 m (40') of the lower Ivy Point Member and the underlying upper Otisco Shale Member near the crest of the east-west trending Fir Tree Anticline (Fig. 10). These are the oldest units exposed at the axis of this fold. This section and that in the adjacent mouth of Shurger Glen afford a direct look at facies far south of the normal east-west outcrop belt some 10 to 15 miles to the north of this area. The southward Ludlowville facies change (distinct coarsening of beds with pervasive shoreward biofacies transitions) along the Cayuga Valley clearly mirrors the eastward facies changes from the Aurora-King Ferry area to the Skaneateles Valley, and we use the terminology that Smith (1935) proposed for the Ludlowville Formation in the Skaneateles Valley in describing the section at Portland Point.

Otisco Shale Member

About 5 m of the upper Otisco Member consists of medium grey, highly fossiliferous, soft silty shale. Shell beds, especially near the top, yield abundant brachiopods, particularly Athyris, Tropidoleptus, Spinocyrtia, small bryozoans, and the rugose coral Stereolasma.

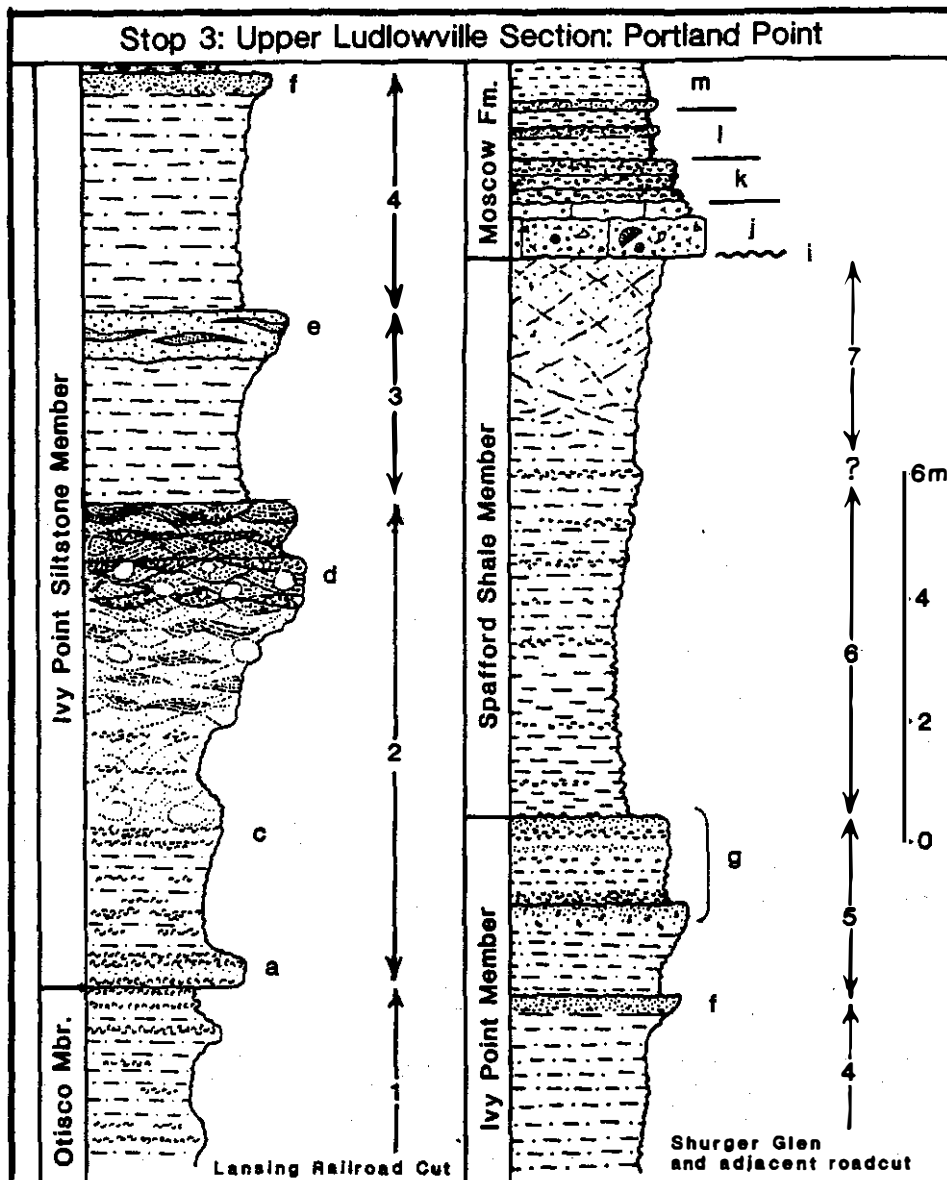


FIGURE 10.--Portland Point Ludlowville section at crest of Fir Tree anticline. Ludlowville stratigraphic terminology adapted from Smith, (1935). Note prominent development of regressive hemicycles capped by discontinuities and/or winnowed, condensed beds; also note that the upper beds of the lower Ivy Point tongue (interval d, equivalent to the coral-bearing upper beds at King Ferry and Cascade) is marked by hummocky cross-stratification, large concretions and coquinites of Allanella tullius. Based on faunal content, the upper part of the Spafford appears to match with the lowermost Jaycox Member in western New York and with beds immediately beneath the Owasco Sandstone in the Owasco to Tully Valley region. The Portland Point Member (j) rests directly on the Spafford, and is overlain by: k, the Rhipidomella-Centronella ("R-C") Bed of Kashong Shale Member; l, the upper Kashong beds including phosphatic pebble layer (see Baird, 1978, 1979); and m, the Windom Shale Member. (For explanation of other symbols see Figure 9).

Ivy Point Siltstone Member

The upper boundary of the Otisco shales with the Ivy Point can be located at the base of a distinctive 65-70 cm (2.1 ft.)-thick silty ledge that contains abundant large brachiopods and bivalves. This unit is the Elmwood Point bed noted at both previous stops. Here the bed contains large brachiopods, such as Spinocyrtia, and rugose corals; hence, in comparing the three sections, this bed displays a gradient of increasing diversity and a change from ambocoeliid to Athyris to Spinocyrtia biofacies. This bed apparently formed by sediment condensation along a paleoslope and thus it provides a natural transect of paleocommunities.

Above the Elmwood Point bed is an 8 meter- (26 ft.)-thick major coarsening-upward cycle which we term the lower siltstone tongue of the Ivy Point Member. It is clearly the interval equivalent of the lower Wanakah (or Aurora submember) cycle seen at Stops 1 and 2, but it is very distinctly coarser-grained. This sequence weathers with a distinctly recessed area near the base, where softer mudstone deposits overlie the Elmwood Point bed (Fig. 10). A remnant of the cap of the middle subcycle at King Ferry Station can still be recognized in several thin, Mucrospirifer- and Tropidoleptus-dominated shell beds associated with subconcretionary pods. Higher units include massive Zoophycos-churned silty mudstones; these, in turn, grade upward into laminated coarse siltstone, displaying evidence of local channeling.

The upper portion of the lower Ivy Point cycle consists of hummocky cross-stratified coarse siltstone and fine sandstone. Joint surfaces in the intensely bioturbated siltstone display a distinctive "fretwork"-type of differential weathering, controlled to some extent by textural and/or compositional differences between sediment within-and surrounding Zoophycos spreiten. Several layers of shell coquinite, composed mainly of the brachiopods "Allanella" tullius, Camarotoechia and minor shell hash, are visible near the top (Fig. 10). Large (up to 0.5 m diameter) calcareous concretions occur within a prominent 1.0 m thick, massive, buff-weathering coarse siltstone bed. This coarse bed caps the major regressive cycle and it is correlative with the coral-bearing siltstone bed at the top of the lower Wanakah cycle at King Ferry and Cascade (STOPS 1, 2); however, no corals have been found here, probably because this facies represents too unstable an environment. A meter-thick uppermost subcycle overlying the concretionary bed displays abundant fossil debris (mainly Allanella, Tropidoleptus, and Mucrospirifer) and is probably equivalent to the Ensenore Ravine bed; this subcycle is also capped by a coarse, slightly concretionary siltstone. Finally, this unit is overlain by softer grey and sparsely fossiliferous shale which grades upward, over about two meters, into another siltstone, probably the Barnum bed-equivalent.

Above the railroad cut, along the access road to Portland Point, a higher, but somewhat similar coarsening-upward cycle marks the position of the upper siltstone tongue of the Ivy Point Member, of Smith's (1935) terminology. This upper bench is capped by about half a meter of extremely shell-rich mudstone with abundant brachiopods, which is clearly correlative with the Bloomer Creek shell bed.

Walk back along the railroad and cross the Portland Point road at small bridge over Gulf Creek; walk upstream along the north bank of the creek for about 100'.

STOP 3B GULF CREEK (SHURGER GLEN)

Locality: Exposures along the gorge of Gulf Creek 0.1 km east of its mouth at Portland Point, Tompkins Co., N.Y. (Ludlowville 7.5' Quadrangle).

References: Cooper (1930); Baird and Brett (1981); Patchen and Dugolinski (1979).

Description of Units: The uppermost 11 meters (36') of the Ludlowville formation are exposed below a falls capped by Tichenor Limestone; these silty shales display a peculiar curved jointing pattern in the falls face (Fig. 10).

The top of the Ivy Point (or King Ferry) exposed near stream level, displays to advantage the Bloomer Creek shell bed which consists of two to three closely-spaced coquinite layers. Fossils at this level include the abundant brachiopods (e.g. Athyris, Pseudoatrypa, Spinocyrtia, Strophodonta demissa), bivalves (e.g. Cornellites, Cypricardella, and Modiomorpha), several bryozoans, and echinoderms, including the grapnel-like holdfast Ancyrocrinus and rare edrioasteroids. About 9 meters (29') of overlying, less fossiliferous, silty mudstone with occasional layers of coquinite is equivalent to Smith's (1935) Spafford Shale Member. The highest Ludlowville member (Owasco Sandstone), if ever present here, was apparently removed by the major sub-Tichenor erosion surface.

At the cap of the falls is Cooper's (1930) "Portland Point" Limestone Member, a very thin (0.9 m), condensed interval, with a basal (Tichenor-equivalent), crinoidal packstone ledge, and a middle silty calcareous shale division, which is equivalent to the Deep Run and Menteth Members of the Moscow Formation (Baird, 1979). Overlying this unit is another shell-rich bed (Rhipidomella-Centronella or "R-C" bed of the Kashong Member). A complete Windom Shale section and falls over the Tully Limestone are observable upstream from the first waterfalls.

DISCUSSION (STOPS 3A AND 3B): The Lansing railroad cut and adjacent Gulf Creek section display pervasively silty Ludlowville deposits which have a strong similarity to the upper Ludlowville sequence in the Skaneateles Valley near Spafford; Smith's (1935) Ivy Point and Spafford successions of the Skaneateles Valley are clearly discernable here. This similarity holds both for lithofacies and biofacies. It should also be remembered that most regional facies changes, and resulting difficulties of correlation, within the Ludlowville interval are encountered within 35 km (20 mi) along the northwest-southeast trending Cayuga Valley. Sparsely fossiliferous grey shales in the lower Wanakah section at Romulus transform southeastward to sandstones at Portland Point, and black shales in the northwestern upper Wanakah sections change to shell-rich grey silty mudstone lithology in this area.

End of field trip.

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THE USE OF JOINT PATTERNS
FOR UNDERSTANDING THE ALLEGHANIAN OROGENY
IN THE UPPER DEVONIAN APPALACHIAN BASIN,
FINGER LAKES DISTRICT, NEW YORK

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REGIONAL SIGNIFICANCE

Abundant evidence (deformed fossils) for layer parallel shortening in western New York indicates the extent to which the Alleghanian Orogeny affected the Appalachian Plateau (Engelder and Engelder, 1977). In addition to low amplitude (< 100 m) long wave length (< 15 km) folds the Upper Devonian sediments of western New York contain many mesoscopic-scale structures including joints that can be systematically related to the Alleghanian Orogeny.

Based on the nonorthogonality of cleavage and joints, the Alleghanian Orogeny in the Finger Lakes District of New York consists of at least two phases which Geiser and Engelder (1983) correlate with folding and cross-cutting cleavages in the Appalachian Valley and Ridge. To the southeast of the Finger Lakes District the earlier Lackawanna Phase is manifested by formation of the Lackawanna syncline and Green Pond outlier and the development of a northeast-striking disjunctive cleavage within the Appalachian Valley and Ridge mainly from the Kingston Arch of the Hudson Valley southwestward beyond Port Jervis, Pennsylvania (Fig. 1). Within the Finger Lakes District, New York, a Lackawanna Phase cleavage is absent; and one finds instead a cross-fold joint set which is consistent in orientation with a Lackawanna Phase compression. In bedded siltstone-shale sequences (i.e. the Genesee Group) this cross-fold joint set favors development in the siltstones (stops 1# and 2#). The Main Phase is seen as the refolding of the Lackawanna syncline and Green Pond outlier, as well as the development of the major folds in central Pennsylvania. Main Phase structures within the Finger Lakes District include an east-west striking disjunctive cleavage in the Tully Limestone (stops 4# and 5#), a pencil cleavage in the Genesee shales (stop 4#), deformed fossils in the Genesee Group (stop 1#), and cross-fold joints which are orthogonal with the cleavage and deformed fossils in shales (stop 1#) (Engelder and Engelder, 1977; Engelder and Geiser, 1979, 1980, 1984). Outcrops of Tully Limestone at Ludlowville (stop 5#) contain a Main Phase disjunctive cleavage which truncates nonorthogonal Lackawanna Phase cross-fold joints, showing that the Lackawanna Phase predates the Main Phase (Engelder, 1985).

Many outcrops of the Appalachian Plateau contain more than one cross-fold joint set (Fig. 2). Those cross-fold joints attributed to the Lackawanna Phase strike counterclockwise from those attributed to the Main Phase. Cross-cutting cleavages in northeastern Pennsylvania have the same relationship with the later shortening event clockwise from the earlier event (Fig. 1). One exception to this rule was discovered recently by Scott (1986) who has documented two shortening directions in the Tully Limestone at Portland Point which is 3 km SSE of Ludlowville. At Portland Point the later shortening direction appears to be counterclockwise from the direction of the earlier shortening event.

LITHOLOGICAL CONTROL OF JOINTING

This field trip will examine the relationship between lithology and jointing. During a study of regional joints in the vicinity of Ithaca, New York, Sheldon (1912) recognized that certain joint sets favored certain lithologies. Strike joints were common in shales but less well-developed in interfingered siltstone beds. In the same region of the Appalachian Plateau, Parker (1942) noted that plumose markings were rare on strike joints but commonly occurred on cross-fold joints. These studies and those elsewhere (e.g. Stearns, 1968; Nelson and Stearns, 1977) make it clear that the host lithology is an important parameter in influencing the development of regional joint sets as well as their surface morphology.

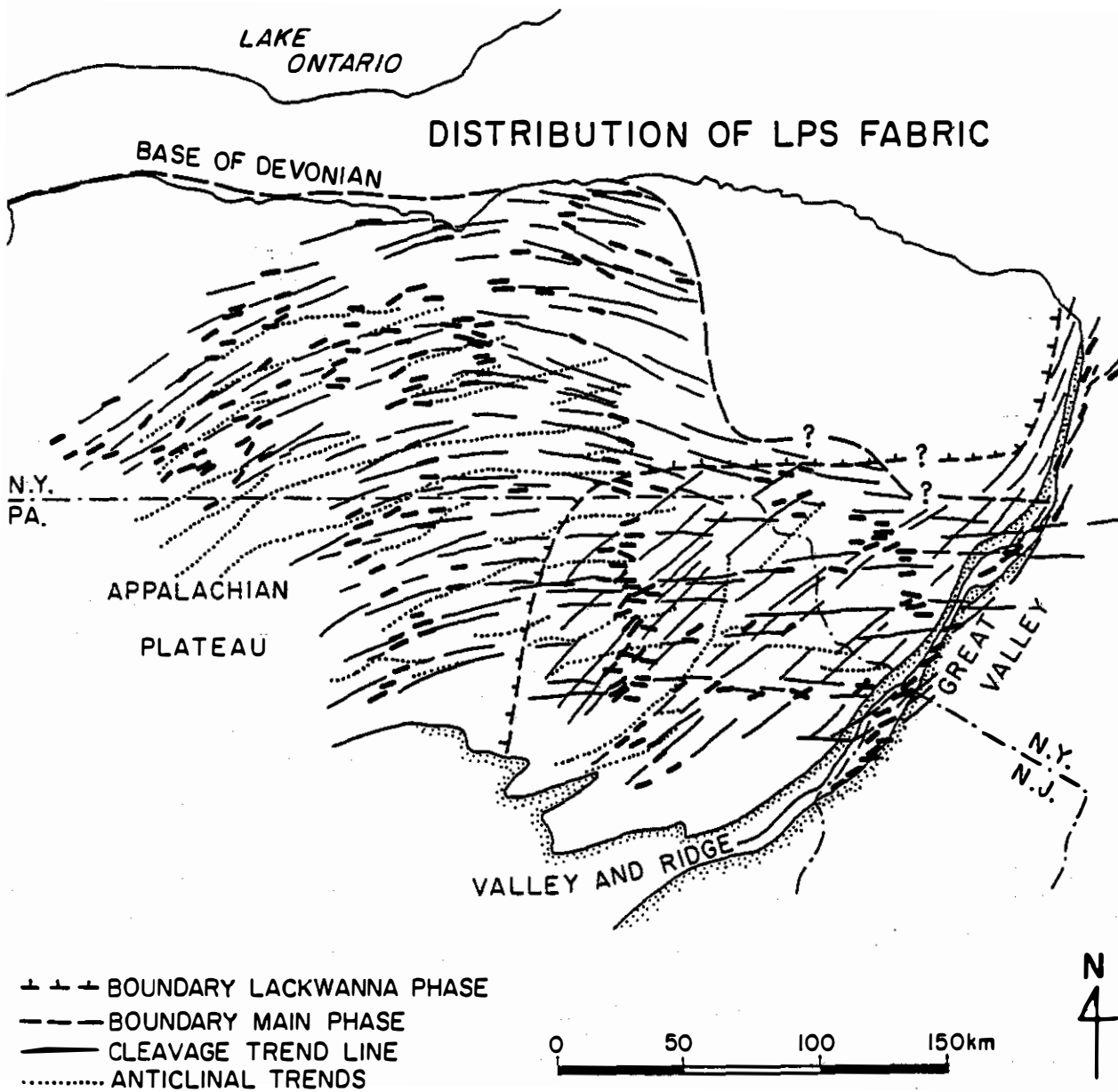


Figure 1. The distribution of layer-parallel shortening (LPS) fabrics across the Appalachian Plateau of New York. The trend-line map was prepared by connecting data points (thick lines) with nearly parallel cleavage planes. The orientation of the cleavage planes is shown by a plot of the strike of cleavage planes (after Geiser and Engelder, 1983).

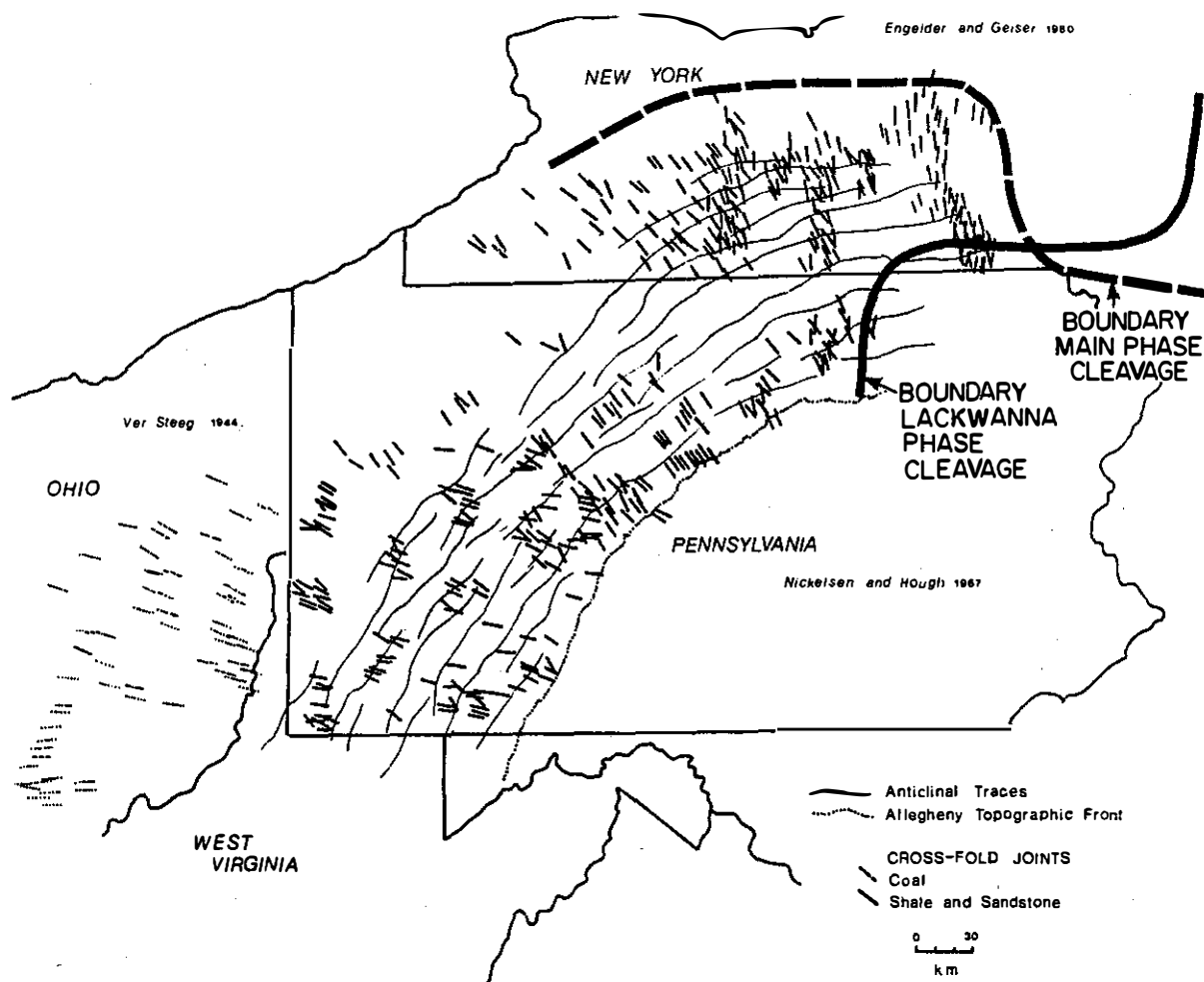


Figure 2. The distribution of cross-fold joints and cleavage within the Central Appalachians (after Engelder, 1985). Cleavage, imprinted during the Lackawanna Phase of the Alleghanian Orogeny, affected the area south of the solid line whereas cleavage imprinted during the Main Phase of the Alleghanian Orogeny affected the area south of the dashed line.

The five stops of this trip will view the same stratigraphic section that Sheldon (1912) examined. The major units include the Genesee Formation (deWitt and Colton, 1978) or Genesee Group (Van Tyne, 1983), the Tully Limestone, and the Hamilton Group (Van Tyne, 1983). Figure 3 shows deWitt and Colton's (1978) stratigraphic column for the Genesee Formation. The first three stops will examine the Ithaca Member of the Genesee Formation. Stop 4# will examine the Tully Limestone, Genesee Shale Member, and the Penn Yan Shale Member. Stop 5# will examine the Moscow Shale of the Hamilton Group and the Tully Limestone.

Three general lithologies will be examined: a bedded siltstone-shale sequence (the Ithaca Member), a homogeneous shale (the Genesee Shale Member and Moscow Shale) and a limestone (the Tully Limestone). Each of these three lithologies has a characteristic joint-orientation pattern and characteristic surface morphology. In addition, surface morphology varies among the various cross-fold joint sets discussed in Sheldon (1912), Parker (1942), and Engelder and Geiser (1980). Surface morphologies have been used to make the case that the cross-fold joint sets mapped by Engelder and Geiser (1980) are fundamentally different from each other.

Genesee Formation

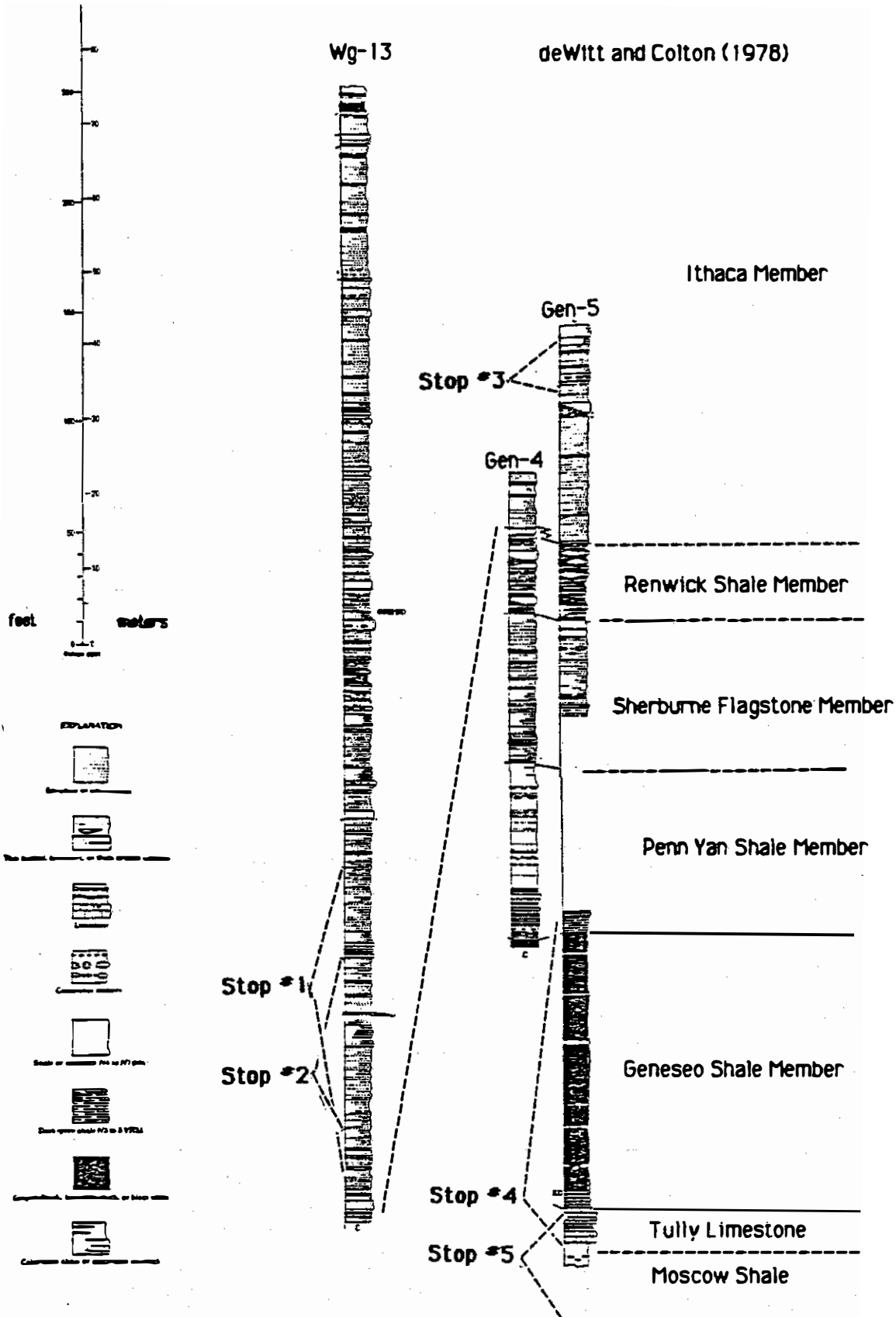


Figure 3. Genesee Formation sections Wg-13, Gen-4 and Gen-5 (after deWitt and Colton, 1978). Stops #1, #2, #3, #4, and #5 are shown in the Ithaca Member, Genesee Shale Member, Tully Limestone, and Moscow Shale respectively.

Within the Appalachian Plateau evidence is overwhelming that the joints formed in extension rather than shear (Engelder, 1982). Surface morphology on joints is distinct from slickensided surfaces on shear fractures. The pattern on joints is called a plumose structure; it constitutes all "delicate tracery of feathery lines" (Woodworth, 1896) diverging from either a straight or sinuous axis. Plumose patterns form during the propagation of a crack (joint) with motion on the crack face normal to the plane of the crack. By the late 19th century geologists recognized that the feather (plume) patterns on joints contained information about the process of joint propagation (Woodworth, 1896). Plumose patterns form on the surface of extension fractures (joints) where the plume records the development of the joint whose rupture front is perpendicular to the barbs of the plume (Fig. 4). Despite the large number of descriptions of the markings on joint faces (Hodgson, 1961; Roberts, 1961; Syme-Gash, 1971; Kulander et al., 1979) none adequately distinguishes the end members of the family of plumose markings observed on the Appalachian Plateau.

STOP 1: LITHOLOGICAL CONTROL OF CROSS-FOLD JOINTS IN THE UPPER GENESEE GROUP, WATKINS GLEN, NEW YORK

This roadcut is best viewed in the late morning when the sun strikes the joint surfaces at a high angle. After about 11:30 AM when the joint surfaces no longer receive direct sunlight, the surface morphology is far more difficult to see.

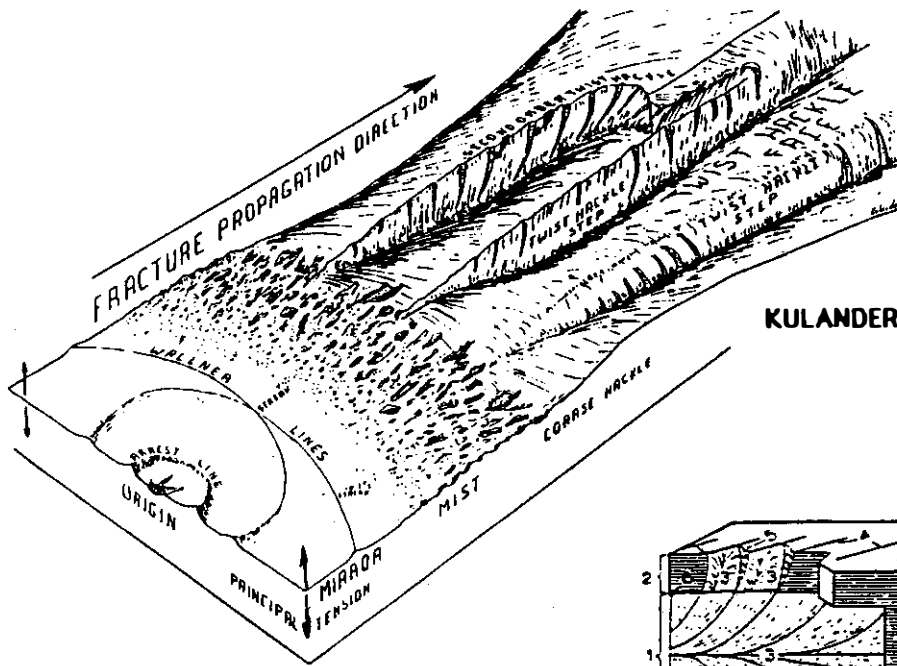
As this roadcut at Watkins Glen was excavated, benches were carved out by taking advantage of the jointed rock. The base of the roadcut is dominated by Upper Genesee Group shales. About mid level in the exposed section thicker siltstone stringers are intercalated with the shale. At the top of the roadcut siltstones dominate. In walking uphill along Route 414 from the town of Watkins Glen, find at the base of the roadcut a 15 cm thick siltstone with plume structures nicely developed on a joint face. Stratigraphic levels within the Genesee Group of this roadcut are referenced from the bottom of this 15 cm thick bed. 10 siltstone beds or groups of beds may be used as markers in describing the 34 m thick roadcut. Key beds are located according to Table I.

Lithological control is fundamental to the development of joints within the Appalachian Basin as is nicely illustrated by this outcrop. Vertical joints within the shales strike at 341° - 343° , whereas vertical joints within the siltstone beds strike at 331° - 334° . Although important in controlling joint development, the differences between siltstone and shale within the Ithaca Member of the Genesee Formation are subtle. Figure 5 shows histograms for the grain size distribution and composition of siltstones and shales from stop #1. At stop #1 the outcrop criterion for distinguishing a siltstone from a shale is based purely on the orientation of the joint set that a particular bed is carrying. Beds that carry "siltstone" joints have a clay/quartz ratio between 0.71 and 1.06 with more than 25% of the grains greater than 30 microns. Beds that carry "shale" joints have a clay/quartz ratio between 1.21 and 2.80 and less than 20% of its grains greater than 30 microns.

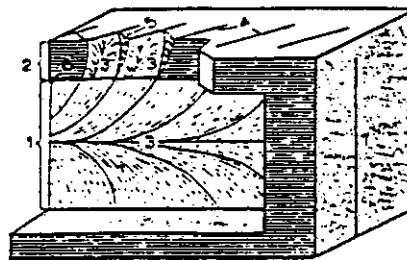
Table I

Bed Number	Stratigraphic Level (m from base)	Landmark or Road Sign
1	0 m	"Slow traffic keep right" sign
2	4 m	Natural Gas Pipe Line
3	12 m	"JCT 14" sign
4	16 m	"Village Speed Limit 30 MPH"
6	22 m	"55 MPH" sign
7	25 m	top portion of thick shale
8	27 m	caution sign for deer

Another unusual aspect is the variety of well-developed markings on the surfaces of joints in both the siltstones and shales. A composite of barbs and arrest lines leaves a delicate plumose structure on the surface of joints in siltstone (Bahat and Engelder, 1984). The barbs consist of a fine roughnesses (low relief elements) on the joint surface which were caused by local out-of-plane crack propagation. This roughness forms ridges

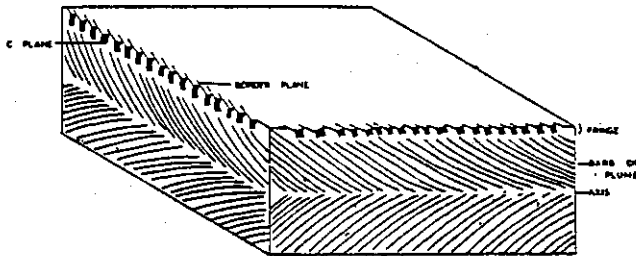


KULANDER ET AL. (1979)

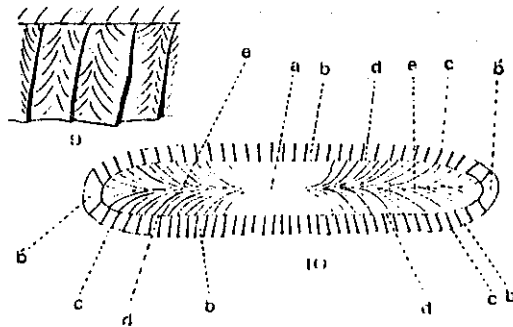


- 1 MAIN JOINT FACE
- 2 FRINGE
- 3 PLUMOSE STRUCTURE
- 4 F-JOINTS (B-PLANES)
- 5 C-FRACTURES (STRAIGHT MEDIAL)
- 6 SHOULDER
- 7 TRACE OF MAIN JOINT FACE

HODGSON (1961)



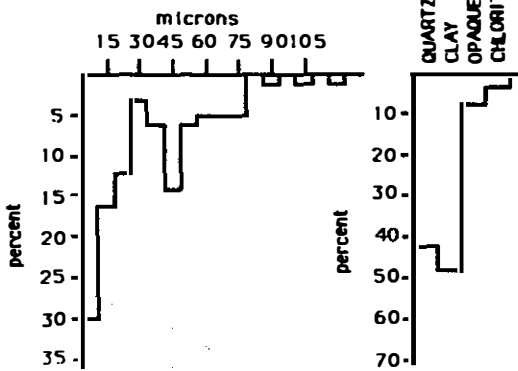
ROBERTS (1961)



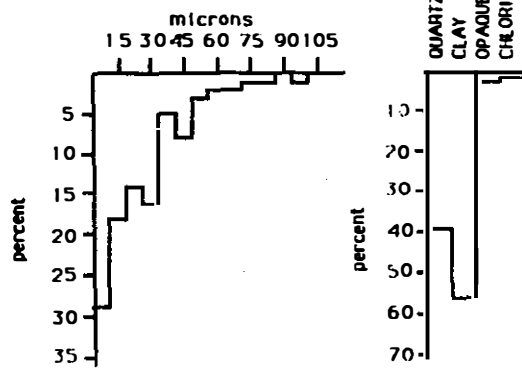
WOODWORTH (1896)

Figure 4. Plumose patterns from Woodworth (1896), Hodgson (1961), Roberts (1961), and Kulander et al. (1979). Woodworth (1896) distinguishes a fringe (b) and a plume axis (e). The fringe of Woodworth (1896), Roberts (1961), and Hodgson (1961) is the area of twist hackle steps of Kulander et al. (1979).

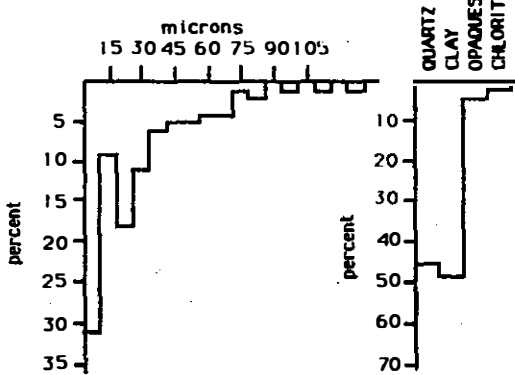
28.1 m (siltstone)



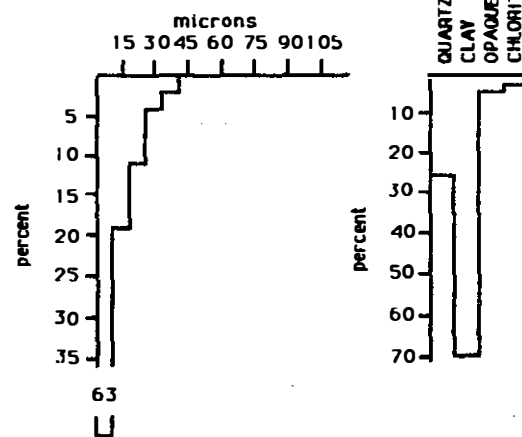
23.7 m (shale)



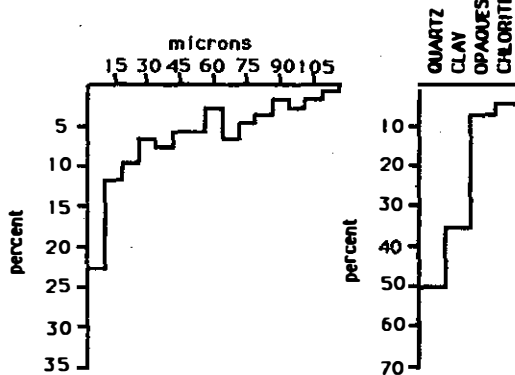
22.2 m (siltstone)



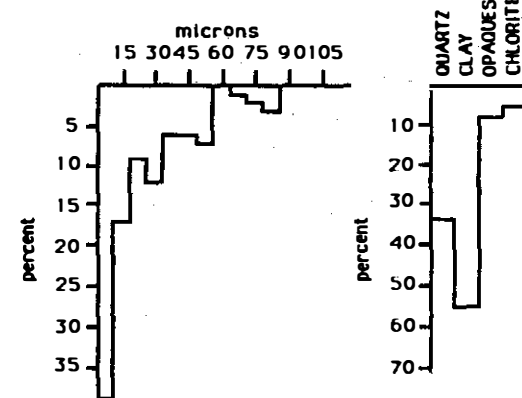
22.9 m (shale)



???. m (siltstone)



20.0 m (shale)



LACKAWANNA PHASE JOINTS

MAIN PHASE JOINTS

Figure 5. Histograms showing the grain-size distribution and composition of "shales" and "siltstones" from Stop #1. Location of each sample designated in terms of vertical distance within the outcrop. The exact location of one "siltstone" sample unknown.

parallel to the direction of rupture propagation. Out-of-plane propagation is believed to be caused by microscopic inhomogeneities, such as grain boundaries in the siltstone. Because the shale is more homogeneous on a microscopic scale there is less tendency for out-of-plane crack propagation. Hence, the shales show no surface morphology equivalent to plumose structures on the siltstones.

Joint faces within siltstones contain three varieties of plumose patterns: the straight or s-type plumose marking which is displayed in the 19 cm thick siltstone bed (#7) at the 25.7 m level (Fig. 6A); the curving or c-type plumose marking which is best displayed in siltstone (bed #8) at the 29 m level (Fig. 6B); and the rhythmic c-type plumose marking which is displayed on the 44 cm thick siltstone bed (#6) at the 22.4 m level (Fig. 6C). The straight plume has a linear axis parallel to bedding whereas the curving plume commonly has an axis which divides into several branches which in turn may themselves divide. Barbs radiate from the plume axes of both the s-type and c-type plume patterns. The barbs form a fine surface morphology which indicates the direction of rupture propagation with the rupture moving from the plume axis outward toward the edge of the joint.

The plumose structures may be traced backward to their initiation point. Cracks initiate at inclusions within the rock such as fossils, concretions, ripple marks, or microcracks. These inclusions are stress risers that permit the magnification of a far field stress to overcome the local tensile strength of the rock. At stop 1# most initiation points are bedding plane boundaries. The 19 cm thick siltstone stringer (#7) at the 25.7 m level contains four or more initiation points at the top of the bed. In contrast, the 44 cm thick stringer (bed # 6) shows initiation points on the bottom of the bed. Higher in the section (\approx 29 m level) siltstone beds are cut with initiation points within the bed.

A feature found on both shale and siltstone joints are arrest lines. These features mark the termination of propagation of individual cracks. The 4 m thick shale bed, at the 24 m level, shows a large arrest line curving on the joint face with the convex side of the line facing in the direction of joint propagation (NNW). This same shale bed contains the 19 cm thick siltstone stringer (bed #7) displaying an s-type plume pattern. Barbs of the s-type plume on the siltstone stringer diverge in the direction of propagation which is toward the NNW and compatible with the large arrest line within the shale. Within the same shale bed another joint terminates against the arrest line after propagating in the SSE direction as is again indicated by the barbs on the s-type plume within the siltstone stringer (bed #7). Arrest lines can be observed on the 44 cm thick siltstone bed (#6) at the 22.4 m level. These arrest lines are part of the rhythmic c-type plume pattern found on joint faces cutting siltstone beds. Here the arrest lines are spaced less than 1 m apart in contrast to those on the thick shales which are separated by more than 50 m.

The closely spaced arrest lines in bed #6 may be interpreted in terms of a jointing mechanism. The deeper portion of the Appalachian Basin is undercompacted (Engelder and Oertel, 1985). Such undercompaction indicates that abnormal pore pressures once prevented normal pore collapse. Pore pressures approaching the weight of the overburden are capable of initiating and driving natural hydraulic fractures. The closely-spaced arrest lines within the siltstone beds suggests a cyclic rupture. Two phases of the cycle are a slow build-up of pore pressure followed by a fast decrease accompanying the incremental propagation of a joint. This process repeats many times to leave a set of closely-spaced arrest lines.

The c-type plumes are found on joints striking at 331° - 333° , (this is considered the normal orientation for joints in siltstone layers at stop 1#) whereas small siltstone stringers (i.e. bed #7) in thick shales show the s-type plumes with joints striking at 341° . The s-type plume in bed #7 is believed to indicate a rapid rupture that extended more than 50 m in a horizontal direction. The length of the rupture is indicated by the distance between the initiation point 50 m to the SSE and the large arrest lines within the 4 m thick shale layers. In contrast, the c-type plumes give the impression of a slower less decisive rupture. Arrest lines spaced at less than a meter on the 332° joints confirm this notion.

The difference between joint propagation in the shales and joint propagation within the siltstones is further understood by placing the timing of their propagation within a regional context. On upper benches of the outcrop (at the 34 m level) deformed crinoid columnals show that the layer parallel shortening (LPS) during the

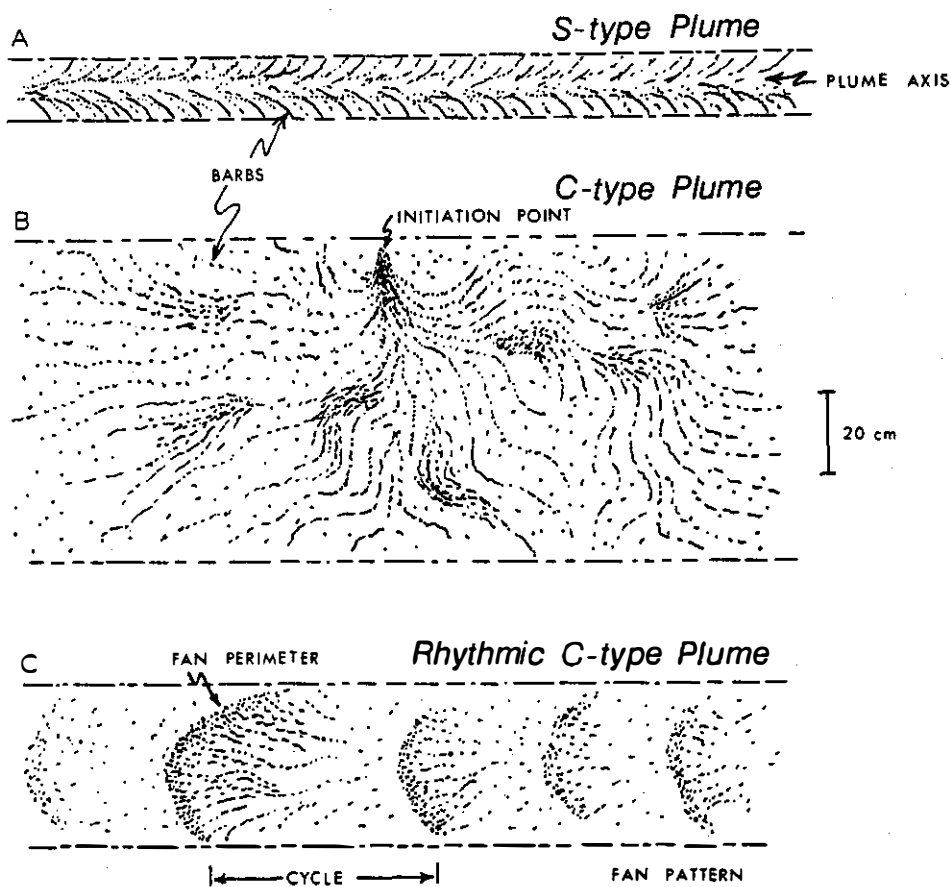


Figure 6. Various plume patterns observed on the surfaces of vertical joints in siltstones of the Finger Lakes District. A. Straight plume (bed #7). B. Curving Plume (bed #8). C. Rhythmic plume (bed #6). (after Bahat and Engelder, 1984).

Alleghanian orogeny was oriented at 341° . Geiser and Engelder (1983) interpret this LPS direction as a principal compression direction during the Main phase of the Alleghanian Orogeny. At Watkins Glen the orthogonality of shale joints and LPS indicated by deformed fossils suggests that the shale joints propagated during the Main Phase. If this is so then when did the joints within the siltstone beds propagate?

The joints within the siltstone are believed to precede those within shales. First, early joints at Ludlow - ville, New York (stop 5#) appear to be cut by Main Phase cleavage. These joints strike a few degrees counter - clockwise from the Main Phase LPS. Secondly, in a deeply buried siltstone-shale sequence the siltstones are known to show a lower least principal stress compared to shales. If joints are hydraulic fractures propagating under high fluid pressures, the joints in beds with the lower least principal stress will propagate first. Third, the preferred orientation of chlorite within the siltstone beds is compatible with a LPS counterclockwise from the LPS affecting the shales (Oertel and Engelder, 1986). Elsewhere in the central Appalachians, the Lackawanna Phase LPS precedes the Main Phase LPS with a counterclockwise compression (Fig. 2).

STOP 2: MULTIPLE JOINTING IN THE UPPER GENESEE GROUP AT WATKINS GLEN

This location just west of Watkins Glen also consists of benches in a roadcut. A 4-5 m thick siltstone unit is the major component of each bench. Here the relationship between lithology and jointing is less straight forward than was found at stop 1#.

A 3-m thick siltstone within the upper bench contains cross-fold joints (334°-336°) which propagate to but not across the contact between the siltstone and the adjacent shales. This again demonstrates how lithological changes act to stop the vertical propagation of joints. However, joints originating within the underlying shale cross into the overlying interface to propagate upward into the siltstone. This latter behavior breaks the rule that lithological changes stop vertical joint propagation. Note that joints within the siltstone are planar whereas the joints in the adjacent shale are less planar.

Both the siltstones and shales of the upper bench carry joints of more than one orientation. At the very south end of the upper bench the siltstone contains joints of two orientations. A joint striking 336° is on top of a joint striking at 343°. Close examination suggests that the joint striking at 343° propagates in a layer that looks more like a shale than the adjacent layer. There are subtle changes in lithology within units that at a distance appear homogeneous to jointing. The shale package below the 3-m thick siltstone has siltier units with joints striking at 336° versus shalier units with joints at 348°

The lower bench of this roadcut contains three cross-fold joint sets with strikes of 334°, 341°, and 004°, respectively. The three joint sets are best viewed just downhill from the north 414-east 79 sign within a face of the lower bench. All three joints have propagated in the same 4 m thick unit of siltstone beds. Again the rule of lithological selectivity of jointing is not obeyed. If it is accepted that the sets at 334° and 341° correlate with those at the previous stop, then the set at 004° requires explanation. We must also wonder if it is possible to fit three cross-fold joint sets into a regional context.

Geologists are in general agreement that tectonic joints are found as orthogonal partners to fold axes within fold thrust belts. The exact details concerning the mechanism for maintaining orthogonality are subject to debate. Engelder and Geiser (1980) took the position that, as fold axes curved around the Central Appalachian trend, members of a single joint set on the Appalachian Plateau changes orientation to maintain a position subnormal to local fold axes. A single joint set may then be traced along strike of the Plateau fold belt for more than 100 km. Nickelsen and Hough (1967) took the opposite view which was that one joint set did not change orientation along strike. Rather individual joint sets maintained a constant orientation within a smaller region and a local joint set of another orientation propagated to maintain "orthogonality" to the local folds of a different orientation.

100 km east of the Watkins Glen area the 004° orientation is far more common than at Watkins Glen. By Engelder and Geiser's (1980) interpretation the 004° joints in this outcrop are stray cross-fold joints with no tectonic significance. By Nickelsen and Hough's (1967) interpretation the 004° joints are the western most manifestation of the common cross-fold joint set found in the vicinity of Binghamton, New York. The distance between outcrops is too large to resolve this issue.

STOP 3: LITHOLOGICAL CONTROL OF MULTIPLE JOINT SETS IN ITHACA SHALE OF THE GENESEE GROUP AT TAUGHANNOCK STATE PARK

This outcrop contains four of the cross-fold joint sets which may be observed in the Finger Lakes district. Despite the complicated pattern of jointing the rule for the silt-shale jointing holds up with joints in the siltstone striking counterclockwise from joints within the shales. Step down into the stream bed just north of the bridge across Taughannock Creek. At this point several benches have been cut into the northeast back of the creek. Joints on those benches fall in three sets: 339°, 345°, and 352°. A fine example of a siltstone joint (340°) over a set of shale joints (352°) is seen just north of this point. Further upstream a joint strikes at 301°.

Here the question is which joint sets correlate with those seen at stops 1# and 2#? Does a 333° joint in siltstone at stop 1# correlate with a 340° joint in siltstone at stop 3#? Likewise, does a 342° joint in shale at stop 1# correlate with, say, a 352° joint in shale at stop 3#? Or should a correlation be made strictly on common orientation such as the 342° at stop 1# with the 340° at stop 3#? These are questions that will probably never be answered to everyone's satisfaction.

Strike joints (080°-090°) are irregular in plan view and widely-spaced in this outcrop. This irregular form is common for strike joints throughout the Appalachian Plateau. Strike joints are believed to propagate upon uplift and, hence, be very late in origin.

STOP 4: DISJUNCTIVE CLEAVAGE AND JOINTING IN THE TULLY LIMESTONE AND GENESEO SHALE AT TAUGHANNOCK FALLS STATE PARK

Taughannock State Park features a U-shaped hanging valley and 50 m waterfall at the head of a 1.5 km-long gorge cut to the level of Cayuga Lake. Outcrops within the park consist of the Tully Limestone and Genesee shales in the stream bed of Taughannock Creek and the lower portion of the Genesee Group (the Genesee shales) exposed on the walls of the gorge. 200 m upstream from the park entrance bedding surfaces of the Tully Limestone may be examined in the stream bed. In another 800 m upstream the stream bed becomes the Genesee shale.

On beds of the Tully Limestone, a disjunctive cleavage is well developed. The cleavage gives a faint herringbone pattern on the gray pavement of the Tully Limestone. Cleavage domains appear as a wavy trace of a dark selvage against the light gray background of Tully Limestone. Individual selvages extend for 10s of cm before ending in many fine branches. The microlithons of Tully Limestone are 5 to 15 cm thick. The spacing of cleavage domains constitutes a weak cleavage according to the classification of Alvarez and others (1978). A general trend for the cleavage of 077° is normal to the compression direction of the Main Phase of the Alleghanian Orogeny in the vicinity of Ithaca, New York. Because the cleavage is wavy any one cm length of selvage might be misoriented from the 077° trend by as much as 15°. Close examination of the selvages will reveal short stylolites pointing in the direction of the Main Phase compression at about 347°. At stop 1# the compression direction was 341°.

The contact between the Tully Limestone and the Genesee shales gives a fine example of the relationship between disjunctive cleavage in the limestone and the development of pencil cleavage in the shales. Best examples are found on the north side of the creek about 400 m from the park entrance. The long axes of the pencils within the Genesee Shales trend at 077° which parallels the strike of disjunctive cleavage within the Tully Limestone. Here the pencils are blocky rectangular solids rather than the long and skinny shape as found in other outcrops of the Appalachian Plateau. The two short dimensions of a pencil cleavage consist of bedding and a disjunctive cleavage normal to bedding.

At stop 4# the Tully Limestone contains none of the cross-fold joints found within the Genesee Group at stops 1#, 2#, and 3#. Limestone affects joint development in a different manner than siltstone and shale. The best developed joints in the Tully Limestone are several sets of en echelon cracks found on the second bench of Tully Limestone about 300 m from the park entrance. Individual cracks within the en echelon set strike at 316° whereas the shear couple indicated by the en echelon zone strikes at 324°. There seems to be no clear relationship between these en echelon cracks and the Alleghanian Orogeny.

On walking upstream Taughannock Creek makes a righthand turn at the point where cross-fold joints appear within the Genesee shales. Here it is common to see later subparallel cross-fold joints (~330°) curving into and abutting earlier cross-fold joints (~340°). This abutting of cross-fold sets is not found at other stops and is difficult to explain in the context of a two phase Alleghanian Orogeny.

The rocks in the walls of Taughannock Creek gorge present an example of the behavior of cross-fold joints within thick (> 50 m) sequences of homogeneous shale. From the Taughannock Creek the tectonic joints can be traced continuously up the valley wall for a large fraction of the exposure of the Genesee shales. The joints

propagate so that their vertical dimension is as large or larger than their horizontal (parallel to strike) dimension. In contrast with stops 1#, 2#, and 3# where vertical joint growth was arrested by bedding interfaces, vertical joint growth was not impeded in the Genesee Shale. This is best seen at the point where the gorge makes a right turn 1000 m from the park entrance. After rounding the right turn in the creek bed, look up on the southeast side of the gorge. Here cross-fold joints are displayed on the southern wall whereas strike joints are displayed on the eastern wall. Across the creek from this point cross-fold joints can be seen on the northern wall. The cross-fold joints are better developed and more closely spaced. This same effect can be seen at the falls where (facing the falls) rocks to the left of the falls show cross-fold joints whereas those to the right of the falls show widely spaced strike joints. Note that cross fold joints become irregular within 50 m above the top of the Tully Limestone.

Engelder (1985) distinguishes tectonic joints (those caused by abnormal pore pressures during tectonic compression) from release joints (those caused by erosion and controlled in orientation by an pervasive fabric such as disjunctive cleavage). At the right turn in Taughannock Creek the cross-fold joints are tectonic joints and the strike joints are release joints. Evidence for the development of abnormal pressures during tectonic (cross-fold) joint propagation include the undercompaction of the Genesee shales (Engelder and Oertel, 1985). Note that good examples of release joints are common in the creek bed upstream from this point. In general the release (strike) joints tend to be less regular in profile than tectonic (cross-fold) joints.

STOP 5: THE RELATIONSHIP BETWEEN DISJUNCTIVE CLEAVAGE AND JOINTING IN TULLY LIMESTONE AND THE UPPER HAMILTON GROUP IN SALMON CREEK AT LUDLOWVILLE, NEW YORK

Rocks within this outcrop include the Tully Limestone (in the bed of Salmon Creek) above the Upper Member of the Hamilton group (seen below the falls). The major mesoscopic structure within the Tully Limestone is a spaced solution cleavage indicating a compression direction of about 005°. Within just about 4 km between stop 4# and stop 5# the compression direction as indicated by cleavage has changed by more than 20°. This is one of the most abrupt changes in LPS within the entire New York Plateau. Less than 2 km south of this outcrop the Tully Limestone is cut by several faults as seen in the Portland Point Quarry. Here local structures overprint a relatively homogeneous strain pattern.

Calcite filled veins striking between 340° and 345° dominate within the Tully Limestone of this outcrop. These veins seem to be cut by the spaced cleavage indicating that the veins propagated first. This cross-cutting relationship is taken as evidence that 345° joints are Lackawanna in age whereas the cleavage correlates with the Main Phase deformation.

Overcoring tests at this outcrop show a residual maximum compressive stress normal to the trend of the cleavage (i.e. 005°) (Engelder and Geiser, 1984) (Fig. 7). This residual stress is believed to have been locked into the Tully Limestone during the Main Phase of the Alleghanian Orogeny.

In the pavement of the stream below the falls the Hamilton Group carries joints striking between 70° and 75°. On the eastern wall of the falls just under the Tully Limestone 340°-345° joints are well developed in the Hamilton Group.

On the far side of the stream a few trilobites may be seen weathering out of the polished surface of the stream bed.

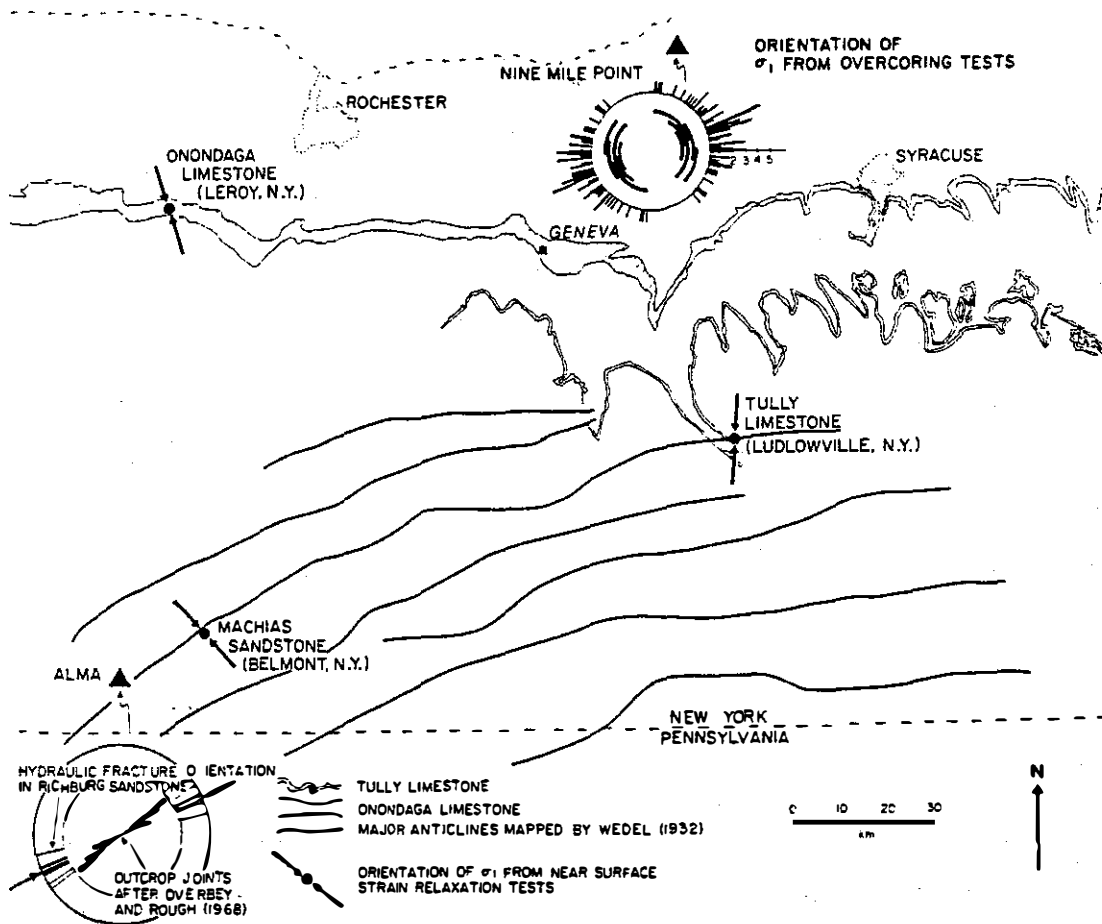


Figure 7. The location of strain relaxation experiments in the Tully Limestone, Machias Sandstone, and Onondaga Limestone. Overcoring data from Nine Mile Point represent the orientation of maximum horizontal compressive stress for 74 measurements (Dames and Moore, 1978). The variation in azimuth from each of six test holes shown inside the rose diagram. The orientations of nine hydraulic fractures from Alma are shown, with the average azimuth located by the inward pointing arrows (Overby and Rough, 1968)

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ROAD LOG FOR THE USE OF JOINT PATTERNS IN UNDERSTANDING THE ALLEGHANIAN OROGENY

CUMULATIVE MILEAGE	MILES FROM LAST POINT	ROUTE DESCRIPTION
0.0	0.0	STOP 1# ---- Travel south on Franklin Street in the town of Watkins Glen, New York and find the intersection of New York State routes 14 and 414 on the south side of town. At that intersection park in the parking lot of the Yankee Pedlar. From the parking lot the outcrop is viewed by walking uphill along route 414.

- Leave stop 1# by driving north along Franklin Street which is both routes 14 and 414. (Montour Falls Quadrangle)
- | | | |
|------|-----|---|
| 0.7 | 0.7 | Turn right to follow route 414 at 4th Street. |
| 2.3 | 1.6 | STOP 2# ---- Intersection of New York State routes 79 and 414. Park downhill from the North 414-East 79 sign. From this intersection stop 2# is viewed by walking downhill and south along route 414. Leave stop 2# by driving east along route 79. |
| 3.7 | 1.4 | Burdett |
| 8.8 | 5.1 | At intersection of New York State routes 79 and 227 bear left along 227 towards Reynoldsville and Perry City. |
| 14.8 | 6.0 | Perry City. Bear left on route 227 to Trumansburg. |
| 18.9 | 4.1 | Trumansburg at the intersection of routes 227 and 96 turn right or south along route 96. |
| 20.0 | 1.1 | At Cemetery Road across from the Trumansburg High School turn left and then take an immediate right to follow Falls Road south. |
| 22.0 | 2.0 | Left on Taughannock Road |
| 22.1 | 0.1 | STOP 3# ---- Park at the Falls Road Bridge which crosses Taughannock Creek. The outcrop is just upstream from the bridge. Leave the outcrop by driving south. |
| 22.4 | 0.3 | Intersection of Jacksonville Road and Gorge Road. Turn left and drive east or downhill on Gorge Road. |
| 24.1 | 1.7 | STOP 4# ---- At the intersection of Gorge Road and Route 89 turn left on 89 and proceed north about 200 m to the parking lot of Taughannock Falls State Park. The outcrop can be viewed by walking east or upstream along Taughannock Creek. Leave stop 4# by driving south along route 89. (Ludlowville Quadrangle). |
| 33.5 | 9.4 | Ithaca at the intersection of New York State routes 89, 96, and 79. Turn left and follow routes 96 and 79 into town. |
| 34.0 | 0.5 | Ithaca at the intersection of New York State routes 13, 34, 79, and 96. Turn left and follow route 13 and 34 north. |
| 35.8 | 1.8 | Exit four-lane highway to following route 34 north towards Auburn. |
| 41.7 | 5.9 | South Lansing at the intersection of New York State routes 34 and 34B. Turn left and follow 34 B to the north. |
| 42.7 | 1.0 | Turn right and follow Brickyard Road to the north. |
| 43.5 | 0.8 | Turn right and follow Main Street into Ludlowville. |
| 43.9 | 0.4 | STOP 5# ---- Enter park at the north end of Main street in Ludlowville. Walk northeast 50 m to the outcrops in the bed of Salmon Creek. |



FACIES AND FOSSILS OF THE ONONDAGA LIMESTONE IN CENTRAL
NEW YORK

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INTRODUCTION

Four members of the Onondaga limestone are recognized in the central part of New York State: Edgecliff, Nedrow, Moorehouse, and Seneca. Detailed descriptions of these members may be found in Oliver (1954, 1956) and Feldman (1985). Paleocommunity analyses of the non-reefal aspects of the Onondaga Limestone have been made by Lindemann (1980) and Feldman (1980) while Lindemann and Feldman (1981) have summarized the stratigraphy, lithofacies and paleocommunities of the Onondaga in the Syracuse area. We believe that the formation in this area has been studied thoroughly, consequently, the purpose of this trip is to sample the fauna in order to: [1] Become familiar with the fossils that are characteristic of the Onondaga, and [2] Attempt to correlate the various taxa with their paleoenvironments. This trip, therefore, will concentrate on collecting the fossils of the Onondaga Limestone in the Syracuse area.

The most abundant invertebrate fossils of the Onondaga which lend themselves to collecting in this extremely dense limestone are the brachiopods and corals. Brachiopod collecting is usually best accomplished in the shaly Nedrow Member, at the base of shale lenses, followed by well weathered Moorehouse and Seneca strata. Corals are most abundant in the Edgecliff Member. Since the Onondaga weathers very slowly it is best to locate bedding planes which have not been touched for several years, if possible.

EARLY HISTORY

Students of New York Geology probably require little introduction to the Onondaga Limestone. Firmly established in the geologic literature as the State's lowermost Middle Devonian formation, the Onondaga is one of its most prominent units due, in part, to the fact that it is laterally continuous from the central Hudson River Valley north to the Helderbergs and west to the Niagara River and the Ontario Peninsula. Between Buffalo and the Helderbergs the formation's resistance to erosion, slight southerly dip, and position in section place it atop an escarpment defining the

southern margins of the Ontario lowlands and the Mohawk Valley; a situation which provided early explorers and geologists with a wealth of natural exposures.

Following on the heels of the American Revolution, the westward expanding settlement of New York prompted exploration for transportation routes, mineral resources, and building materials. In the bad old days long before central heating, finished lumber, and fiberglass insulation, quarried-stone and kiln-produced mortar and plaster were, where available, valuable building materials. Historic records document the former workings of hundreds of quarries and kilns by which the Onondaga supplied construction industries of the early 1800's. This gives the formation a role in human history and, as 1986 marks the sesquicentennial of the New York Geological Survey, it seems appropriate to note some of the highlights in the history of geologic activities leading to our current understanding of the Onondaga Limestone.

Geologic observations made during the pre-Survey years are summarized by Wells (1963). Even a casual reading of Wells' book leaves little doubt that fossils figured significantly among descriptions recorded during the early 1800's. In 1803 Compte de Volney collected fossil specimens from the strata now referred to as the Onondaga and sent them to Lamarck in France. The first published illustration of a fossil from the New York Devonian, which appeared in 1807, is that of a "sheep horn" collected from the Onondaga. Wells (1963, p. 14), reaching a slightly different conclusion, considered the specimen to be the gyroconic nautiloid Goldringia cyclops. The year 1810 found future New York governor De Witt Clinton in the field checking on potential routes for a proposed canal linking Lake Erie and the Hudson River. Noting details and fossils of the limestone escarpments which traverse the State, Clinton made some of the earliest interpretations of Devonian water depths (p. 18). Furthermore, Wells (1963, p. 38) reported that in 1822 Clinton anonymously published a series of geologic observations among which can be found the statement "...a great limestone ridge runs throught the whole of this country, east to west...north of it a ledge of gypsum commences...." Obviously this is not the first observation of the Onondaga Formation, but it may be the earliest to demonstrate knowledge of a discrete limestone body extending fully across the State. During this time Amos Eaton, Senior Professor of the Rensselaer School (RPI), was hard at work trying to document and make sense of New York's geologic column. The conceptual necessity to reproduce the strata of Europe severely hampered Eaton's work. However, in 1824 (Wells, 1963, p. 42) he recognized a limestone unit with abundant "hornstone" (chert) which he named the Cornitiferous limerock. Four years later, Eaton (1828, p. 153) recognized "lower or compact" and "upper or shelly" divisions within the unit. This then was the status of what would later become the Onondaga Limestone when, in 1836, Governor William Marcy established the Survey of New-York. Within the next decade

work of the Survey's principal geologists Emmonds, Hall, Mather, and Vanuxem would alter the course of geologic investigations in North America. Eaton did not live to see his work completed; he died on Tuesday, May 10, 1842.

The completed reports of the original four district surveys brought the Onondaga Limestone into a fairly modern aspect. Since Emmond's Second District lies beyond the formation's limits and Mather contributed little original information in his First District report, we will concentrate on the reports of Vanuxem (1842) and Hall (1843). In examining the writings of these two pioneer geologists we should bear in mind that Vanuxem considered the term "formation" to be ambiguous and eschewed its use while Hall, holding to the principle that similar products result from similar processes, saw no such ambiguity and promoted the term. This dichotomy is clearly shown by the ways in which the two geologists treated similar observations. Relative to the Onondaga Limestone, Vanuxem was a "splitter" and Hall a "lumper."

Following Hall's (1841) nomenclatural lead, Vanuxem (1842, p. 132) recognized the then Onondaga Limestone and described its "light grey color, crystalline structure, toughness, and its organic remains which are very numerous." He noted that the Onondaga contained numerous "smooth encrinal stems" exceeding a half inch in diameter and that the unit maintained a thickness of 10-14'. Succeeding the Onondaga, Vanuxem (1842, p. 139) recognized, as a discrete unit, the cherty Corniferous Limestone, measuring 60-80' in thickness at Cherry Valley. Above the Corniferous, Vanuxem (1839, p. 275) identified a set of chert-free limestone beds with abundant Strophomena ("Chonetes") lineata which he referred to as the Seneca Limestone. It is evident from his annual report (1839) and final report (1842) that, while in print he treated it as a subdivision of the Corniferous, Vanuxem considered the Seneca to be a separate unit. He only refrained from completing the divorce in the final report due to uncertainties resulting from the absence of a Seneca lithology and fauna in the Helderberg area.

Working in the Fourth Geological District, which covers much of New York's western half, Hall (1843) expanded on the detail of Vanuxem's report. Beneath the Onondaga Limestone Hall (1843, p. 151) reported on "a few inches of sandstone" which he referred to the Schoharie Grit. He (1843, p. 152) described the Onondaga as 1-40' of limestone with "light grey color often approaching to white, more or less crystalline in structure, and containing numerous fossils." Noting extreme variations in chert abundance as well as alternating chert-rich and chert-free strata, Hall included the "cherty mass" with the Onondaga rather than with the Corniferous Limestone above. Attending to details, Hall described the now famous Onondaga reefs, complete with various facies, and devoted considerable attention to interpreting depositional environments. Working at a time when many people, geologists included, still believed the earth to be very young, he (p. 156) noted that "the simple fact of the successive growth of

coral upon deposits covering other corals, of itself proves a great lapse of time; for the growth of all these forms is exceedingly slow."

Above the Onondaga, Hall found that the Corniferous Limestone thickened westward from less than 30' in Seneca County to in excess of 71' at Leroy, and that its "hornstone content" showed considerable variation between localities. Recognizing, as had Vanuxem, that the upper Corniferous strata lacked chert, it is interesting to note that, in a quarry near Waterloo, Hall (1843, p. 163) found a "separation of the higher and lower strata by a 'wayboard,' or a seam of (yellowish) clay about four inches thick." This is an early description of the Tioga Bentonite which, in modern stratigraphic terminology separates the Moorehouse and Seneca members of the Onondaga Formation. As we have seen, the Onondaga of today was not recognized at the time that Hall wrote the report of the Fourth District. However, noting vertical and lateral lithologic variations within both the Onondaga and Corniferous (including Seneca) limestones which approximate the differences between them, Hall (several pages) was adamant that they be united as "one formation" (p. 152) with divisions based on fossil content. This brought the earlier observations of De Witt Clinton to a full circle in establishing a single unit and set the stage for today's concept of the Onondaga Limestone.

LITHOSTRATIGRAPHY AND BIOSTRATIGRAPHY

In time, the formational name Onondaga Limestone came to encompass the 1836 Geological Survey's Onondaga, Corniferous and Seneca limestones. Based on observations made in Hall and Vanuxem's type areas of Onondaga and Seneca counties, Oliver, (1954) formally divided the Onondaga into four members. In vertical succession these are the Edgecliff (formerly Onondaga), Nedrow, Moorehouse (formerly Corniferous), and Seneca limestones.

The Edgecliff is a thick bedded to massive, light gray to pink, poorly washed to unsorted biosparite. The basal bed(s) of the member contains quartz sand which, in some places, is sufficiently abundant to constitute a quartz arenite. The Edgecliff fauna abounds with rugose corals, tabulate corals and crinoids. At Split Rock, the type locality, the Edgecliff is 8' (2.3 m) thick. Its thickness is highly variable to the west and generally increases eastward to about 48' (15 m) in the Mid-Hudson Valley.

Succeeding the Edgecliff in central and eastern New York, the Nedrow Member is a 10-14' (3-4.3 m) sequence of thinly bedded, dark gray, argillaceous, fossiliferous to sparse biocalcissiltites. While brachiopods dominate the megafauna and the unit is typified by platyceratid gastropods and the small rugosan Amplexiphyllum hamiltoniae, thin sections reveal that in many places the Nedrow is numerically and volumetrically dominated by the microfossil Styliolina fissurella. The Nedrow undergoes an eastward facies change

and thickens to approximately 43' (13 m) near Catskill. In western New York it overlies the Clarence Member and is of uncertain thickness.

Gradational with and succeeding the Nedrow, the Moorehouse Member is a 24' (7.3 m) unit of laminated to thick bedded, dark gray, argillaceous, cherty, sparse biocalcissiltite. Shale partings typically separate the limestone strata. The fauna is dominated by a diversity of brachiopods with lesser numbers of trilobites and gastropods. Both east and west of its central New York type section at Jamesville, the Moorehouse thickens and undergoes a facies change to cleaner and coarser grained textures with crinoid-coral faunas.

The Moorehouse is overlain by 4-6' (10-15 cm) of yellowish clay, the Tioga Bentonite, which is the lowermost bed of the Seneca Member. A 26' (7.8 m) unit of dark-gray, fossiliferous and sparse biocalcissiltites, the Seneca becomes increasingly argillaceous upward. In central New York the fauna is dominated by the diminutive strophomenid brachiopod "Chonetes" lineata which is found in a zone approximately 10' (3 m) above the Tioga. The Seneca is overlain by black shales of the Marcellus Formation.

The base of the Onondaga Limestone has long been considered to coincide with the base of the Middle Devonian in New York State (Rickard, 1975). The biozones of conventional megafossils such as brachiopods, corals and cephalopods are summarized by several authors in Oliver and Klapper (1981); all support this age assignment relative to other North American faunas. Unfortunately, these megafossils are restricted to North America and direct correlation with European biozones is not possible. This difficulty in correlation was recently compounded when the International Union of Geological Sciences ratified the decision of the Subcommittee on Devonian Stratigraphy that the base of the Middle Devonian Series and of the Eifelian Stage coincide with the first occurrence of the conodont Polygnathus costatus partitus (Ziegler and Klapper, 1985). The subspecies partitus is the second in a lineage of three which are sequentially related. Thus, the bottom of the patulus Zone is high in the Emsian Stage and the bottom of the costatus Zone is well within the Eifelian. Klapper (in Oliver and Klapper, 1981) reported that the upper Nedrow beds at Cherry Valley yield both Polygnathus costatus costatus and P. c. patulus, placing the member's top well within the partitus Zone. Noting this along with the fact that P. c. partitus is unknown from the Onondaga, Ziegler and Klapper (1985) suggested, with question marks, that the Edgecliff Member is within the patulus Zone and correlative to the Emsian Stage of the Lower Devonian Series. At this time, the absence of definitive taxa from the Edgecliff, Clarence, and lower Nedrow members leaves the precise age of the lower Onondaga in a very gray (N 3?) area.

ONONDAGA LITHOFACIES

Lindemann (1980) identified six carbonate lithofacies of the Onondaga Formation on the basis of relative abundances of calcisiltite, bioclasts, cement, argillaceous mud and pyrite as point-counted in thin section. Four of these lithofacies are well represented in central New York and figure in this discussion. Mean abundances of their constituents are shown in Table 1 and descriptions follow. As the lithofacies do not exactly correspond to formally named carbonate lithologies, they are referred to by Roman numerals.

Lithofacies VI. Lithofacies VI consists of thick bedded to massive, light gray, poorly washed to sorted biosparites. Varying abundances of quartz sand are present in samples from the lower beds of the Edgecliff Member. Comminuted crinoids and bryozoans dominate the fossils seen in thin sectioned samples from central New York. While evidence of bioturbation is rarely observed, vertical burrows are common as are shell lag concentrates and cross-laminae. Rare cryptalgal laminae, oncolites and the calcareous alga Asphaltina are present. This lithofacies is characteristic of the Edgecliff throughout the state and also occurs higher in the formation in eastern and western New York.

Lithofacies VI is interpreted as having been deposited under shallow shelf conditions in wave-agitated waters of very low turbidity.

Lithofacies II. Lithofacies II consists of medium bedded to massive, medium gray packed biocalcisiltites. Crinoids and bryozoans dominate the fossils in thin section and in the field. Intimately associated with Lithofacies VI, and occurring in the Edgecliff throughout the state and the upper Onondaga to the east and west, Lithofacies II differs primarily in its paucity of interparticulate cement. There are additional faunal differences, but these do not figure in lithologic descriptions.

Lithofacies II is interpreted as having been deposited under nonturbid carbonate shelf conditions quieter than, but similar to, those of Lithofacies VI. Stratigraphic distribution and association with other moderate energy lithofacies indicate that II was deposited just offshore from, or in slightly deeper water than VI. Lagoonal conditions are not indicated in central New York.

Lithofacies III. Lithofacies III consists of laminated to medium bedded, dark gray, argillaceous calcisiltite, fossiliferous calcisiltite and sparse biocalcisiltite. The pyrite content of this lithofacies does not exceed 1 percent in central New York. Comminuted crinoids and trilobites dominate the megafossils in thin section and the microfossil Styliolina fissurella reaches its maximum abundance. Fossils are occasionally concentrated in thin stringers associated with argillaceous laminae. However, most fragments were scattered by intense bioturbation. This lithofacies predominates in the Nedrow of central New York and in the Moorehouse elsewhere in the state.

Lithofacies III is interpreted as having been deposited in quiet, moderately turbid water offshore from Lithofacies VI and II. Restricted circulation and low oxygen levels are not indicated. The sediment's fine grained nature suggests a flocculent or soupy sediment-water interface, a condition not particularly conducive to colonization by the larvae of sessile organisms. This accounts for the relative abundance of calcisiltites and planktonic styliolines.

Lithofacies V. Lithofacies V consists of laminated to medium bedded, dark gray, highly argillaceous fossiliferous calcisiltites and sparse biocalcisiltites. Trilobites dominate the fossils in thin section and share dominance with brachiopods in field observations. Chondrites and general bioturbation are abundant. This facies is virtually restricted to the Moorehouse and Seneca members of central New York where it is intimately associated with Lithofacies III. It differs from III in containing about twice as much argillaceous mud and slightly more pyrite (Table 1).

Lithofacies V is interpreted as having been deposited in quiet, relatively deep and turbid water in and near the subsiding axis of the Appalachian Basin. Because this facies occurs in the area of the formation's lowest rate of sedimentation, probably less than half that of some sites to the east for example, the magnitude of real day-to-day turbidity required to attain its approximately 20 percent argillaceous content is uncertain. While fluctuations in argillaceous influx are evident as shale laminae, it appears that the depositional conditions of lithofacies V differ from those of III primarily in geographic proximity to the relatively carbonate-starved and more restricted axis of the Appalachian Basin.

Table 1. Mean percent abundances of lithofacies' constituents in the Onondaga Limestone, central New York.

Facies	Calcisiltite	Bioclasts	Cement	Detrital Mud	Pyrite
VI	15	61	21	4	1
II	38	53	3	4	1
III	84	2	0	10	2
V	67	10	0	21	2+

DEPOSITIONAL HISTORY

The succession of Onondaga members is closely associated with specific lithofacies, as represented by outcrops in the general Syracuse vicinity. The interpretation below pertains only to this area and does not apply elsewhere.

Onondaga deposition began in clean, shallow, wave-agitated waters on a relatively flat erosional surface during a generally westward transgression of the sea. Quartz sand eroded from Early Devonian units was rounded, sorted, and redeposited as a quartz arenite at the formation's base. As the shoreline progressed westward, the shallow carbonate shelf conditions of Lithofacies VI were established and the supply of quartz sand diminished, eventually ceasing all together. Northward (today) migration of the Appalachian Basin into the area beginning in late Edgecliff time produced a gentle down-warp in the sea floor in which progressively increasing water depths are indicated by the succession of lithofacies VI, II and III. Flocculent bottom conditions reduced the sessile benthonic fauna, thus minimizing carbonate production and paving the way for the lithofacies III, V succession. In the face of argillaceous influx coupled with increasingly deleterious benthonic conditions, shale deposition gradually became dominant over carbonate deposition until the latter virtually ceased. The Marcellus Shale gradually prograded westward over the Onondaga Limestone.

MEGAFOSSILS OF THE ONONDAGA LIMESTONE IN CENTRAL NEW YORK

The predominant megafossils of the Onondaga Limestone in the Syracuse area that will be collected on this trip are brachiopods (Table 2, figs. 1-4) and corals (Table 3). Therefore, these groups will be treated in more detail than other, less common fossils, such as gastropods, trilobites, cephalopods, crinoids, bryozoans and sponges (Table 4). Some common corals and a lithistid sponge, illustrated in Shimer and Shrock (1944), are shown in figures 5 and 6.

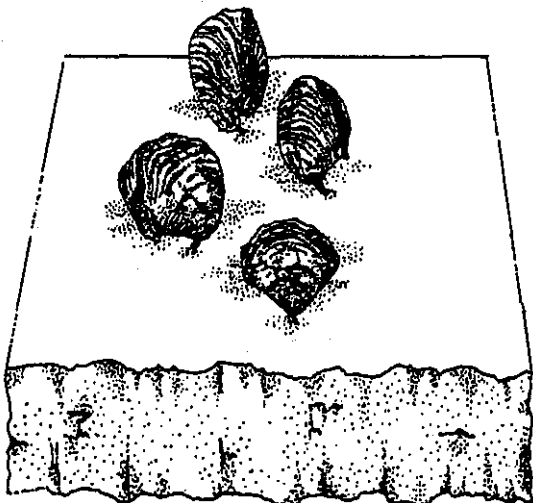


Figure 1. *Athyris* sp. A in plenipedunculate life position on the sea floor. This life strategy refers to those brachiopods in which the pedicle is a single, unbranched muscular structure apart from its distal tip.

Table 2. Brachiopods of the Onondaga Limestone in central New York from Cherry Valley to Syracuse.

Taxon	Common	Rare	Very Rare
<u>Acrospirifer duodenaria</u>			X
<u>Athyris</u> sp. A		X	
<u>Atrypa "reticularis"</u>	X		
<u>Amphigenia</u> sp.			X
<u>Atlanticocoelia acutiplicata</u>		X	
" <u>Chonetes</u> " aff. <u>lineata</u>	X		
<u>Charionoides</u> aff. <u>doris</u>			X
<u>Costistrophonella</u> cf. <u>punctulifera</u>			X
<u>Coelospira camilla</u>		X	
<u>Dalejina</u> aff. <u>alsa</u>		X	
<u>Gypidula</u> sp.			X
<u>Leptaena</u> aff. " <u>rhomboidalis</u> "	X		
<u>Levenea</u> aff. <u>subcarinata</u>		X	
<u>Meristina</u> cf. <u>nasuta</u>			X
<u>Megakozlowskiella raricosta</u>	X		
<u>Megastrophia</u> sp.		X	
" <u>Mucrospirifer</u> " cf. <u>macra</u>			X
<u>Nucleospira</u> aff. <u>ventricosa</u>			X
<u>Orthotetacid</u> indet.	X		
<u>Pentagonia unisulcata</u>			X
<u>Penatmerella arata</u>		X	
<u>Schizophoria</u> cf. <u>multistriata</u>			X
<u>Strophodonta</u> cf. <u>demissa</u>			X
<u>Stropheodontid</u> indet.			X
<u>Trematospira</u> sp.			X

Note: Although this table denotes relative abundance of brachiopod taxa in the central part of the state in terms of common, rare and very rare, it should be noted that some species are more abundant in specific horizons or beds and are relatively rare throughout the remainder of the formation. For example, "Chonetes" aff. lineata occurs abundantly ten feet above the Tioga Bentonite but rarely in the rest of the Onondaga and Amphigenia sp. occurs only in the basal Edgecliff Member where it is locally abundant.

Table 3. Corals of the Onondaga Limestone in central New York from Cherry Valley to Syracuse.

Taxa	Common	Rare
<u>Tabulates</u>		
<u>Aulopora</u>	X	
<u>Syringopora</u>	X	
<u>Favosites</u>	X	
<u>Lecfedites</u>	X	
<u>Rugosans</u>		
<u>Amplexiphyllum</u>	X	
<u>"Heterophrentis"</u>	X	
<u>Acinophyllum</u>	X	
<u>Breviphrentis</u>		X
cf. <u>Syringaxon</u>		X
<u>Cystiphyllodes</u>	X	
<u>Siphonophrentis</u>	X	
<u>Heliophyllum</u>	X	

Brachiopods. Since most of the brachiopods that will be collected loose consist of articulated specimens, the descriptions below do not include interior morphologies. For more detailed descriptions, including internal features, refer to the Treatise or Feldman (1985). Collecting brachiopods from the Onondaga Formation in central New York is best accomplished by carefully searching weathered outcrops in the Nedrow Member, especially reentrants, which form due to the more rapid weathering of the shale as compared to the dense limestone. Collecting in the other members varies, but is generally difficult. As a rule, the longer the outcrop has weathered the better the chances of finding suitable specimens. Occasionally, due to local diagenetic factors, it is possible to locate silicified material ranging in quality from poor to fair. As one approaches the western and especially eastern areas of the outcrop belt silicification tends to increase in amount and quality.

Acrospirifer duodenaria - Biconvex shells transversely subelliptical in outline; hinge line long and straight; medial open delthyrium with no preserved deltidial plates; pedicle valve bears narrow, triangular, moderately deep, noncostate sulcus; brachial valve bears corresponding fold; five to six rounded plications on each pedicle flank with U-shaped interspaces; anterior commissure uniplicate.

Athyris sp. A - Shells transversely suboval in outline, and subequally biconvex with pedicle valve slightly deeper than brachial valve; ventral beak suberect terminating in, small round foramen; brachial beak

smaller and less noticeable; pedicle valve bears shallow sulcus with corresponding low fold on brachial valve; anterior commissure weakly uniplicate; some forms nonsulcate and rectimarginate; fine, concentric growth lines on both valves.

Atrypa "reticularis" - Dorsibiconvex shells with well rounded radial costellae which increase in size and number anteriorly; costellae separated by U-shaped interspaces; concentric growth lamellae cross costellae becoming more distinct and frilly anteriorly; anterior commissure rectimarginate or slightly deflected towards brachial valve.

Amphigenia sp. - Large shells, costellate, elongate in outline; anterior commissure rectimarginate to broadly sulcate; delthyrium wide, triangular and unmodified by deltidial plates; round pedicle foramen present.

Atlanticocoelia acutiplicata - Subcircular in outline with length almost equal to width; brachial valve gently convex, pedicle valve slightly more so; weak pedicle sulcus sometimes noticeable on larger specimens; no corresponding dorsal fold; hinge line very short and becomes rounded anteriorly; no interareas present; anterior and lateral commissures crenulate; ten to twelve plications with U-shaped interspaces; concentric growth lines, two or three per shell, common on epehebic forms.

"Chonetes" aff. lineata - Shells small, subsemicircular in outline and concavoconvex in lateral profile; interareas very narrow; no delthyrial structures preserved; greatest width at hinge line or anterior to midlength; valves covered with fine capillae which increase anteriorly by bifurcation.

Charionoides aff. doris - Biconvex, elongate shells; fold and sulcus often poorly defined; pedicle beak slightly incurved with round foramen; delthyrium triangular; valve exteriors smooth with fine, concentric growth lines present at lateral margins of both valves.

Costistrophonella cf. punctulifera - Shells wider than long, subsemicircular in outline; hinge line long, straight; interarea evident; numerous, subangular, radiating costae which increase anteriorly by intercalation and bifurcation; interspaces broad and shallow with 11 to 12 costae per 5 mm near anterior commissure at midline; costae crossed by about 7 evenly spaced growth lines.

Coelospira camilla - Small, concavoconvex to planoconvex, subcircular to suboval in outline; small, distinct pedicle foramen on incurved pedicle beak; no interarea evident; maximum width about one-third valve length in adults; pedicle valve bears two medial plications usually at least as large as remaining radial plications on flanks; interspaces U-shaped; brachial valve bears medial plication which generally bifurcates at one-third valve length; median interspace usually flat but sometimes bears small ridge; plications broader

on flanks and thinner toward lateral commissure; several well-defined, concentric growth lines evident near anterior commissure in adult forms.

Dalejina aff. alsa - Shells ventribiconvex, transversely suboval to subcircular in outline; hinge line very short and straight in apical area but becomes rounded as lateral margins approached; maximum width at or just anterior to midlength; pedicle valve bears slight median depression; brachial valve often bears corresponding median ridge; anterior commissure most often recti-marginate to slightly sulcate; ventral interarea short, narrow; numerous radial costellae which increase anteriorly both by intercalation and bifurcation; at anterior commissure there are 18 to 20 costellae per 5 mm, near midline; costellae occasionally crossed by concentric growth lines near anterior margins.

Gypidula sp. - Elongate oval to subcircular in outline; pedicle valve swollen; costate to multicostate; almost identical to Pentamerella arata (see description below) but can be differentiated by a pedicle fold and brachial sulcus whereas Pentamerella has a pedicle sulcus and brachial fold.

Leptaena aff. "rhomboidalis" - Transversely subquadrate in outline, concavoconvex to slightly biconvex with pedicle valve strongly geniculate at anterior and lateral commissures; brachial valve correspondingly geniculate within pedicle trail; hinge line straight, pedicle interarea flat; ornamentation consists of radial costellae which extend past point of geniculation and continue on trail of valves; concentric rugae cross costellae becoming larger anteriorly.

Levenea aff. subcarinata - Shells small to medium sized, transversely suboval in outline, ventribiconvex in lateral profile; brachial valve bears shallow, rounded sulcus which broadens anteriorly; length slightly greater than width; maximum width at or just anterior to midlength; ventral interarea short, slightly incurved; triangular delthyrium encloses angle of approximately 60 degrees; delthyrium often widens apically into small, circular foramen; ornamentation consists of rounded, radial costellae which increase in number anteriorly by bifurcation.

Megakozłowskiella raricosta - Shells subtransverse in outline, strophic, medium to large, ventribiconvex; hinge line straight; pedicle interarea moderately narrow with striae which parallel hinge line; brachial interarea extremely narrow; distinct slightly flattened fold on brachial valve and corresponding deep, U-shaped sulcus on pedicle valve; commonly three plications on flanks; delthyrium includes angle of approximately 60 degrees; no deltidial plates preserved; anterior commissure uniplicate; strong, concentric growth lamellae with anterior frills; radial ornamentation consists of very fine striae or capillae.

Megastrophia sp. - Medium sized to large, subsemicircular to transversely suboval in outline; somewhat alate, concavo-convex in lateral profile; maximum width attained at hinge line; unequally parvicostellate to subuniformly costellate; pseudodeltidium flat, complete, with narrow median ridge; chilidium flat, complete, with median ridge; hinge entirely denticulate.

Meristina cf. nasuta - Convex, elongate and suboval in outline with no noticeable interarea. Unequally biconvex, with pedicle valve much deeper than brachial valve; maximum width commonly anterior to midlength; delthyrium broad, triangular and opens apically into semicircular foramen; faint pedicle sulcus modified by development of low, rounded medial plication that extends anterior commissure in tongue-like projection; concentric growth lamellae evident at anterior portion of valves but remainder of shell smooth.

"Mucrospirifer" cf. macra - Small to large alate shells transversely subtrigonal to subsemicircular in outline; biconvex in lateral profile with brachial valve slightly flatter than pedicle valve; ventral interarea moderately high, long, somewhat curved; ventral beak, posterior to interarea, short and stubby; open, triangular delthyrium present which divides interarea medially; dorsal interarea long, thin, ribbon-like; brachial valve bears high, medial fold flattened at top; pedicle valve bears corresponding U-shaped sulcus; surface of shells covered by sharply defined plications ranging from U-shaped to subangular in cross section; numerous, concentric, frilly growth lines present; no fine radial ornamentation.

Nucleospira aff. ventricosa - Small, transversely suboval in outline, biconvex in lateral profile with pedicle valve slightly deeper than brachial valve; hinge line curved; brachial beak fits into anterior end of delthyrium which is partially covered by concave pseudodeltidium in some specimens; both beaks erect, no interarea evident; shell surface lacks radial ornamentation; no fold or sulcus present; pedicle valve shows faint median depression in some specimens; concentric growth lamellae present, more concentrated towards rectimarginate anterior commissure.

Orthotetacid indet. - Small to medium sized shells, generally poorly preserved as internal impressions; hinge line straight; ornamentation finely costellate;

Pentagonia unisulcata - Medium sized, nonstrophic, pentagonal in outline when viewed posteriorly; beak suberect, dorsibiconvex with greatest width attained between midlength and anterior commissure; brachial valve cariniform due to presence of raised, rounded fold bearing narrow, median groove; in some forms groove widens slightly anteriorly forming two parallel to subparallel ridges extending almost half the valve length; flanks concave, dropping steeply away from sulcate fold; sulcus broad, shallow with two

distinct ridges which define sulcus laterally and extend from umbo across posterolateral margins of flanks to uniplicate anterolateral commissure; vague, concentric growth lines on anterior portion of shell.

Pentamerella arata - subglobose and broadly pyriform in outline with strongly convex pedicle valve and weakly convex brachial valve; hinge line short, curved, narrow; no interarea evident; pedicle valve beak short, strong, incurved, not closely pressed against brachial beak; brachial beak small, less erect and less incurved; maximum width attained at or about midlength; weak sulcus present on anterior half of pedicle valve with corresponding fold on brachial valve (note: this morphological feature is the key to differentiating Gypidula from Pentamerella; see Gypidula above); both valves ornamented with numerous, rounded, bifurcating plications which become narrower on lateral slopes than near midline; interspaces between plications U-shaped and wider than plications which tend to become slightly V-shaped in cross section on some specimens; About 5 plications in sulcus and 6 on fold; concentric growth lines more numerous anteriorly.

Schizophoria cf. multistriata - Shells medium sized, suboval to subquadrate in outline, unequally biconvex; brachial valve deeper and more uniformly convex; in juveniles both valves become almost equally biconvex; pedicle valve develops broad, shallow sulcus on adult forms; brachial valve bears indistinct fold; hinge line short, slightly rounded; maximum width attained at or just past midlength; ventral interarea triangular, fairly high in larger shells, relatively narrow in younger ones; dorsal interarea narrower; interareas of both valves equal to about one-half width; ornamentation consists of rounded to subangular radial costellae with broad, flat inter spaces; about 11 costellae in a 5 mm space near anterior commissure at midline.

Strophodonta cf. demissa - Subcircular to shield shaped shells, concavoconvex in lateral profile; shells wider than long; point of maximum width at hinge line; lateral margins almost straight posteriorly; anterior margins evenly rounded; all margins crenulate; anterior commissure rectimarginate; costellae coarse, bifurcating with angular interspaces in cross section.

Stropheodontid-indet. - Small, subcircular in outline, alate; Smooth exterior with irregularly spaced growth lines.

Trematospira sp. - Elongate, suboval, moderately biconvex; dorsal fold and ventral sulcus; hinge line curved, narrow interarea; conjunct deltidial plates; pedicle foramen mesothyridid; approximately 45 rounded, bifurcating costae on dorsal exterior and 19 on ventral exterior; beak pointed and apically twisted.

Table 4. Other faunal constituents of the Onondaga Limestone in central New York.

Taxa	Common	Rare	Very Rare
Gastropods			
<u>Platyceras</u>		X	
<u>Tropidodiscus</u>			X
<u>Straparollus</u>			X
<u>Liospira</u>			X
<u>Ecculiomphalus</u>			X
<u>Loxonema</u>			X
<u>Platystoma</u>			X
Euomphalacean fragments			X
Cephalopods			
<u>Foordites</u>			X
<u>Halloceras</u>			X
cf. <u>Goldringia</u>			X
Trilobites			
<u>Phacops cristata</u>	X		
<u>Odontocephalus</u>	X		
<u>Dechenella</u>			X
cf. <u>Otarion</u>			X
cf. <u>Proetus</u>			X
Dalmanitid fragments			X
Crinoids			
Camerate columnals	X		
Calyces			X
Bryozoans			
<u>Dyoidophragma</u>			X
<u>Fistulipora</u>			X
Sponges			
<u>Hindia</u>	X		

DISTRIBUTION OF MEGAFOSSILS

Corals dominate the shallow water habitat of the Edgecliff and Lithofacies VI. Common rugosans include Acinophyllum, Cystiphyloides, Heliophyllum, Heterophrentis, and Siphonophrentis while the tabulates are represented by Emmonsia, Favosites and Lecfedites. Brachiopods include Amphigenia, Atrypa, Elytha, Leptaena and Levena. The trilobite Phacops is present in small numbers; platyceratid gastropods are locally common and crinoid columnals abound. Bryozoans are rare and almost never intact.

In the field trip area, the fauna of the upper Edgecliff and lithofacies II differs little from that of Lithofacies VI. While most of the fauna remains unchanged, though somewhat less abundant, Lecfedites and fenestrate bryozoans become more common. This appears to be in response to diminishing water agitation.

Brachiopods, including Atrypa, Megakozlowskiella and Megastrophia, dominate the fauna of the Nedrow member and Lithofacies III. The lithistid sponge Hindia is quite common low in the section. The rugose corals Amplexiphyllum, Heliophyllum and Heterophrentis are present, though not particularly common in central New York. Phacops and Odontocephalus are common trilobites while the gastropods are represented by Platyceras and Platystoma and the cephalopods by Goldringia? and Foordites.

The offshore habitat of the Moorehouse and Seneca members and Lithofacies V is, in the field trip area, dominated by trilobites (Phacops, Odontocephalus) and a diversity of brachiopods including Atrypa, Athyris, Coelospira, Leptaena, Mucrospirifer and Strophodonta. "Chonetes" is relatively common in the Moorehouse and extremely abundant high in the Seneca. The rugose coral Heterophrentis as well as the tabulates Aulopora and Syringopora are often present though not at all common.

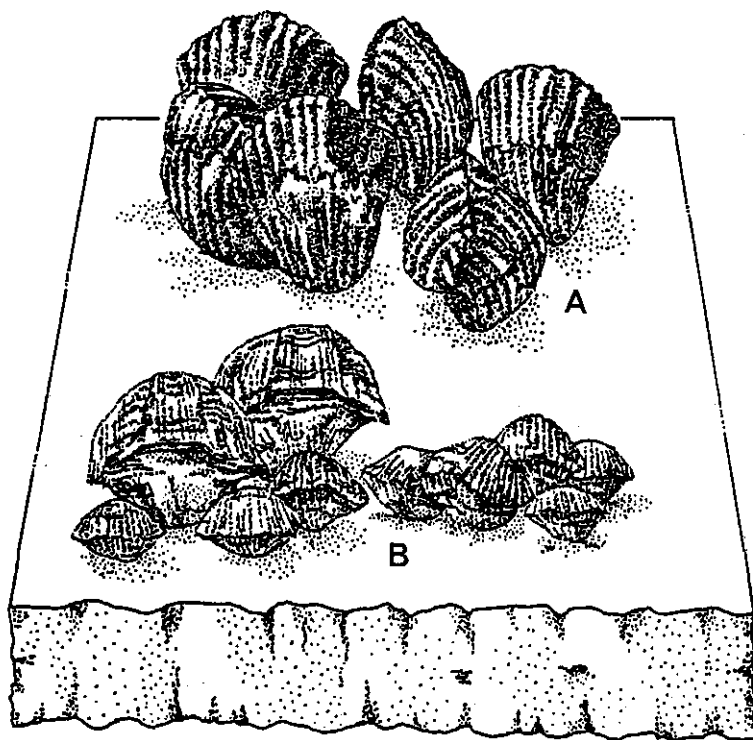


Figure 2. A, Pentamerella arata and B, Atrypa "reticularis" shown in cosupportive life habit. This refers to those ambitopic brachiopods which maintain an umbo-down posture and are packed tightly together often growing on one another.

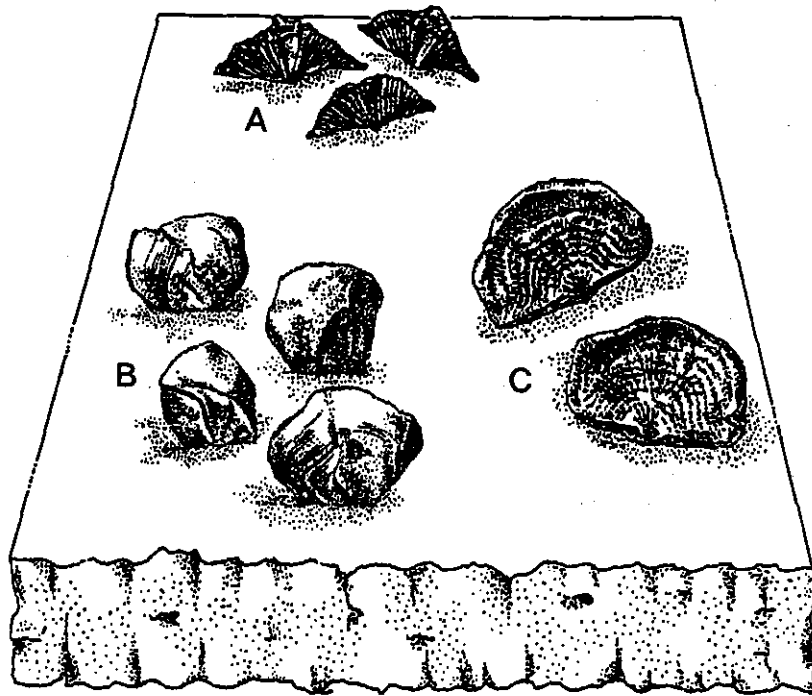


Figure 3. A, Mucrospirifer cf. macra (rhizopedunculate); B, Pentagonia unisulcata (ambitopic); C, Lep-
taena aff. "rhomboidalis" (quasi-infaunal).

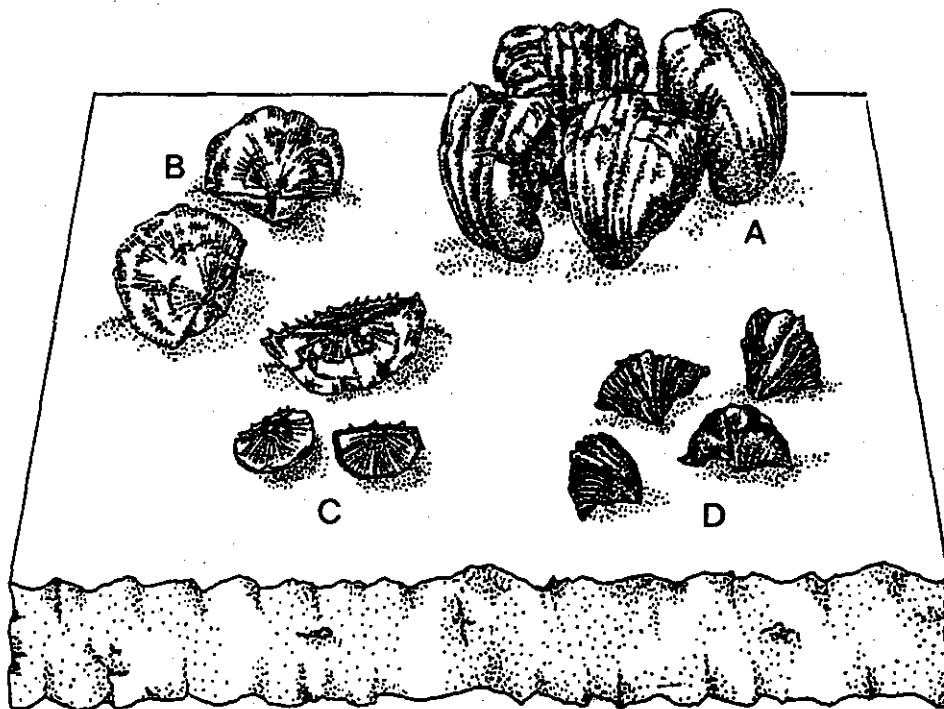
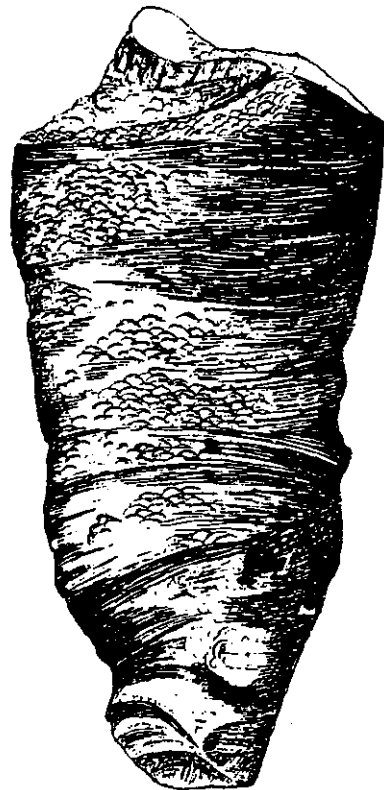


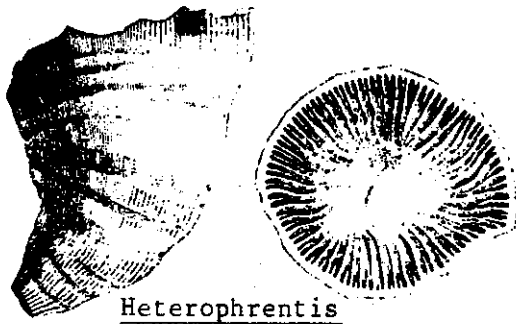
Figure 4. A, Gypidula sp. (Cosupportive); B, Orthotetacid indet.; C, "Chonetes" sp.; D, Cyrtina hamilton-
ensis (extremely rare in central New York) (all ambitopic).



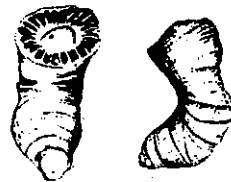
Acinophyllum



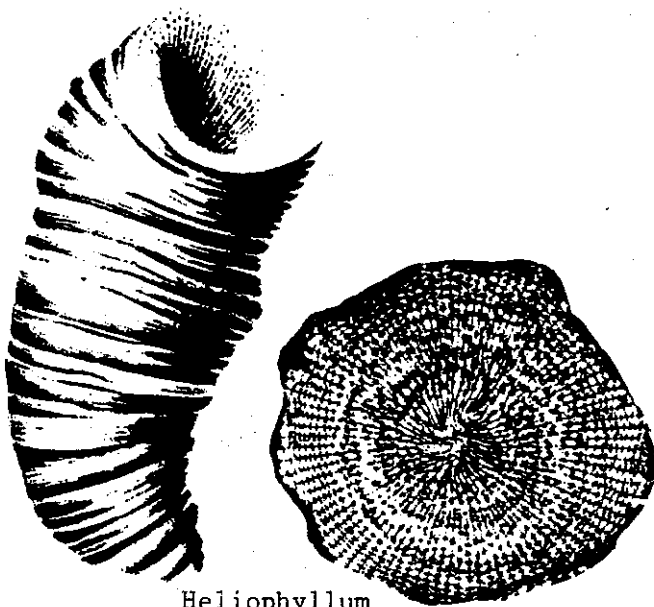
Cystiphyllodes



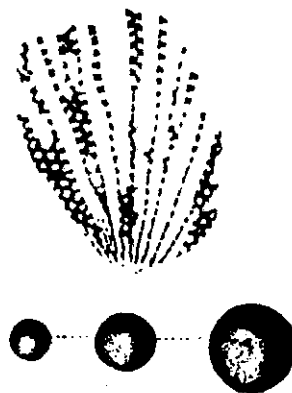
Heterophrentis



Amplexiphyllum

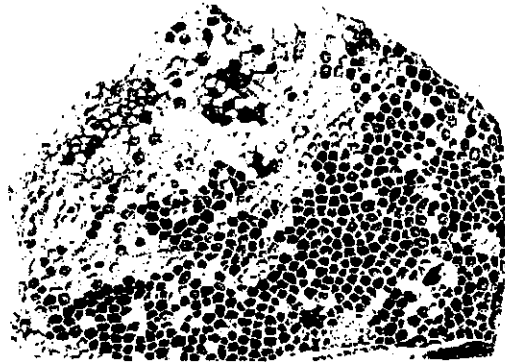


Heliophyllum



Hindia

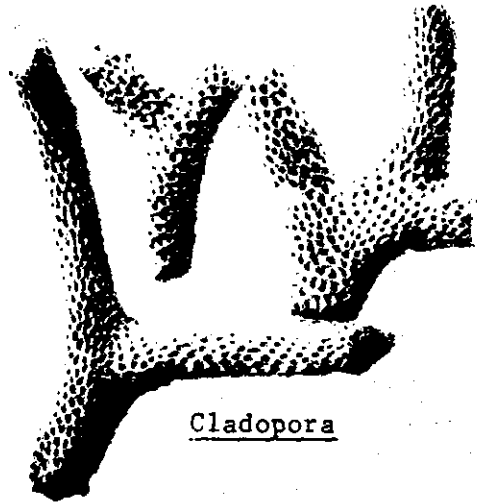
Figure 5. Corals and sponges of the Onondaga Limestone.



Favosites



Emmonsia



Cladopora



Lecfedites



Syringopora



Aulopora

Figure 6. Corals of the Onondaga Limestone.

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ROAD LOG FOR THE ONONDAGA LIMESTONE IN CENTRAL NEW YORK

CUMULATIVE MILEAGE	MILES FROM LAST POINT	ROUTE DESCRIPTION
		Leaving Ithaca take N.Y. Route 13 east to I81 north, approximately 27 miles.
0.0	0.0	Leave I81 at exit 16 and turn left onto U.S. Route 11 north, approximately 24 miles. Log mileage begins at this point.
0.7	0.7	Immediately after passing beneath I81 turn left from Route 11 onto Quarry Road.
1.0	0.3	STOP 1: Park on the road shoulders and walk into the south end of the quarry which lies between Quarry Road and the Interstate.
<p>NOTE: This quarry is on the Onondaga Indian Reservation. Do not, under any circumstances, enter without prior permission from the Tribe!</p> <p>The quarry's south end is floored by the Edgecliff Member. Above this about 13' of Nedrow is exposed (type section). The Moorehouse is 19' thick and is overlain by an incomplete section of the Seneca Member. The Tioga Bentonite horizon is evident as the glacially scoured surface above the quarry's northeast wall where it forms a reentrant high in the south-east wall (it can be located between two chert horizons).</p>		
1.3	0.3	STOP 2: Return to Route 11 and park in the lot on the road's west side. Walk across the road to the outcrop between Route 11 and the I81 exit ramp.
<p>The entire Edgecliff and lower Nedrow are exposed here in a weathered condition which may be more suitable for collecting than the quarry exposures.</p> <p>Leave the parking lot and head left (north) on Route 11.</p>		
4.2	2.9	Turn right (east) onto Route 173.
8.3	4.1	In Jamesville turn left (north) onto Solvay Road.

9.0 0.7 Turn right (east) onto the entrance road to the Jamesville Quarry.

STOP 3: Continue along the entrance road and stop at the quarry office.

This is reputed to be the largest quarry in New York State. Hard hat and prior permission are required. The entire thickness of the Onondaga, from the phosphatic "Springvale Sandstone" to the Seneca and Marcellus Shale, is exposed near the quarry's southeast corner. However, quarry operations have left only the "Springvale," Edgecliff and Nedrow section readily accessible. The Tioga Bentonite and Seneca/Marcellus at the quarry top can be safely seen only from a distance. Exercise extreme caution in proximity to the quarry walls; hard hats offer little protection from large falling rocks.

9.7 0.7 Leave the quarry via Solvay Road, return to Route 173 and turn right (west) back toward Syracuse.

18.9 9.2 Turn left onto Split Rock Road. The road sign may not be visible, so watch for a yellow-and-blue state historical marker.

19.6 0.7 STOP 4: Continue to the end of Split Rock Road and into the quarry entrance.

This is the type section of the Edgecliff Member which is named for Edgecliff Park, located just to the west. The member's full thickness of 2.3 m is exposed in the upper areas of the main quarry and its erosional base is marked by the presence of phosphate nodules. It is interesting to note the pronounced difference between the "Springvale" horizon here and in the Jamesville Quarry. In the southern quarry wall, 3 m of Nedrow are exposed, but the top of the Nedrow has been removed by glacial scour.

END OF TRIP

STRATIGRAPHY AND PALEONTOLOGY - AROUND CAYUGA LAKE ONCE AGAIN

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INTRODUCTION

When the Association last met at Cornell on May 9, 1959, the three of us were soon to be graduate students at the University. The classic Devonian localities visited on Trip C were our formal introduction to a succession of rocks that has enlightened and challenged generations of students. Perhaps nowhere else in the State can such a variety of platform, slope and basin facies and faunas of the Catskill Delta be so profitably seen on a one-day excursion. Much progress has been made in understanding these rocks in the past 27 years and our purpose today is to retrace the route of Trip C in light of subsequent study. The advances in lithostratigraphy, paleoecology and goniatite and conodont biostratigraphy have been particularly notable. It seems especially appropriate that in the 150th Anniversary year of the founding of the New York State Geological Survey, we return to the geographic center of the Devonian outcrop and the succession described in the famed district reports of Vanuxem (1842) and Hall (1843).

The sequence of stops of the 1959 trip is followed and an optional stop (Stop 7) in the Renwick Member near Ithaca, has been added (Fig. 1). The stop descriptions in the 1959 Guidebook have been recast and revised where appropriate and discussions of economic and environmental factors have been added. Recent summaries with references to work since 1959 are Rickard (1975, 1981), Oliver and Klapper (1981), Woodrow and Sevon (1985) and Kirchgasser and others (1985).

Lower and Middle Devonian

The Devonian section in New York State begins mainly with carbonates (Helderberg Group and Onondaga Limestone) deposited during Early and Middle Devonian time, which are overlain by a thick, upward coarsening, regressive, clastic sequence that represents a westward migrating deltaic complex (Catskill Delta) constructed during later Middle and Late Devonian time.

The Lower Devonian Helderberg Group in the Cayuga Lake region is approximately 50 feet thick and represents a peritidal, lagoonal depositional setting. The nearly unfossiliferous Rondout is overlain by the low diversity Manlius Limestone containing the brachiopod *Hewellella vanuxemi*, tentaculitids and some stromatoporids.

The overlying 4 feet of Oriskany Sandstone (Stops 4 & 5), resting unconformably on the Helderberg Group, represents a transgressive, shallow water, high energy, near-shore environment. This inference is supported by the unit's coarse grain lithology and a fauna dominated by large, epifaunal, filter feeding brachiopods such as *Costispirifer arenosus*, *Hipparionyx proximus* and *Rensselaeria ovoides*. Gastropods (*Platyceras*) and bivalves are rare.

The Middle Devonian Onondaga Limestone, approximately 100 feet thick, disconformably overlies the Oriskany Sandstone and is characterized by a high diversity fauna dominated by corals, crinoids, brachiopods, and a variety of snails, bivalves, and trilobites. A low turbidity, relatively clear water, shallow shelf environment prevailed during Onondaga time in the Cayuga Lake region.

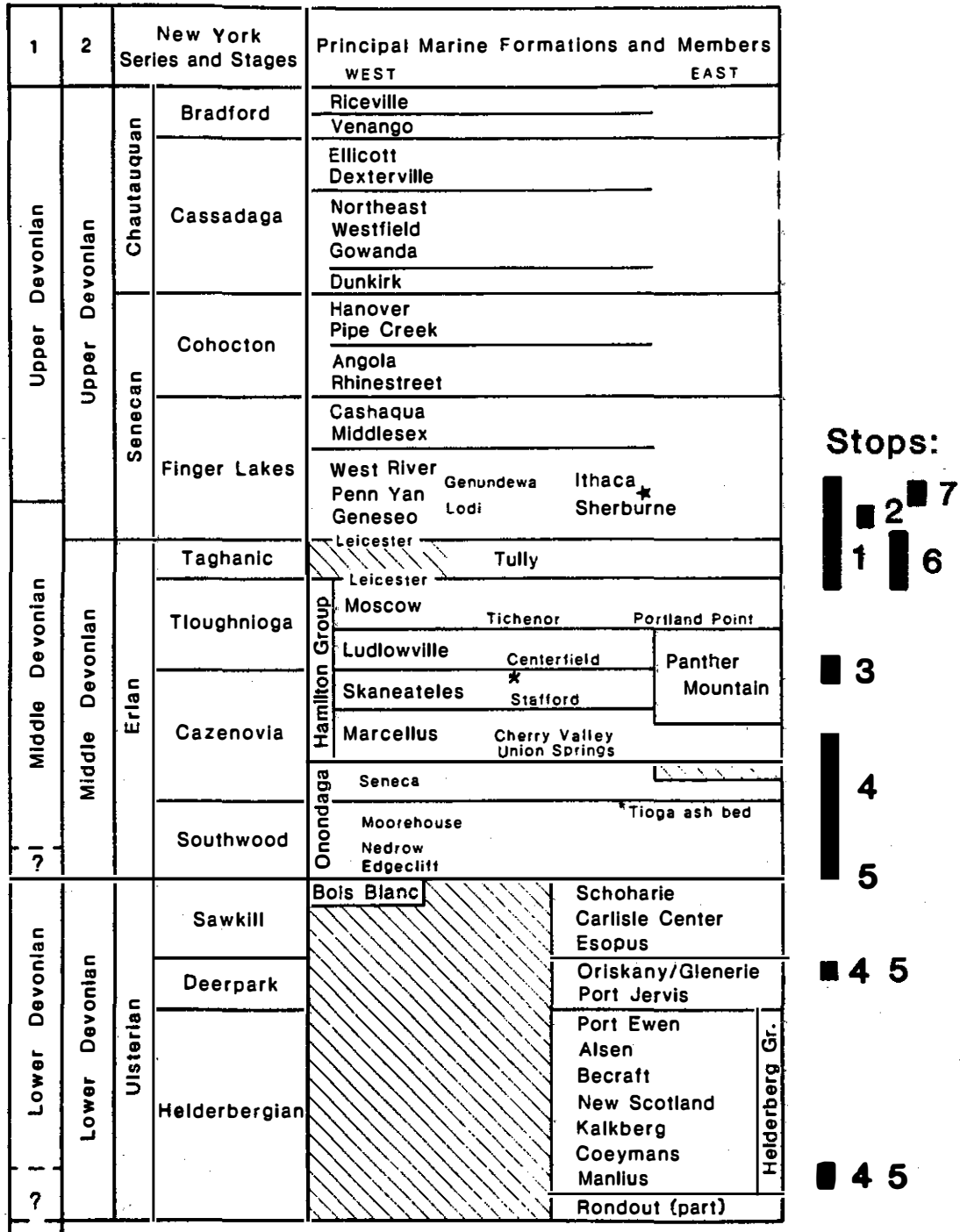


Fig. 1. Devonian section in New York with locations of field trip stops. Asterisk = Ledyard Member; Star = Renwick Member. From Kirchgasser and others (1985).

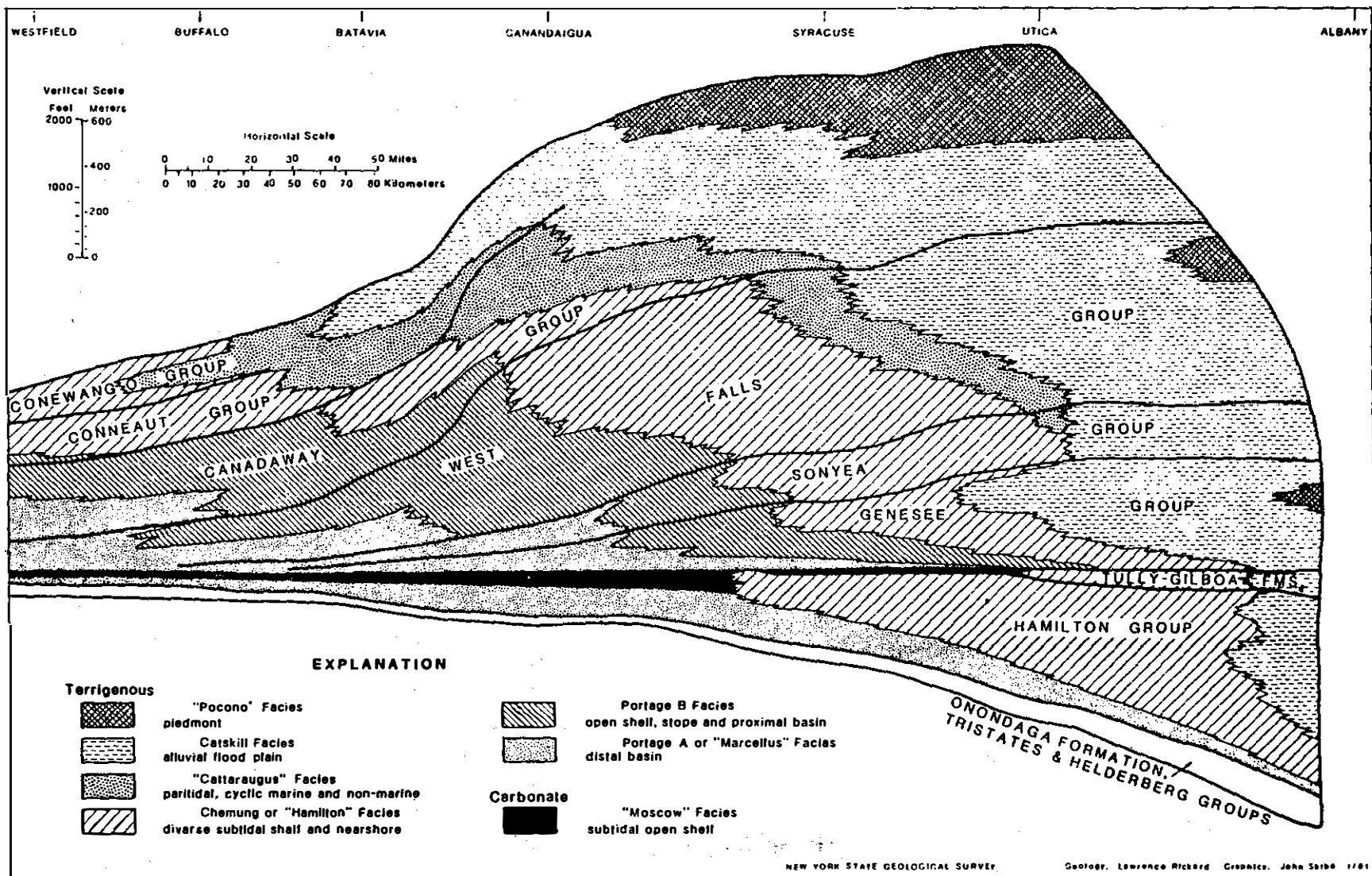


Fig. 2. Generalized stratigraphic cross section of New York Devonian. From Rickard (1981).

The Hamilton Group overlies the Onondaga Limestone and records the initial clastic influx of the Catskill wedge. Detrital sediments were derived from the Acadian landmass east and southeast of New York. The highly fossiliferous Hamilton shales, mudstones, and minor limestones of western New York pass eastward into thicker and coarser siltstones and sandstones in central New York. In the Catskill Mountain region of eastern New York the Hamilton Group consists primarily of non-marine red beds (Fig.2).

Although Lardner Vanuxem (1842) and James Hall (1843), in their monumental reports on the third and fourth geological districts respectively, laid the foundation for subsequent work, the general stratigraphic relationships of the Hamilton Group presently understood were first clarified by Cooper (1930, 1933).

Within the Hamilton Group are three geographically extensive carbonate horizons which serve as key beds. These relatively thin units, do not change facies laterally as rapidly as the synjacent units. They pierce the clastic wedge, subdividing the Hamilton into four units, in ascending order: the Marcellus, Skaneateles, Ludlowville, and Moscow Formations (Fig.1). The lowest of these carbonate units, the Stafford Member (west of Cayuga Lake) and the Mottville (east of Cayuga Lake) marks the base of the Skaneateles Formation and separates it from the underlying Marcellus Formation. The other Hamilton carbonate horizons that serve as formational boundaries are the Centerfield Member (Stop 3), between the Skaneateles and Ludlowville Formations, and the Tichenor Member (western New York) and the Portland Point Member (central and eastern New York), between the Ludlowville and Moscow Formations.

Since the carbonate horizons occur in a predominately clastic deltaic sequence, the assumption had been made that they represent brief episodes of transgressive maxima (McCave, 1973). The studies of Brett and Baird (1982, 1985), Baird and Brett (1983), Brett and others (1983), Grasso (1983, in press), Gray (1983,1985), Selleck (1983) and others indicate that these, as well as other Hamilton units, are parts of shallowing-upward or regressive sequences. Eustatic lowering of sea level may have been the causal agent for the Hamilton regressive cycles. Johnson and others (1985) recognize several such events in the Devonian of North America and elsewhere. The vertical succession of litho- and biofacies within the Mottville, Centerfield and Portland Point clearly indicate upward shallowing sequences. These units are in turn sandwiched between deeper water deposits above and below (in ascending order: Oatka Creek, Levanna, Ledyard and Kashong). Therefore, the time-equivalent units in the non-marine facies that correspond to the Mottville, etc., should be found at or near the top of the regressive tongues rather than at the base of transgressive tongues.

Hamilton paleogeography, as proposed by Brett and Baird (1985), consists of a western muddy shelf separated from an eastern predominately silty and sandy shelf by a centrally located, subsiding trough. The trough-axis trends from the northern Cayuga region southwest through the central Seneca Lake region. The basin also opens out or becomes wider in this direction. In general, as Hamilton stratigraphic units are traced from the west into the Seneca-Cayuga Lakes region, they become thicker, less carbonate rich, more shaley, and less fossiliferous. This has been well documented for the Centerfield Member by Gray (1985) and for the Portland Point Member by Baird (1979).

Middle and Upper Devonian

Tully Formation. The shallow water, moderately fossiliferous limestones of the Tully Formation record a major break in the Devonian succession, separating the platform sediments of the Hamilton Group described above from the transgressive-regressive cycles of clastic sediments of the later Devonian progradational phase of the delta (Fig.2). The Tully marks the beginning of a major transgressive event (Taghanic Onlap) following late Hamilton regression and erosion that culminated in the succeeding anoxic Genesee black shale (Genesee Formation). This major mid-Devonian deepening event has been recognized around the world (House,1983; Johnson and others,1985). The stratigraphic complexities (disconformities, facies and thickness changes) and tectonic controls of sedimentation through the Tully interval have been documented in major studies by Johnson and Friedman (1969) and Heckel (1973).

A late Middle Devonian age for the Tully is indicated if the Middle/Upper Devonian conodont boundary recommended by the Devonian Subcommittee (Ziegler and Werner, 1985) is accepted. However, there are

historical and faunal arguments (Tully *Pharciceras*) for the boundary to be at the level of the Tully (House, 1982b; Kirchgasser and others, 1985; and Rickard, 1985) (Fig.3).

The conodont age of the Tully is Middle and Upper *varcus* Zone (Ziegler and others, 1976; Klapper, 1981). On the goniatite scale the occurrence of *Pharciceras amplexum* and *Phar.* sp. indicate a position low in the *Pharciceras* Stufe of the European standard (House, 1985; House and others, 1985; Kirchgasser and others, 1985).

The benthic fauna of the Tully, dominated by brachiopods and numbering some 250 species listed by Cooper and Williams (1935), is of predominantly Hamilton affinity and belongs to the provincial Eastern North American Faunal Realm. Species ranges through the Tully interval indicate a major turnover in the upper part associated with the major deepening event referred to above. The Tully-Genesee transgression (Taghanic Onlap) resulted in the breakdown of faunal provinciality as western North American (Old World) elements entered the Appalachian Basin (Johnson, 1970,1971).

Genesee Formation - The Genesee Formation is the lowest of the major transgressive-regressive cycles of clastic facies that characterize the prograding post-Tully Catskill Delta (Fig.4). At Cayuga Lake the Genesee cycle consists of the transgressive Genesee black shale member (distal basin facies) overlain by progradational siltstone facies of the Sherburne (Penn Yan in lower part) and Ithaca Members (proximal basin, slope and open shelf facies of the delta-front) interrupted by a second major transgressive black shale (Renwick Member).

A major correlation change since the 1959 meeting resulted from tracing of the Middlesex black shale. This initial transgressive tongue of the next cycle (Sonyea Formation) is equivalent to a level at the top of the Ithaca at Cayuga Lake (Sutton, 1959, 1965; deWitt and Colton, 1959, 1978). The Sherburne and Ithaca were shown to thin and interfinger basinward with the Penn Yan and West River shales (distal basin) of the Genesee Formation. They do not correlate with the western Portage or Naples beds (Cashaqua shale) of the post-Middlesex Sonyea Formation as was thought at the time of the 1959 meeting. The classic benthic Ithaca fauna and associated pelagic fauna (including goniatites and conodonts) are not contemporaneous with the pelagic Naples fauna of the Cashaqua but are older and distinct Genesee faunas (Kirchgasser, 1985).

Sections published in deWitt and Colton (1978) document the broad outline of correlations within the Genesee Formation. Further refinement has come from tracing of thin black shale, pyrite, shelly carbonate and siltstone horizons (some with goniatites, datable conodonts and distinct gamma-ray signatures) between the basin and delta-front (Figs.4 and 5). The facies shifts around the delta-front are complicated in detail and lithostratigraphic boundaries are difficult to draw. Some revisions based on work in progress by Gordan Baird and Carlton Brett are included here (Fig.5). Newly discovered is an interval in the upper Genesee (Hubbard Quarry shale), between the Fir Tree (pyrite/inarticulate debris/siltstone) and Lodi horizons, that thickens and becomes more silty southward (shoreward) along Cayuga Lake. This interval has a bearing on the placement at Cayuga Lake of the Devonian Subcommission Middle/Upper Devonian boundary (see below).

Biostratigraphic work on the Genesee is well advanced (Fig.3) but much of the taxonomic documentation is unpublished and work on several of the faunas is incomplete. The goniatite sequence, initially outlined in a seminal paper by House (1962), correlates with the *Pharciceras* Stufe and lower *Manticoceras* Stufe (Fig. 3). In the Cayuga Lake region *Pharciceras* sp. (Fauna 13) occurs in the transition beds between the Tully and Genesee (Fillmore Glen Mbr.) and *Epitornoceras* and *Ponticeras* enter high in the Genesee (Fauna 14). *Ponticeras perlatum* ranges from the top of the Genesee into the basal Sherburne (Lodi Limestone) and continues to the base of the Renwick (15). Above, in about the middle Ithaca, *Pont. cf. regale* (Fauna 16) is followed by *Koenenites* (17) with *Hoeninghausia* entering near the top of the member (17c). These Ithaca goniatites were previously thought to be representatives of *Manticoceras* from the Naples Fauna (Kirchgasser, 1985).

The conodont correlations of the Genesee have been refined since publication of a monograph by Huddle (1981) (Klapper,1981; Kirchgasser, and others, 1985). Elements indicating ages from the *hermanni-cristatus* to Lower *asymmetricus* Zones occur along the Hamilton-Genesee disconformity west of the Tully outcrop (Fig. 3). The *disparilis* Zone fauna has been found at the top of the Genesee Member at Seneca Lake (Fig. 5). Immediately above, in the Lodi Limestone (lowermost Sherburne or Penn Yan) the Lowermost *asymmetricus*

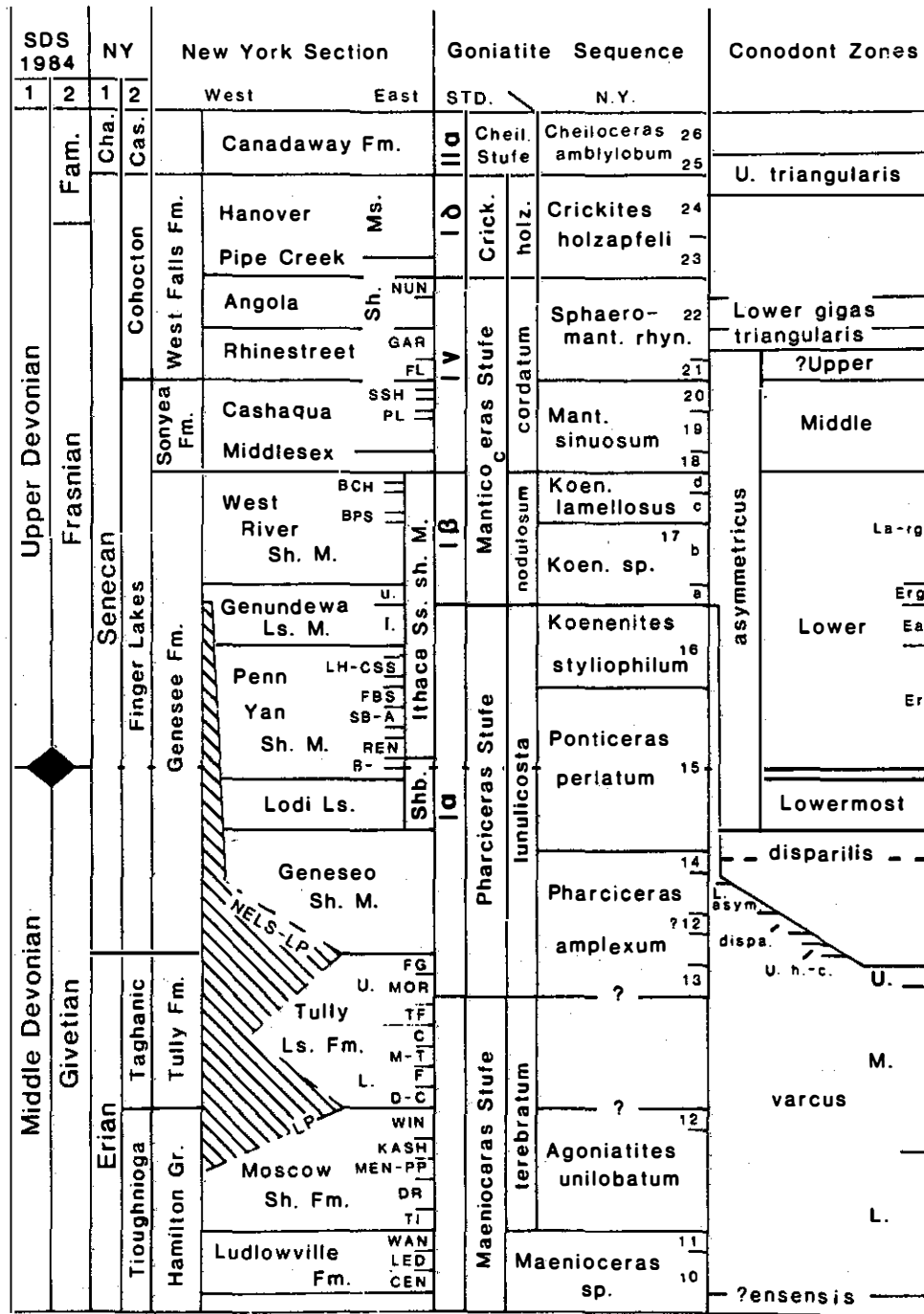


Fig.3. Goniatic (ammonoid) and conodont sequences around Middle/Upper Devonian boundary (diamond symbol) proposed by the Devonian Subcommittee. Shb.=Sherburne Member, B=black shale B, REN=Renwick Member. Modified from Kirchgasser and others (1985).

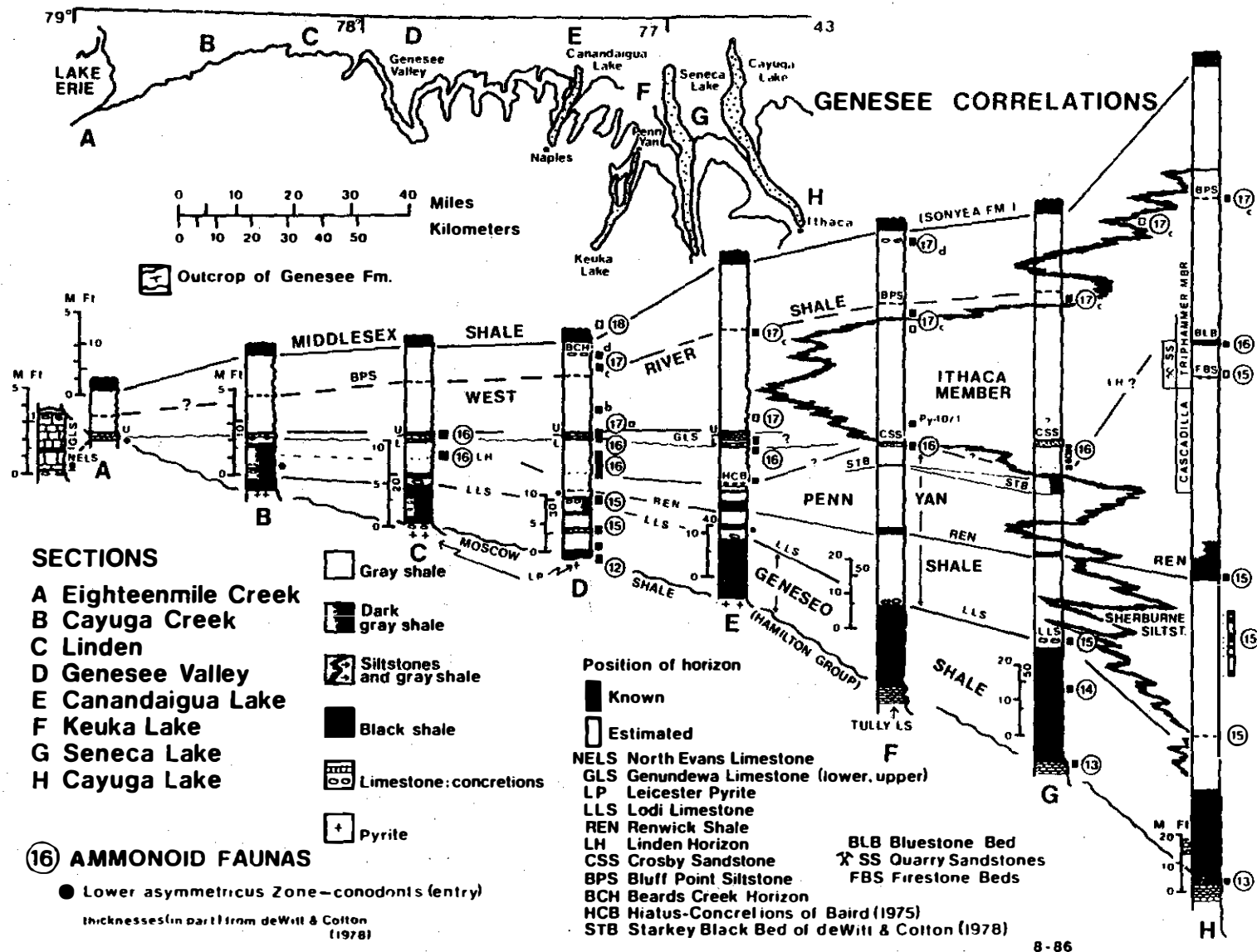


Fig.4. Genesee correlations. Modified from Kirchgasser(1985).

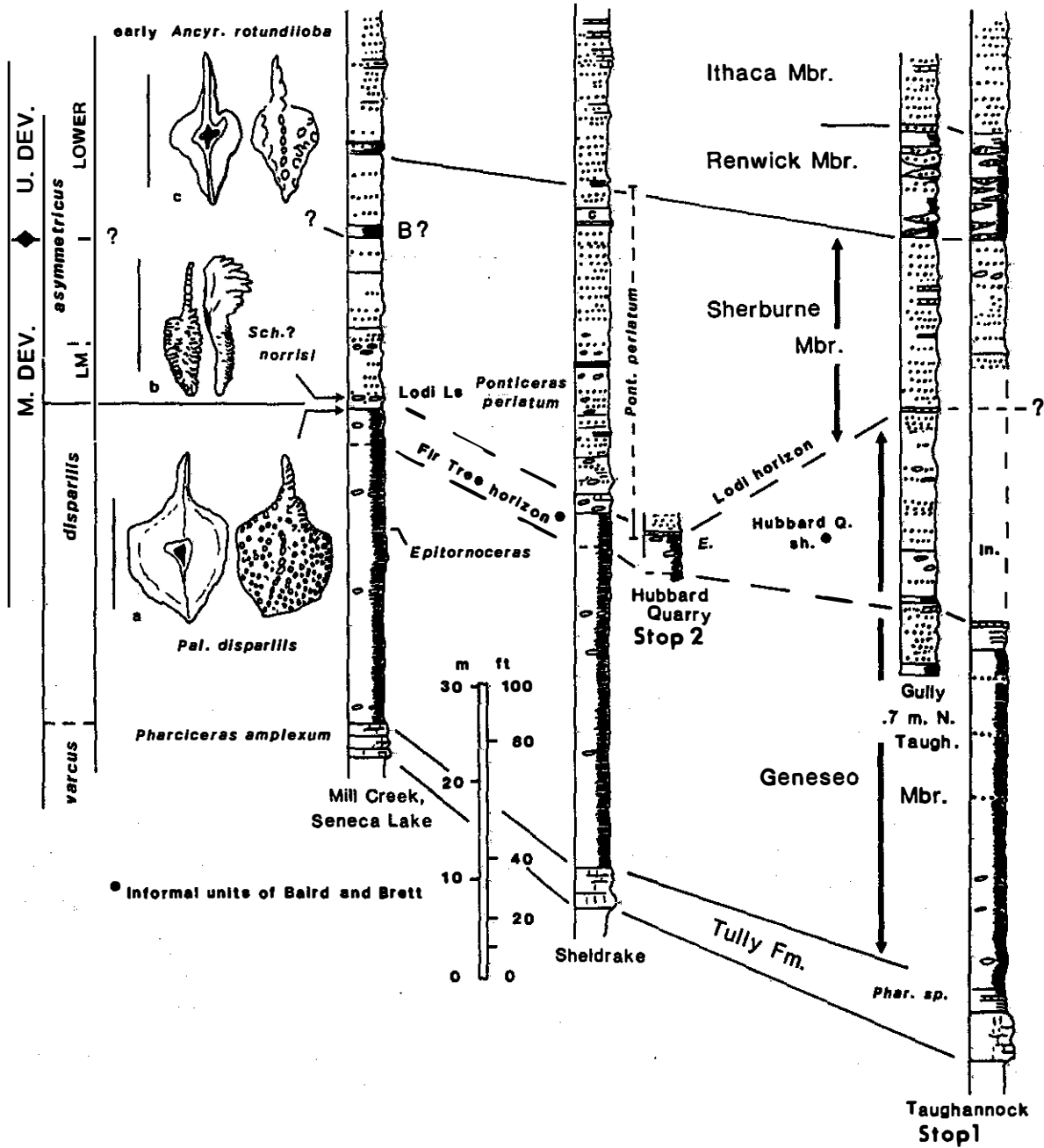


Fig.5. Lower Genesee correlations between Seneca and Cayuga Lakes. Modified from sections in deWitt and Colton (1978, Pl.3). Sketches of conodont species (scale=1mm) based on a. Ziegler and others (1975), b. G.Klapper (manuscript), c. Klapper (1985). (STOPS 1 and 2).

Zone fauna (with *Schmitognathus? norrisi*) occurs. The entry of *Ancyrodella rotundiloba* (Er or early *rotundiloba* form of Klapper, 1985), which marks the start of the lower *asymmetricus* Zone and is the international Middle/Upper Devonian boundary recommended by the Devonian Subcommittee (Ziegler, 1985) has not been documented in the Seneca-Cayuga Lake section. However, based on its entry at the base of black shale B in the lower Penn Yan in the Honeoye Valley, its position must be high in the Sherburne (pre-Renwick).

Thayer (1974) identified several faunal associations in the delta-front facies of the Genesee Formation. The sparse fauna in the black shales is dominated by planktic styliolines and epiplanktic brachiopods, which, along with plant debris and cephalopods, settled on the anoxic bottom (*Styliolina*-brachiopod facies). Above the Genesee shale, the silty, firmer and more oxic bottom conditions of the regressive (prograding) Sherburne supported a fauna characterized by the branching tabulate coral *Cladochonus* and in the lower Renwick the brachiopod, *Warrenella* (*Cladochonus* facies). Above, in the black shales of the Renwick the delicate plumose coral *Plumalina* occurs (*Plumalina* facies). Above the Renwick, in the progradational complex of delta-front facies comprising the Ithaca Member, is the classic benthic Ithaca Fauna of Williams (1884) and Kindle (1896), subdivided into the *Ambocoela* and *Schizophoria*-strophomenid facies faunas. Within this sequence key transgressive or discontinuity horizons with conodonts and goniatites have been correlated with the faunal succession in the basin facies to the west (Kirchgasser, 1985). Thayer (1974) argued that the distribution of the benthic faunal associations around the delta-front was controlled by the relative rates of delta-progradation.

Economic and Environmental Geology

In the years since the NYSGA last met in Ithaca, the public perception of earth science and the utilization of geologic data has changed dramatically. Of paramount importance has been the implementation of the State Environmental Review Act of 1975. This legislation developed a systematic interdisciplinary approach to the consideration of the environmental factors concerning any proposed project. The review mechanism is an effective method to explore ways to minimize adverse environmental impacts. On a regular basis geologists serve as members of project teams preparing terrestrial resource impact assessment, especially in the areas of mining, real estate development, highway realignment, solid waste disposal and hydrogeology.

Mining activity is now regulated by the New York State Mined Land Reclamation Act of 1975. This legislation, administered by the Department of Environmental Conservation, was enacted to foster and encourage the development of the mining industry of the state, while preventing pollution associated with mining activity and assuring the reclamation of mined lands in such a manner as to render them suitable for future productive use. The Department of Transportation is responsible for evaluating the nature of materials and their processing in order to ensure the quality and uniformity of aggregate produced for state contracts.

We will visit commercial mining operations at Canoga (Stop 4) and at Portland Point (Stop 6), at which time the implementation of these regulations will be discussed.

As regional development has progressed in the Cayuga Lake area, the mining of materials required for construction has correspondingly developed. Local development has been necessary due to the relatively high costs of transporting such materials into the market area (in excess of \$0.50/Ton/mile). Despite significant physical and structural constraints, the Cayuga Crushed Stone quarry, producing in the Tully Limestone at Portland Point, has been greatly expanded since 1959. Today, over 9 million dollars worth of concrete and bituminous material, and crushed limestone aggregate are produced annually. This quarry serves as the only source of rip rap for erosion abatement and stream control in the central Southern Tier. Likewise, the quarry at Canoga, now operated by Seneca Stone Company, has been expanded and deepened so that the entire sequence of Onondaga limestone is being processed in order to serve the Seneca Falls- Waterloo-Geneva area.

In 1970 the former Cayuga Rock Salt Company mine at Portland Point was acquired by Cargill Salt. This room-and-pillar mining operation has been renovated and expanded. Production has more than doubled -- from 450,000 Tons/year to over 1 million Tons/year. Until 1962 the International Salt Company operated brine wells at Myers about one mile north of the Cargill mine shaft. The danger of water invasion from this brining operation, and the presence of less deformed salt and better mining conditions below the 1800 foot level has

required the development of a deeper production level. As production is now under Cayuga Lake (part of the State Waterway System), royalties of 1/4 million dollars / year are being paid to the State of New York.

In 1959 the primary natural gas exploration objectives in the region were Oriskany structural traps. Although these still hold some promise of production, the main emphasis in recent years has been the development of Medina/Queenston stratigraphic traps. These tight sand reservoirs, concentrated around the northern end of Cayuga Lake, include the Waterloo Field (1960), the Seneca Falls Field (1965) and the West Auburn Field (1959)(Fig.6). Depending on market conditions, the development and extension of these fields continues. With governmental support and financing, area educational institutions (such as the Union Springs School System and Wells College) have developed wells for onsite utilization.

In recent years, deep seismic surveys have evaluated the potential of overthrust production. Even with the present depressed market conditions, exploration continues at a significant level.

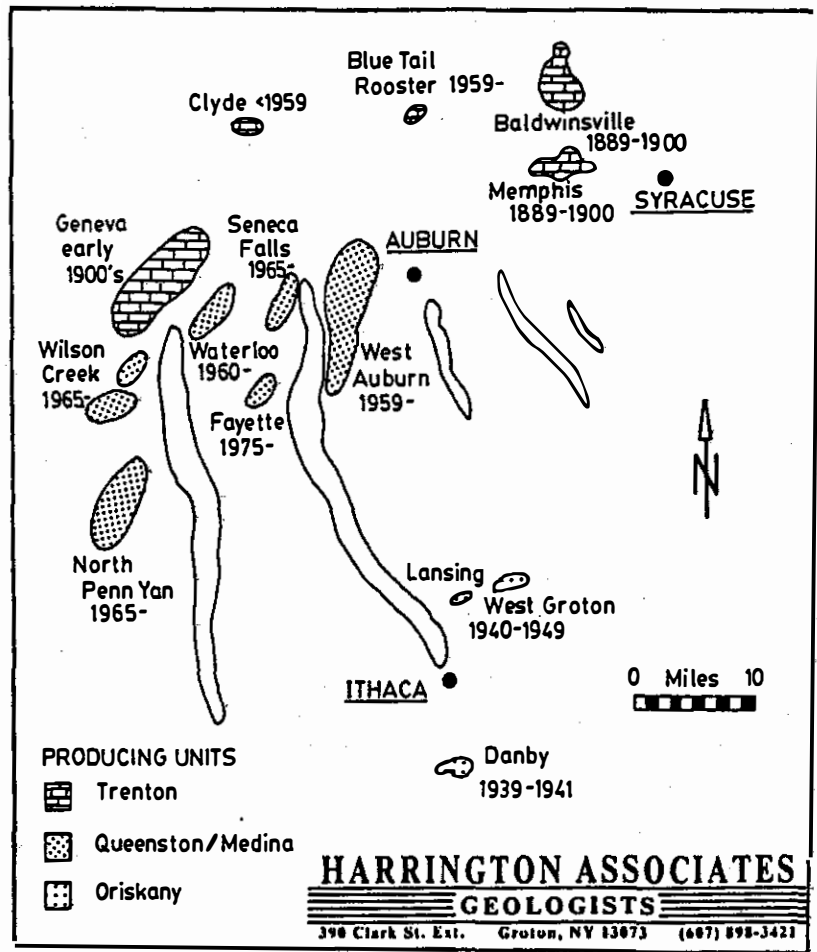


Fig. 6. Gas Fields of the eastern Finger Lakes region. Modified from Harrington Associates reports.

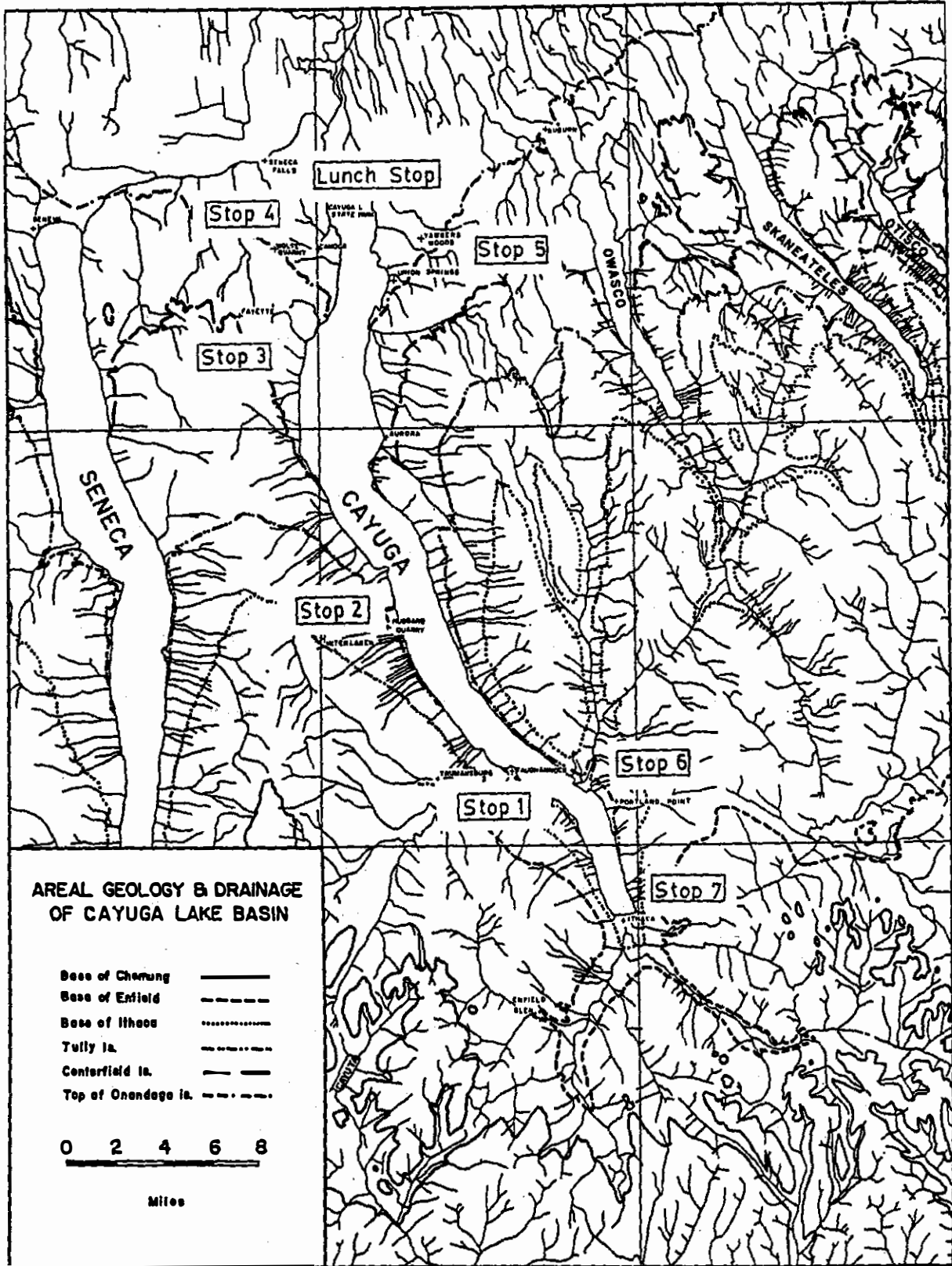


Fig. 7. Route Map for Trip Around Cayuga Lake. Modified after Cole and others (1959).

DESCRIPTION OF FIELD TRIP STOPS

STOP 1A. TULLY LIMESTONE SECTION, LOWER FALLS, TRUMANSBURG CREEK (FIG.8).

No Hammers. Collecting is not permitted in the State Park but will be possible at a similar section across the lake at Portland Point (Stop 6). Five beds of the TULLY FORMATION are exposed in the falls and embankment. The comments on lithologies and benthic faunas which follow are based primarily on data from Heckel (1973).

The contact between the CARPENTER FALLS and TAUGHANNOCK FALLS beds is a widespread discontinuity surface that marks a major lithologic and faunal break within the TULLY. The TULLY as a whole records a progressive rise in sea level (Taghanic Onlap) following late Hamilton erosion but the sedimentary controls are complicated by regional tectonism (Heckel, 1973). A major deepening even has been recognized worldwide at about this time (Middle *varcus* Zone) House (1985), Johnson and others (1985).

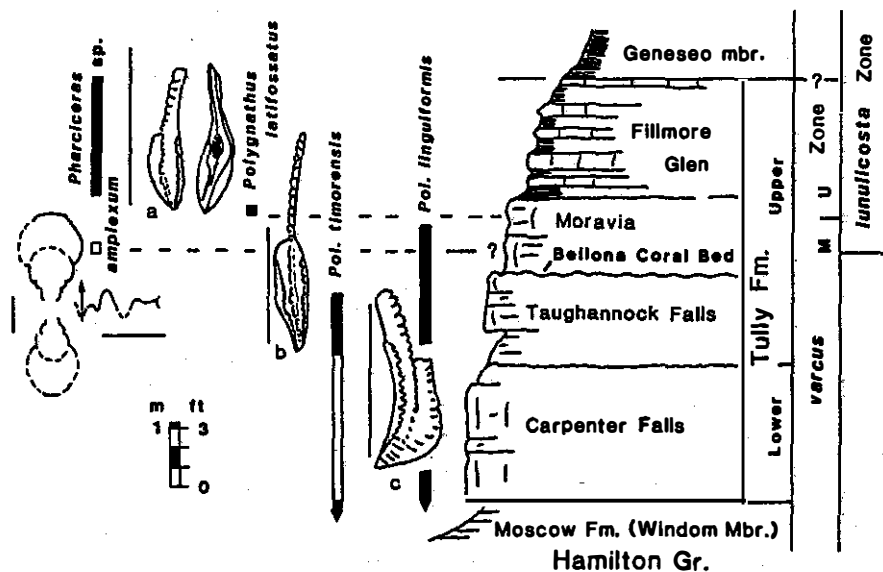


Fig.8. Tully Limestone section in Lower Falls of Taughannock Creek. Adapted from Ziegler and others (1976), Heckel (1973) and deWitt and Colton (1978). Goniatic ranges from Kirchgasser and House (1981); conodont ranges from Ziegler and others (1976) and Klapper (1981). Goniatic sketches based on House and Kirchgasser (manuscript). Conodont sketches a, c, based on Ziegler and others (1976), b, Klapper and others (1970).

The upper CARPENTER FALLS bed consists of thick-bedded burrowed calcilutites with some pelmatozoan calcarenites, which disconformably overlies the WINDOM SHALE of the MOSCOW (several eastern units of the lower TULLY are missing here). The diverse fauna is dominated by brachiopods, including *Emanuella*, *Schuchertella* and *Atrypa*; other fossils include trilobites, styliolines, tentaculites, auloporid corals and ostracods. The main *Hypothyridina* (*H. venustula*) bed of the TULLY occurs in this area at the base of the unit.

The TAUGHANNOCK FALLS BED (5 ft.) at the base of the Upper TULLY marks the start of a new depositional regime and the major faunal turnover in the TULLY. The bed consists of well-bedded burrowed skeletal calcilutite with the "knobby zone" (re-entrant) in the lower part. The fauna is dominated by small horn corals, auloporid corals, styliolines, echinoderms and trilobites. Brachiopods and bryozoans are essentially restricted to the upper part of the bed (*Elytha fimbriata* Zone); Lower TULLY brachiopods are rare or absent. The BELLONA CORAL BED contains rugose, tabulate and auloporid corals including *Heliophyllum*,

Cystiphylloides, *Favosites* and "*Heterophrentis*". The MORAVIA BED is a 4 foot ledge of wavy-bedded limestone lithologically and faunally similar to the TAUGHANNOCK FALLS BED.

The FILLMORE BED (6 feet) conformably overlies the MORAVIA BED and consists of interbedded dark shaley calcilititic limestone and dark calcareous shale transitional to (but separated by a sharp conformable contact) the black GENESEO SHALE. The sparse planktic fauna of the FILLMORE GLEN, which reflects continued rise in sea-level and anoxic bottom conditions, includes styliolines, inarticulates, and rare goniatites.

The boundary between the Middle and Upper *varcus* Zones is near the top of the MORAVIA BED (Ziegler and others, 1976, Table 3; Klapper, 1981 fig. 2). The three elements illustrated in Fig.8 are the common *Polygnathus linguiformis*, *Pol. timorensis*, a *varcus*-type species, and *Pol. latifossatus* which indicates the start of the Upper *varcus* Zone. In the goniatite sequence, the *Pharciceras amplexum* horizon is near the base of the MORAVIA BED but specimens have not been reported from this locality (House, 1962, Kirchgasser and House, 1981). *Phar. amplexum* correlates with a position at the base of the *Pharciceras* Stufe (House and others 1985).

The controversy over the position of the TULLY with respect to the Middle/Upper Devonian boundary continues (Kirchgasser and others, 1985, Rickard, 1985). At the time of the 1959 meeting the mid-Genesee (GENUNDEWA LIMESTONE) boundary of Cooper and others (1942), where the goniatite *Manticoceras* enters, was the accepted position but since 1964, and until recently, the boundary has been placed around the TULLY (Kirchgasser, and others, 1985, Fig. 4). The boundary recommended by Devonian Subcommittee (base of Lower *asymmetricus* Zone) is somewhere in the SHERBURNE and is well above the TULLY.

STOP 1B. TAUGHANNOCK FALLS OVERLOOK, GENESEE FORMATION (FIG. 9).

The main falls and amphitheater, at the head of a one mile long post-glacial gorge with 200 to nearly 400 foot high walls, display a spectacular but mostly inaccessible section in the upper GENESEO, SHERBURNE, RENWICK and lower ITHACA MEMBERS of the GENESEE FORMATION. The top of the falls (215 ft high) is controlled by resistant siltstones in the SHERBURNE MEMBER and by the jointing. The re-entrant at the crest of the falls remains virtually unchanged since the last major rockfall between 1888 and 1892. Near the base of the falls on the left are altered alnoite dikes in N-S joint planes.

About 90 feet of black GENESEO SHALE is exposed in the walls from the level of the plunge pool to the sharp contact with lighter colored shales and siltstones halfway up the falls. Contacts higher up are not easily defined in their fresh joint surfaces. Recent work by Gordan Baird and Carlton Brett shows that the contact at the top of the black shale band is not the same level as the GENESEO/SHERBURNE contact at Hubbard Quarry to the north (Stop 2) and at Seneca Lake to the west. In the gully .7 miles north of Taughannock (Fig.5), the FIR TREE HORIZON is just above the black shale contact. Above are the siltstones and shales of the HUBBARD QUARRY INTERVAL at the top of which is the LODI HORIZON, which defines the GENESEO/SHERBURNE contact (Stop 2). The HUBBARD QUARRY INTERVAL is a silty wedge at the top of the GENESEO MEMBER which intervenes at the south end of Cayuga Lake between the top of the black shale and the LODI. The FIR TREE and LODI HORIZONS have not been seen at Taughannock. The LODI, at the base of the SHERBURNE MEMBER, should be somewhere close to the lip of the falls.

The black shales and lenticular siltstones of the RENWICK MEMBER form a distinctive band in the walls above the level of the falls, and near the top are the lower siltstones of the ITHACA MEMBER. The RENWICK MEMBER, with the *Warrenella* beds at the base will be examined at Stop 7.

The goniatite (*Epitornoceras* and *Ponticeras*) and conodont distributions through the section in this region are illustrated in Fig. 5. The uppermost GENESEO MEMBER is in the *disparilis* Zone and the LODI has a Lowermost *asymmetricus* fauna. Datable conodonts have still to be recovered in this part of the Cayuga Lake section. The proposed Middle/Upper Devonian boundary at the base of the Lower *asymmetricus* Zone should occur at Taughannock somewhere in the SHERBURNE, above the lip of the falls and below the RENWICK.

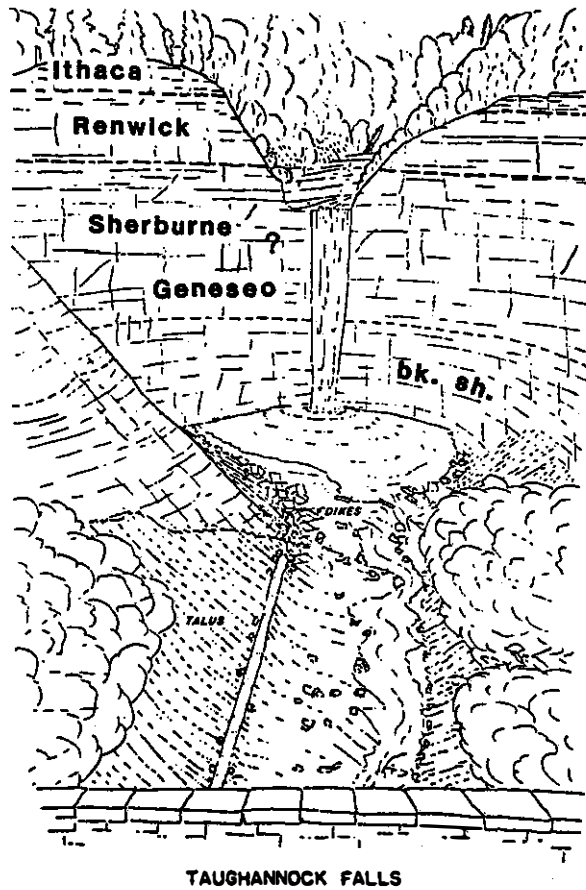


Fig.9. Overlook of Taughannock Falls. Modified from Cole and others (1959)

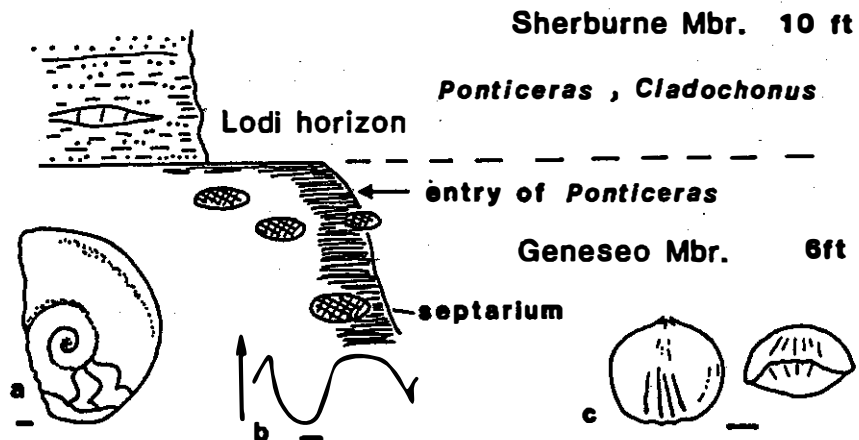


Fig.10. Upper Genesee and lower Sherburne at Hubbard Quarry. Modified from Cole and others (1959). Sketches of fossils based on a. *Pont. perlatum* : Kirchgasser (1985), b. *Epitornoceras peracutum* : House and Kirchgasser manuscript), c. *Leiorhynchus quadracostatus*: Johnson (1970).

STOP 2. GENESEO, LODI LIMESTONE, SHERBURNE SECTION, HUBBARD QUARRY, (FIG.10).

The contact between the GENESEO and SHERBURNE MEMBERS is well exposed and accessible in the west wall of the quarry. The limey bands in the lower SHERBURNE are the LODI LIMESTONE, a horizon that defines the GENESEO/SHERBURNE contact.

The upper 12 feet of the GENESEO MEMBER at this locality are the HUBBARD QUARRY beds of Brett and Baird (personal communication, 1985), at the base of which is the FIR TREE Horizon, which here is a bedding plane of pyrite and inarticulate debris exposed at the back of the quarry.

The septarian concretions in the upper black shales of the GENESEO are mostly unfossiliferous but they contain "an interesting variety of minerals: barite, calcite, ankerite, quartz, marcasite, sphalerite, and galena, in order of decreasing abundance." (Cole and others, 1959).

The epiplanktic fauna of the GENESEO near the contact (*Styliolina*-brachiopod facies of Thayer, 1974) includes: *Styliolina* sp., *Barroisella spatulata*, *Orbiculoidea lodensis*, *Schizobolus truncatus*, *Leiorhynchus quadracostatus*, *Pterochaenia fragilis*, *Ponticeras perlatum*, *Epitornoceras peracutum*, fish and plant fragments. The clusters of *Leiorhynchus* are of special interest because the presence of this western North American genus is evidence of eastward faunal migration during the Taghanic Onlap (Johnson, 1970)

In the siltstones of the lower SHERBURNE MEMBER (and especially in the LODI horizon) the fauna is a mixed assemblage of the *Styliolina*-brachiopod facies and the benthic *Cladochonus* fauna of Thayer (1974), *Cladochonus* sp., *Leiorhynchus quadracostatus*, *Loxonema noe*, *Palaeotrochus praecursor*, *Panenka* sp., breviconic nautiloids and *Ponticeras perlatum* occur (Cole and others, 1959). This fauna is well developed at Seneca Lake and has been found in the LODI as far west as the Genesee Valley (Figs. 4,5). Presumably the LODI represents a period of reduced clastic influx or hiatus at the end of a transgressive phase. Cole and others (1959) correctly predicted that the LODI would correlate to the west below the level of the GENUNDEWA (*Styliolina*) LIMESTONE, the horizon with which it had been correlated earlier.

The goniatite *Ponticeras perlatum* enters near the top of the GENESEO and ranges through the SHERBURNE to the base of the RENWICK. Specimens may be seen in the black shales around the upper septarian horizon, where they occur with the large involute tomoceratid *Epitornoceras peracutum*, and in the LODI horizon. In Europe these are *Pharciceras* Stufa genera. Huddle (1981) reported conodonts from the LODI at this locality but the species were not zonally diagnostic. At Seneca Lake the GENESEO/SHERBURNE contact corresponds to the *disparilis*/Lowermost *asymetrica* conodont Zone boundary, of the upper Middle Devonian (as defined by the Devonian Subcommittee)(Fig.5).

STOP.3. FAYETTE TOWN QUARRY, CENTERFIELD LIMESTONE AND LEVANNA SHALE, FIG.11.

This quarry, operated by the town for road material, exposes the upper part of the LEVANNA MEMBER of the SKANEATELES FORMATION and the lower 20 feet of the overlying CENTERFIELD MEMBER of the LUDLOWVILLE FORMATION. Gray (1985) has carefully mapped the CENTERFIELD and subdivided it into several submembers based on detailed stratigraphic tracing of distinctive units. He has named the lower CENTERFIELD member, from Buffalo Creek to Owasco Lake, the YORK SUBMEMBER. The quarry is a good collecting site for the dark shale *Leiorhynchus* fauna of the LEVANNA and the more normal bottom *Tropidoleptus* (Hamilton) fauna of the CENTERFIELD.

The LEVANNA is a dark, almost black shale with a typical dark shale fauna of mostly *Leiorhynchus*, *Ambocoelia* and chonetids. Common forms include: *Leiorhynchus multicostus*, *Ambocoelia umbonata*, *Strophalosia truncata*, *Chonetes scitula*, *Orbiculoidea media*, *Nuculites*, *Styliolina fissurella*, *Tornoceras uniangulare*, *Lyrioceras*, *Palaeoneilo*, *Buchiola halli*, *Panenka*, *Pterochaenia fragilis*, *Bucanopsis leda*, *Euryzone rugulata*, *Geisonoceroidea*, *Michelinoceras* and *Spyroceras*.

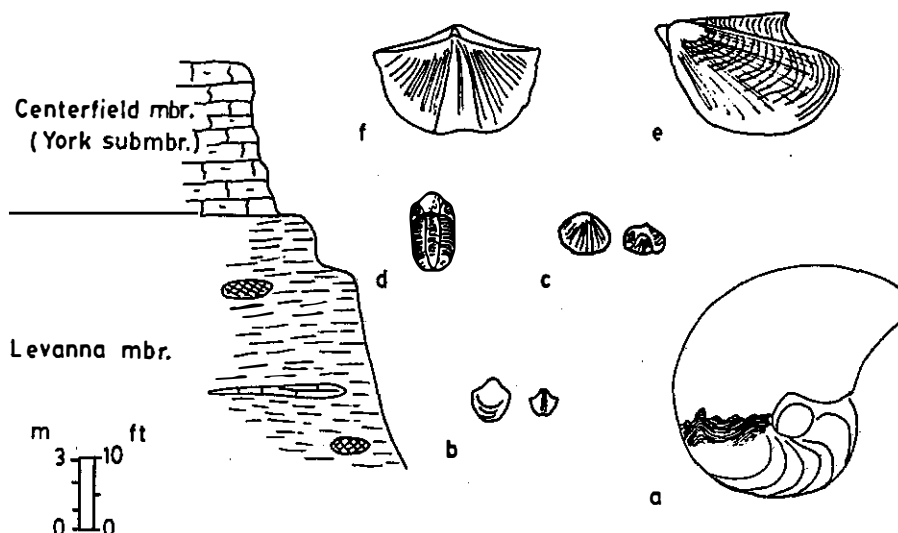


Fig.11. Section in Fayette Town Quarry. Modified after Cole and others (1959). Fossil sketches: a. *Agoniatites vanuxemi*, b. *Ambocoelia umbonata*, c. *Leiorhynchus multicosus*, d. *Phacops rana*, e. *Actinopteria decussata*, f. *Spinocyrtia granulosa*.

Terrestrial plants from lands to the east drifted into this area and their carbonized remains are not uncommon: *Drepanophycus* (lycopsid), *Hostimella*, *Loganiella* (psilopsids), and macerated pieces of the seed-fern *Eospermatopteris*. Fish remains, especially fragments of the armor plate of *Dinichthys halmodeus*, have been found.

Femow (1961) suggested that the LEVANNA was deposited in a relatively high stressed, low energy environment such as relatively deep dysaerobic water and, along with the UNION SPRINGS, OATKA CREEK and LEDYARD Members, may represent the deepest water conditions that existed in Hamilton time along the Cayuga Lake meridian. This is indicated by the fine, dark, carbonaceous aspect of the shale and by the paucity of a well developed, high diversity benthonic assemblage of larger, epifaunal invertebrate taxa.

In the lighter-colored, slightly calcareous, worm-riddled bands near the top of the LEVANNA, a more normal benthonic fauna occurs with *Phacops rana*, *Mucrospirifer*, *Mediospirifer* and chonetids. Large specimens of *Agoniatites vanuxemi* has also been found in these layers.

The limy bands record episodes of more aerobic water, thereby allowing the establishment of a more normal benthonic assemblage. Although free oxygen content was higher than the enclosing dark shales, it probably was not as high as that which prevailed during CENTERFIELD time.

The lighter-colored, slightly calcareous, drab-weathering YORK SUBMEMBER of the CENTERFIELD in the upper part of the quarry represents the lower part of the member, and is here rich in pelecypods such as *Aviculopecten princeps*, *Actinopteria decussata*, *Modiomorpha mytiloides*, and *Leiopteria*; and well-preserved brachiopods: *Meristella barrisi*, *Tropidoleptus carinatus*, *Spinocyrtia granulosa*, *Fimbrispirifer venustus*, and *Mediospirifer audaculus*. A few favositid and rugose corals may be found at or near the top of the unit.

The CENTERFIELD at this locality represents the start of a shallowing upward sequence that does not become fully developed because the section is incomplete. The shallowest facies (peak regression or core CENTERFIELD) at other localities in the Cayuga Lake region is found in the overlying VARICK SUBMEMBER (Gray, 1985). Above this the WHEELER GULLY SUBMEMBER represents a return to

slightly deeper water conditions paralleling those that existed in the YORK SUBMEMBER, as discussed below. The overlying LEDYARD MEMBER above the CENTERFIELD mirrors the LEVANNA MEMBER.

The YORK SUBMEMBER represents a transitional environment between the deepest water LEVANNA below and the shallowest water VARICK above. Generally low energy conditions prevailed as indicated by the fine grained lithology. The diverse benthonic assemblage suggests well oxygenated water. The presence of an occasional trilobite, with well developed eyes, suggests deposition in the photic zone.

STOP 4A. SENECA STONE QUARRY, UPPER SECTION, CHERRY VALLEY/UNION SPRINGS. FIG.12. (Note: Permission must be obtained before visiting this quarry.)

In 1959 only the upper member of the ONONDAGA, the SENECA, was exposed at this location. The quarry had been worked off and on for at least a century, and at that time was being worked on a small scale by Warren Bros. Road Co. to provide crushed stone for bituminous paving. Subsequently purchased by Seneca Stone Corp., the quarry has been greatly enlarged and deepened. The entire ONONDAGA sequence, as well as the overlying UNION SPRINGS SHALE and CHERRY VALLEY LIMESTONE, and the underlying SPRINGVALE HORIZON, ORISKANY SANDSTONE, and uppermost MANLIUS, may now be seen.

Walk west along the upper bedding plane of the CHERRY VALLEY LIMESTONE. This unit has long been famous for its diverse and abundant cephalopods (Clark,1901; Flower,1936; and Rickard, 1952). *Agoniatites vanuxemi* and *Striacoceras typum* are common cephalopods at this locality. Other forms include the tabulate coral, *Aulopoa*, and the trilobite, *Proetus*. Please do not attempt to collect from the bedding surface. Loose blocks at the west end of the outcrop provide better collecting.

To the north a low angle thrust fault may be seen cutting the UNION SPRINGS BLACK SHALE and the upper portion of the ONONDAGA. Several structural depressions may also be seen in the quarry wall. These are sag structures produced by local solution of underlying bedded gypsum. These structures are well developed along the plateau front, near deep valleys where there has been deep ground water circulation (Phillips,1955). They are generally considered to be post-glacial in origin (Gilbert,1891; Fairchild, 1909)

Return to the bus and proceed to the quarry floor.

STOP 4B. SENECA STONE QUARRY, LOWER SECTION, ONONDAGA/ORISKANY/MANLIUS SECTION, FIG.12.

As we pass beneath the crushing plant the TIOGA METABENTONITE BED can be readily see. In 1959 this horizon was approximately one foot above the quarry floor. The TIOGA METABENTONITE provides an excellent time marker. Eastward from here it is found progressively nearer the top of the ONONDAGA, indicating the time-transgressive nature of the unit -- older in the east, younger in the west.

The floor of the main quarry is on the ORISKANY SANDSTONE and the overlying lowermost EDGECLIFF MEMBER of the ONONDAGA. The nodular phosphatic SPRINGVALE HORIZON may be observed, as may the uppermost beds of the MANLIUS FORMATION. The complete EDGECLIFF, NEDROW, MOOREHOUSE and SENECA MEMBERS of the ONONDAGA are exposed in the highwall. In the quarry floor loose blocks of EDGECLIFF and ORISKANY provide excellent collecting. Particularly abundant in the ORISKANY are the large brachiopods *Costispirifer arenosus*, *Acrospirifer purchisoni*, *Rensselaeria ovoides*, and *Hipparionyx proximus*.

To the west in Ontario the SPRINGVALE consists of a sandstone lithologically very similar to the ORISKANY (from which it was derived by reworking) and to the basal "Zone A" sandstone of the EDGE - CLIFF. No sandstone of SPRINGVALE age is present in New York (Hodgson, 1970). However, a nodular phosphatic horizon, representing very slow rates of sedimentation, is well developed here and at Yawgers Woods (Stop 5). Depending on the locality, this horizon rests either on the ORISKANY or on older Silurian dolomites.

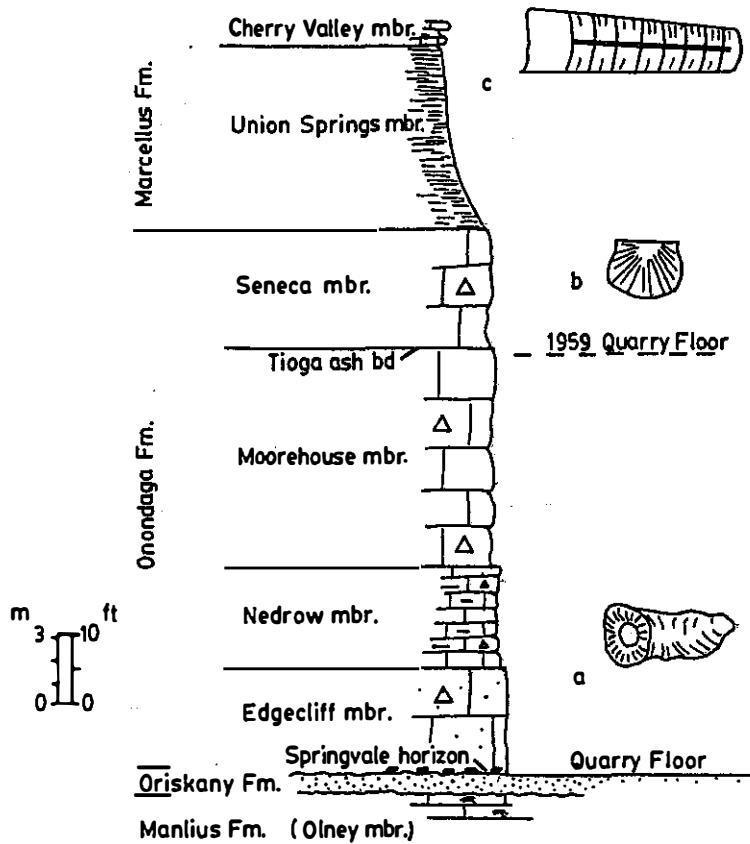


Fig.12. Section at Seneca Stone Quarry, Canoga. Modified after Cole and others (1959). Fossil sketches: a. *Amplexaphyllum*, b. *Chonetes lineatus*, c. *Striacoceras typum*. For additional illustrations see Figs. 11,13.

STOP 5. YAWGERS WOODS, MANLIUS/ORISKANY/ONONDAGA SECTION, FIG.13.

Here the resistant ORISKANY SANDSTONE, a single bed 4 feet thick, outcrops in the woods a half mile west of the road, where it forms a low escarpment facing west. A few feet of MANLIUS (ONLEY MEMBER) LIMESTONE limestone can be seen below it and one or two feet of the basal ONONDAGA (SPRINGVALE and EDGECLIFF) LIMESTONE rest on top.

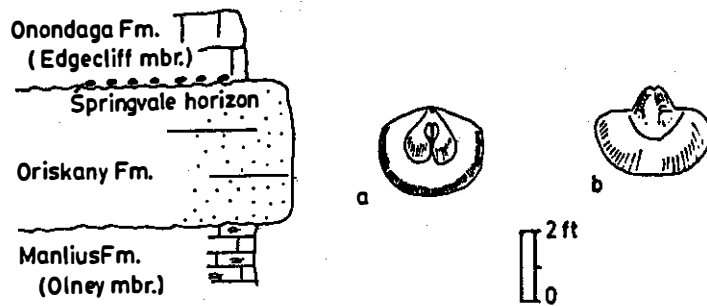


Fig. 13. Section at Yawgers Woods. Modified after Cole and others (1959). Fossil sketches: a. *Hipparionyx proximus*, interior, b. *Costispirifer arenosus*, interior.

In a creek less than 1000 yards to the south the ORISKANY is absent and the ONONDAGA lies directly and disconformably on the MANLIUS. In 1959, the ORISKANY was thought to be absent from the opposite side of the lake westward to Buffalo. Its distribution can now be extended west to the Canoga exposure (Stop 4). The sandy SPRINGVALE horizon in the base of the ONONDAGA has often been mistaken for the ORISKANY in the areas where the latter actually is lacking.

This and a nearby locality are among the oldest known and most famous of American Devonian fossil localities. It was visited as early as 1810 by DeWitt Clinton, later governor of the state, on his return to Albany from an expedition to explore the route of the then proposed Erie Canal (1817-1825). The quote below is extracted from Clinton's "Private Canal Journal" found in Campbell (1849, p. 167-168).

"August 12th., Sunday. We left Ithaca at five..... We dined at Henry Moore's tavern four miles from Cayuga Lake..... Moore is a Republican, as all emigrants from Suffolk county are...

About half-a-mile from his house, and three and-a-half from Cayuga Lake, there is on Lot 69 of the Cayuga Reservation, containing 240 acres and owned by him a ledge of rocks and stones extending a mile in a parallel direction with the lake. The higher stratum is composed of limestone (Onondaga), and the next adjoining one of sandstone (Oriskany) embedded with marine substances. There is but one stratum of sandstone, of the thickness of two or three feet, and below and beneath as well as above it there is limestone (Lower ls. = Manlius). The sandstone contains several marine shells, which appear to be strange, and I should therefore pronounce them oceanic. There are littoral ones also, such as scallops (probably *Costispirifer arenosus*) and in one instance a periwinkle (*Platyceras?*) was found and sent to Peale's Museum in Philadelphia. One strange substance is larger than a scallop, and one is like a horse shoe in miniature (*Hipparionyx proximus*)... This collection of sandstone demonstrates the existence of the ocean here."

A few years later (1815), Clinton remarked of these fossils: "These petrifications are worthy of a more minute examination. I have no doubt but that a very interesting set of shells might be made from this immense stratum of sandstone."

In 1819 David Thomas of Aurora noted that the fossils in the sandstone are mostly in the bottom of the bed, due to the "shells sinking more speedily than the sand" in the Flood, i.e., diastatic settling of the sedimentologists!

Benjamin Silliman sent some fossils from Yawgers Woods to Alexandre Brongniart in Paris in 1820. Brongniart was unable to name them: "...the sandstones of Cayuga, containing terebratulids which I shall perhaps be able, at some future time, to give you the exact name." By 1829, however, Brongniart was able to correlate, more or less correctly, "le gres blanc de Cayuga:" with sandstones of Devonian age in Europe, but it was many years before Conrad, Hall, and Vanuxem figured and described the fossils of the ORISKANY.

Fossils are scarce in the MANLIUS at this locality. The fauna is small, consisting of benthonic forms indicative of hypersaline conditions: *Howellella vanuxemi*, *Brachyprion varistriata*, *Schuchertella interstriata*, *Tentaculites*, *Leperditia alta*.

The fauna of the ORISKANY is characterized especially by large brachiopods in enormous numbers, with occasional pelecypods and gastropods but few other forms: *Costispirifer arenosus*, *Rensselaeria ovoides*, *Acrospirifer murchisoni*, *Hipparionyx proximus*, *Cosstellirostra peculiaris*.

Only the lowest one or two feet of the ONONDAGA LIMESTONE cap the ORISKANY at Yawgers Woods, representing the basal SPRINGVALE sand horizon (Zone A of Oliver, 1954), and the lowest beds of the EDGECLIFF MEMBER (Zone C of Oliver). Collecting is poor, but some loose blocks of the SPRINGVALE in the field at the east edge of the woods contain corals and the black, phosphatic sandy nodules characteristic of the SPRINGVALE. Purple fluorite is sometimes found in intradissepimental cavities in the corals. The SPRINGVALE represents weathered ORISKANY reworked by the westwardly-transgressing Onondaga sea.

STOP 6. PORTLAND POINT QUARRY. MOSCOW/TULLY/GENESEO SECTION. FIGS.14,15.

(Specific stops within this quarry will depend on the stage of operations at the time of our visit).

This quarry, the only commercial TULLY LIMESTONE mine in New York, contains one of the most fossiliferous exposures in the Cayuga Lake Basin. Over 100 species of corals, bryozoans, crinoids, brachiopods, pelecypods, gastropods, cephalopods, and trilobites, have been collected from the top few feet of the MOSCOW SHALE exposed in the quarry floor.

Until 1948 Penn-Dixie Cement Company operated a large cement plant on the Cayuga Lake shore. Cement was transported by barge up Cayuga Lake to the Erie Canal system. Limestone mined in this quarry was conveyed to the plant by aerial cableway. Since 1961 the mine has been operated by Cayuga Crushed Stone, Inc. Material is utilized for aggregate, concrete, bituminous material and rip rap. The level of mining and extent of operation has changed dramatically since the 1959 NYSGA visit to this quarry. Reserves in the present mine are virtually exhausted. An extension of the mining operation is now being developed which will provide limestone to the southern Finger Lakes for the next 10 to 15 years. Limestone will be mined north of Portland Point Road, and will be transported over the road by conveyor belt to the existing crushing plant (Fig.14).

Of particular significance in this mine is the unusual thickness of glacial overburden and black shale that is economically stripped in order to remove limestone. In the southern portion of the mine, over 15 feet of glacial till and 40 feet of GENESEO SHALE must be stripped to remove 10 feet of capstone (a transitional sequence of interbedded limestone and shale - the FILLMORE MEMBER of Heckel (1973) and 19 feet of TULLY LIMESTONE. To control product quality the capstone must be isolated by selective mining. The favorable location of this quarry for marketing stone throughout the Southern Tier makes this extreme depth of stripping economically feasible. Most of the quarry floor is now occupied by spoil piles of GENESEO SHALE. This material will be backfilled against the highwall during reclamation. In the near future only in those areas mined prior to 1975 will bedrock exposure remain.

The quarry is situated on the crest of the Fir Tree (Portland Point) Anticline. The south limb of this fold is cut by a low-angle thrust fault overthrust to the northwest. On the east side of the quarry are several kimberlite dikes. These dikes, described by Sheldon, 1921; Martens, 1924; and Broughton, 1950, vary from 6" to 1'5" in width and strike parallel to the prominent N 5 W joint set. Glacial striae are exposed on TULLY LIMESTONE surfaces at the north end of the quarry.

The black, bituminous GENESEO SHALE is sparsely fossiliferous, but contains a mixture of pelagic marine fossils (*Orbiculoidae*, *Styliolina fissurella*, *linguloids*, *Paracardium*, *Leiopteria*, *Tornoceras uniangulare*), freshwater fish (*Dinichthys*, crossopterygian scales), and land plants (*Aneurophyton*).

The TULLY LIMESTONE is fossiliferous, but fossils are hard to extract. The guide fossil *Hypothyridina venustula* is found in clusters in the lower part of the basal CARPENTER FALLS MEMBER. The thin dark shaly BELLONA CORAL BED, in the upper part of the TULLY, is a widespread datum plane from Skaneateles Lake across the Cayuga Lake Basin nearly to the western limits of the Tully east of Canandaigua Lake. The corals of this zone--*Heliophyllum*, *Heterophrentis*, *Siphonophrentis*, *Cystiphyllodes*, *Favosites*, and *Alveolites*, are not found in the TULLY outside of this bed, and represent a recurrence of Hamilton forms.

The upper 5 feet or so of the MOSCOW FORMATION (WINDOM MEMBER) exposed in the quarry floor, is very fossiliferous. The fauna is the typical normal shale *Tropidoleptus* fauna. Especially abundant or characteristic are: Corals - *Amplexiphyllum hamiltoniae*, *Bethanyphyllum robustum*, *Cystiphyllodes americanum*, *Eridophyllum archiaci*, *Favosites hamiltoniae*, *Favosites turbinata*, *Favosites arbuscula*, *Heliophyllum halli*, *Heterophrentis* spp., *Stereolasma recta*, *Stewartophyllum intermittens*, *Trachypora vermiculosa* (= *T. romingeri*, *T. limbata*); Bryozoa - *Fenestrellina* spp. *Fistulipora fruticosa*, *Fistulipora frurcata*, *Ptilopora striata*, *Polypora multiplex*, *Sulcoretopora incisurata*, *Taeniopora exigua*; Brachiopods - *Athyris spiriferoides*, *Atrypa "reticularis"* *Mediospirifer audaculus*, *Camarotoechia sappho*, *Chonetes coronatus*, *Cryptonella planirostra*, *Cyrtina hamiltonensis*, *Douvillina inaequistriata*, *Elytha fimbriata*, *Megastrophia*

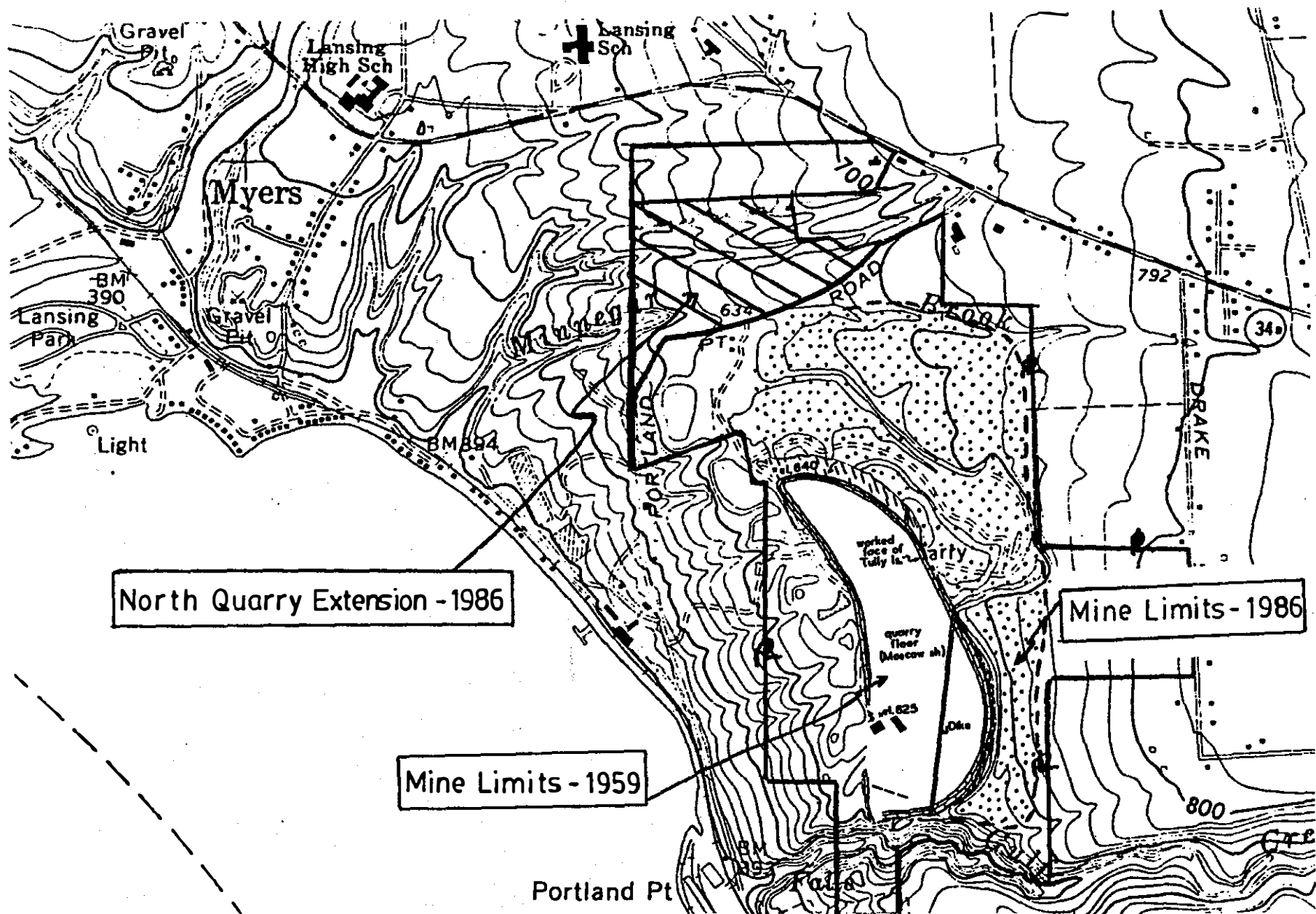


Fig. 14. Location and extent of mining. Cayuga Crushed Stone, Inc. Quarry. Portland Point. Enlarged portion of USGS Ludlowville 7.5 Quadrangle. Modified from Cole and others (1959) and Harrington Associates NYS-DEC and NYS-DOT Open File Reports (1982-86).

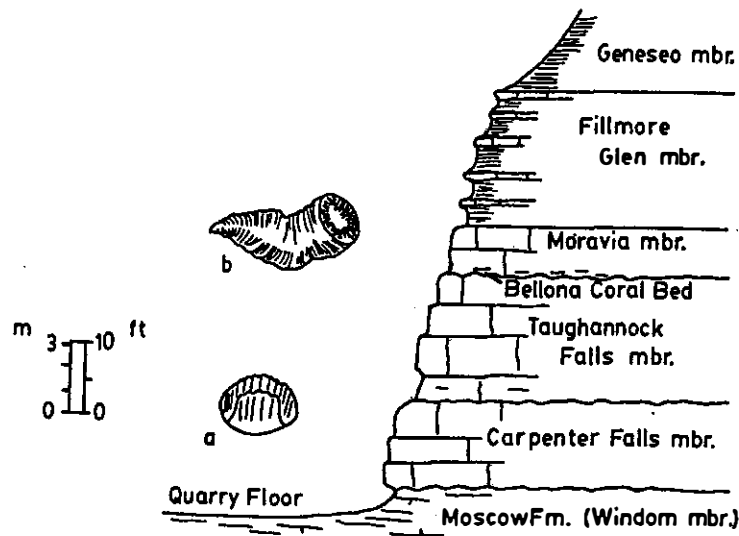


Fig. 15. Section at Cayuga Crushed Stone Quarry, Portland Point. Modified after Cole and others (1959) and Harrington Associates NYS-DOT Open File Reports (1984, 1985). Fossil sketches: a. *Hypothyridina venustula*, b. *Heliophyllum halli*, both $\approx 5X$.

concava, *Mucrospirifer mucronatus*, *Pholidostrophia nacrea*, *Protoleptostrophia perplana*, *Rhipidomella vanuxemi*, *Spinocyrtia granulosa*, *Spinocyrtia marcyi*, *Stropheodonta demissa*, *Tropidoleptus carinatus*; Gastropods - *Naticonema lineata*, *Platyceras erectum*; Pelecypods - *Actinopteria boydii*, *Actinopteria decussata*, *Aviculopecten princeps*, *Cypricardella bellistriata*, *Grammysia arcuata*, *Leiopteria greeni*, *Lyriopecten interradiata*, *Modiomorpha mytiloides*, *Goniophora hamiltoniae*, *Palaeoneilo muta*, *Plethomytilus oviforme*; Cephalopods - *Michelinoceras* sp., *Nephriticerina juvenis*, *Spyroceras* sp.; Trilobites - *Dechenella rowi*, *Dipleura dekayi*, *Greennops boothi*, *Phacops rana*.

STOP 7. RENWICK SECTION. STEWART PARK ACCESS TO ROUTE 13N. FIG.16.

A continuous section through the RENWICK MEMBER of the GENESEE FORMATION is accessible on the south side of the access road above Lake Street. The section begins close to the base of the RENWICK but the contact with the underlying SHERBURNE siltstones is not exposed.

In the vicinity of Ithaca, the RENWICK MEMBER, as defined by deWitt and Colton (1978), consists of sparsely fossiliferous brownish-black and dark gray shale interbedded with gray silty shale and siltstones (some with sole marks) and siltstone-filled scour channels. The RENWICK records a transgression between major stages of delta progradation (SHERBURNE and ITHACA).

The well-known WARRENELLA LAEVIS BEDS (Zone) (Williams 1884; Kindle 1906) which characterize the SHERBURNE/RENWICK transition in this area are well displayed here. This emigrant brachiopod from western North America is an important faunal marker around Cayuga Lake but has not been reported farther west in New York. At the foot of Ithaca Falls, in nearby Fall Creek, Williams (1884) and Kindle (1896) describe a diverse fauna from the *laevis* interval, including the goniatite *Ponticeras perlatum* (highest occurrence of the species) but the fauna here is rather sparse and goniatites have not been seen.

Above the *laevis* beds are a series of transgressive/regressive cycles marked by black shale band interbedded with siltstones, some with sole markings. The distinctive interval of scour channels (channel-fillings). Interpreted by Williams (1881) as iceberg scratchings, is well displayed. de Witt and Colton report channels in this region up to 3 feet deep and 30 feet wide. Scant data indicate southwest trends for the channels and current flow from east to west but with some eastward flow as well (de Witt and Colton 1978, p. A12).

The scour channel interval contains the famous *Plumalina* beds with the delicate fern-like fronds "called *Lycopodites vanuxemi* by Dawson, but considered to be allied with the graptolites and named *Plumalina plumaria* by Hall..." (Williams, 1881). They are now thought to be either hydroid corals or alcyonarian corals (Sass and Rock, 1975).

At the top of the section are dark shales with *Lingula* and *Leiorhynchus*. In the early classifications of the section around Ithaca the dark *Lingula* shales of the ITHACA beds included all of what is now called

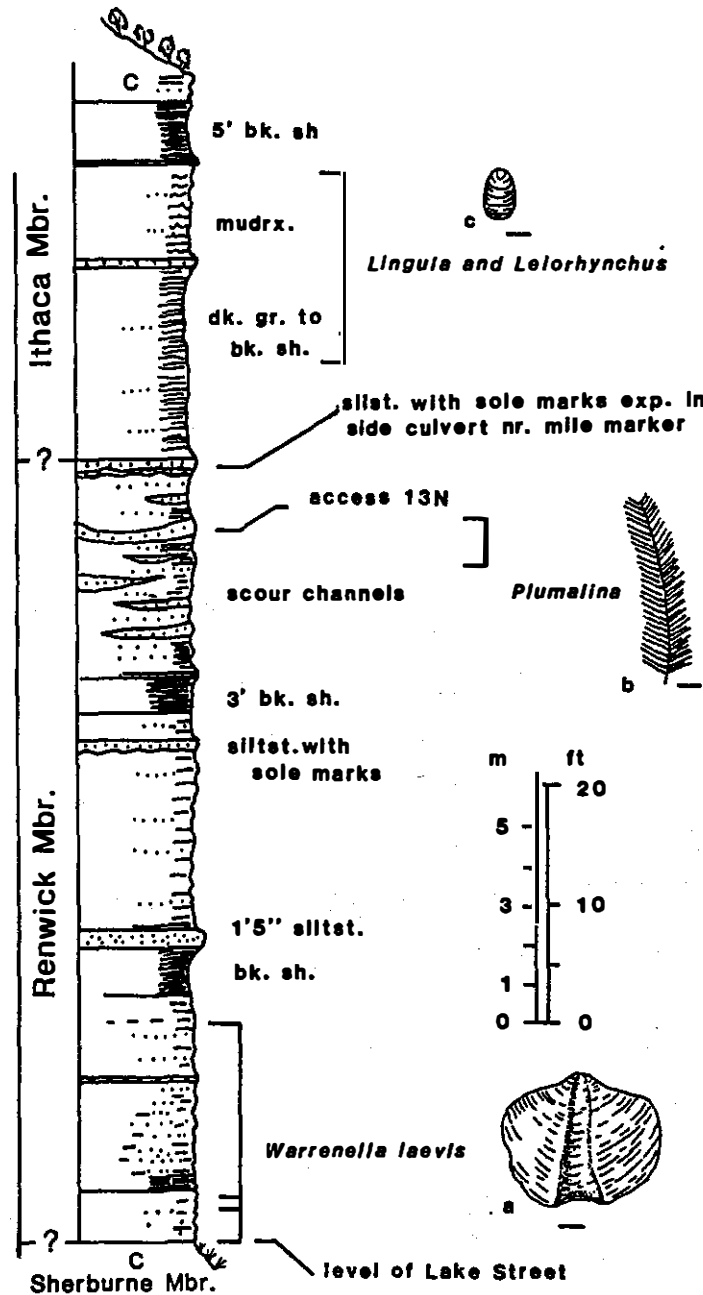


Fig. 16. Section in lower Renwick Member along Stewart Park access to Rte 13N, above Lake Street intersection, Ithaca, N.Y. Fossil sketches based on Palmer and Brann (1966): a. *Warrenella laevis*, b. *Plumalina plumaria*, c. *Lingula complanata*

RENEWICK. The RENEWICK/ITHACA boundary remains poorly defined. deWitt and Colton recognize 60 feet at the nearby type locality at Renwick Brook, which would place the contact near the top of the scour channel interval.

Exposures farther east along Rte. 13 are in the ITHACA MEMBER. The section around the cloverleaf at the Cayuga Heights exit has been examined but no lithologic or faunal horizons have been found which can be correlated to particular horizons in the gorge sections at Fall Creek, Cascadilla Creek, and farther south. It may be possible to tie the sections together by correlating conodont horizons, if Klapper's subdivision of the Lower *asymmetricus* Zone can be recognized. Further work on the paleoecology (Thayer, 1974) and biostratigraphy (Kirchgasser, 1985) of ITHACA MEMBER must include taxonomic revision of the faunas, particularly the brachiopods.

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ROAD LOG FOR TRIP AROUND CAYUGA LAKE

CUMULATIVE MILES FROM MILEAGE	LAST POINT	ROUTE DESCRIPTION
0.0		Mileage begins at intersection of Rtes 13, 89, & 96, i.e. at the "Octopus"; turn <u>hard right</u> onto Rte 89 N.
0.5	0.5	Cayuga Inlet on right (home of Cornell Crew).
1.2	0.7	Exposure of Sherburne Formation on left.
4.3	3.1	Glenwood
5.5	1.2	View across Cayuga Lake at Portland Point (Firtree Point) Anticline.
8.8	3.3	Willow Creek. At this locality deWitt and Colton (1978) measures 11 feet of Genesee, 88 feet of Penn Yan, 78 feet of Sherburne and 8 feet of Renwick. The Tully Limestone forms a waterfall just east of the road.
9.1	3.6	Taughannock Falls State Park. Park in small lot on left (before bridge). Follow path to stream.

STOP 1A TAUGHANNOCK FALLS STATE PARK, LOWER FALLS-MOSCOW SHALE/TULLY LIMESTONE/GENESEO SHALE.

		Return to bus. Proceed north on Rte 89. Note modern delta being deposited at the south of Trumansburg Creek.
9.6	0.5	Bear left onto Taughannock Falls Road. Follow signs to Falls overlook.
9.8	0.2	Travelling over a series of hanging deltas.
10.3	0.5	Turn left into parking lot at Falls Overlook. (Restrooms are available at the east end of the parking lot).

STOP 1B TAUGHANNOCK FALLS STATE PARK, FALLS OVERLOOK, GENESEE FORMATION.

		Return to bus. Turn left and proceed west on Taughannock Falls Road.
11.	0.7	Turn right. Follow sign to Rte 96 (Taughannock Park Road). Bear left along the west side of the stream. The Ithaca Formation is exposed in the stream bed. Note the mature nature of the upper portion of Trumansburg Creek.

12.2	1.2	Turn right onto Rte 96.
12.9	0.7	Enter Village of Trumansburg
14.9	2.0	View east of the glaciated upland.
16.5	1.6	Enter Hamlet of Covert.
20.4	3.9	Village of Interlaken.
20.6	0.2	Junction of 96A
20.7	0.1	Turn right onto Detour Rte 89 (Cayuga Street)
22.0	1.3	Turn left onto Rte 89.
22.6	0.6	Park on right side of road

STOP 2 HUBBARD QUARRY, GENESEO/LODI/SHERBURNE.

Return to bus. Proceed north on Rte 89.

27.0	3.6	Groves Creek on right with water falls over Tully Limestone. Note elevation of Tully compared to its elevation at Taughannock State Park. Excellent fossil collecting in the upper Moscow below falls.
33.1	6.1	Turn left onto Ernsberger Road.
35.4	2.3	Turn right at junction of Rte 414.
38.1	2.7	Enter Town of Fayette.
38.7	0.6	Turn left on Poorman Road. Offshore bar of Glacial Lake Dana to the south.
39.0	0.3	Turn left into Fayette Town Quarry.

STOP 3 FAYETTE TOWN QUARRY, CENTERFIELD/LEVANNA.

Return to bus. Turn right and proceed east on Poorman Road.

39.3	0.3	Turn left onto Rte 414.
41.9	2.6	Turn right (east) on to Canoga Road.
42.2	0.3	Bear right on Reed Road.
43.0	0.8	Pass entrance to Seneca Stone Quarry. <u>This will be Stop 4. We will return to this point after lunch.</u>
43.8	0.8	Turn left (north) at T-intersection.
44.1	0.3	Bear right.
44.8	0.7	Take left fork.

- 45.0 0.2 Turn left onto Rte 89 North at Canoga.
- 46.8 1.8 Enter Village of Seneca Falls. To the left was the site of Cayuga Bridge.

Cayuga Bridge crossed Cayuga Lake as part of the Great Western Turnpike before what is now US Route 20 was built skirting the north end of the lake. Cayuga Bridge was constructed of wood and joined Bridgeport on the westside of the lake with the village of Cayuga on the east shore. Construction commenced in May 1799 and was completed in September 1800, by the Manhattan Company of New York for \$150,000. Its length was 1 mile, and at the time of its completion it was the longest bridge in North America, and perhaps the world. A traveller remarked in 1800 that only 5 years ago "the Indians possessed the shores of the lake, imbosomed in almost impenetrable woods." The first bridge was built on mud sills, the second on piles, the third and last was erected in 1833. It burned sometime thereafter.

- 48.1 1.3 LUNCH STOP - CAYUGA LAKE STATE PARK
Return to bus. Turn left and proceed south on Rte 89 to Canoga.
- 51.4 3.3 Turn left on Canoga Road.
- 52.4 1.0 Bear left.
- 53.7 1.3 Turn right at intersection onto Reed Road.
- 54.5 0.8 Turn right into Seneca Stone Quarry

STOP 4 - SENECA STONE QUARRY, MANLIUS/ORISKANY/ONONDAGA/UNION SPRINGS/CHERRY VALLEY.

Return to bus. Leave quarry turning left onto Reed Road.

- 57.6 3.1 Return to Canoga and turn right (north) onto Rte 89.
- 63.0 5.4 Driving along top of glacial offshore bar.
- 63.8 0.8 Cross the Cayuga-Seneca Canal.

This canal, extending from the old Erie Canal at Montezuma to Cayuga and Seneca Lakes, was constructed in 1826-28. It was enlarged during 1847 -62. In 1918 it became part of the Barge Canal System by canalization of the Seneca River.

- 65.5 1.7 Turn right (east) onto Rte 20.
- 65.8 0.3 On the right is the Montezuma Wildlife Refuge. This swamp is a remnant of proglacial Lake Iroquois.
- 67.3 1.5 Cross Seneca River.
- 67.4 0.1 Cross line of the old Cayuga-Seneca Canal.
- 67.5 0.1 Turn right (south) onto Rte 90.
- 67.8 0.3 Excellent drumlin on left.
- 69.1 1.3 On right is the bed of the old Cayuga-Seneca canal.

- | | | |
|------|-----|---|
| 70.0 | 0.9 | Enter the Village of Cayuga. |
| 70.5 | 0.5 | Turn left (east) at blinker light (West Genesee St). At the foot of the street to the right was the eastern end of Cayuga Bridge. |
| 71.0 | 0.5 | Passing through drumlin field. |
| 72.9 | 1.9 | Enter the West Auburn Gas Field. |

First drilled in 1959, the productive horizon in this field is the Ordovician Queenston formation. Throughout most of New York State this unit consists of a sequence of shales with very low effective porosity. An exception exists in the Auburn area where fine-grained sandstones possessing sufficient porosities for gas production occupy approximately 40 feet of the upper portion of the formation.

At present, the major producer in the area is Miller Brewing Company with over 50 wells being utilized to provide gas for their bottling plant at Sennet, New York. To the south, stepout wells have been drilled to provide onsite gas at Union Springs and at Aurora.

- | | | |
|------|-----|--|
| 72.9 | 1.0 | Cross railroad tracks at Relius. |
| 73.4 | 0.5 | Turn right (south) onto Oakwood Road. |
| 74.4 | 1.0 | On the east is the "Lot 69" Oriskany/Onondaga locality . See discussion for Stop 5. |
| 75.9 | 1.5 | Enter Oakwood. Quarry in the Onondaga is on the right. At stop sign bear right (west) onto Rte 326 |
| 78.0 | 2.1 | Turn right onto deadend lane (Weed Road). Outcrop is in the woods west of lane. |

STOP 5 YAWGERS WOODS, MANLIUS/ORISKANY/ONANDAGA.

Return to bus. Proceed west on Rte 326.

- | | | |
|------|-----|---|
| 80.1 | 2.1 | Turn left (south) onto Rte 90. |
| 80.3 | 0.2 | Village of Union Springs. On right is the Union Springs High School gas well. This well, a stepout of the Queenston West Auburn Field, provides much of the energy utilized by the school. |
| 82.1 | 1.8 | On the left in the old Union Springs Quarry is exposed the Cherry Valley/Union Springs/Onondaga sequence. Note old quarrymen's houses constructed of Onondaga Limestone on the right. |
| 83.4 | 1.3 | Passing the site of Cayuga Castle, "Goi-o-Gouen", a principal Cayuga Indian village destroyed in the Sullivan Campaign September 23, 1779. This was also the site of an early Jesuit Mission. |
| 85.4 | 2.0 | Enter the Hamlet of Levana. The type exposure of the Levana Shale is in the cliff along the lake shore. |
| 87.3 | 1.9 | Enter Village of Aurora. |
| 87.7 | 0.4 | On the right is the Aurora Inn (1833). |

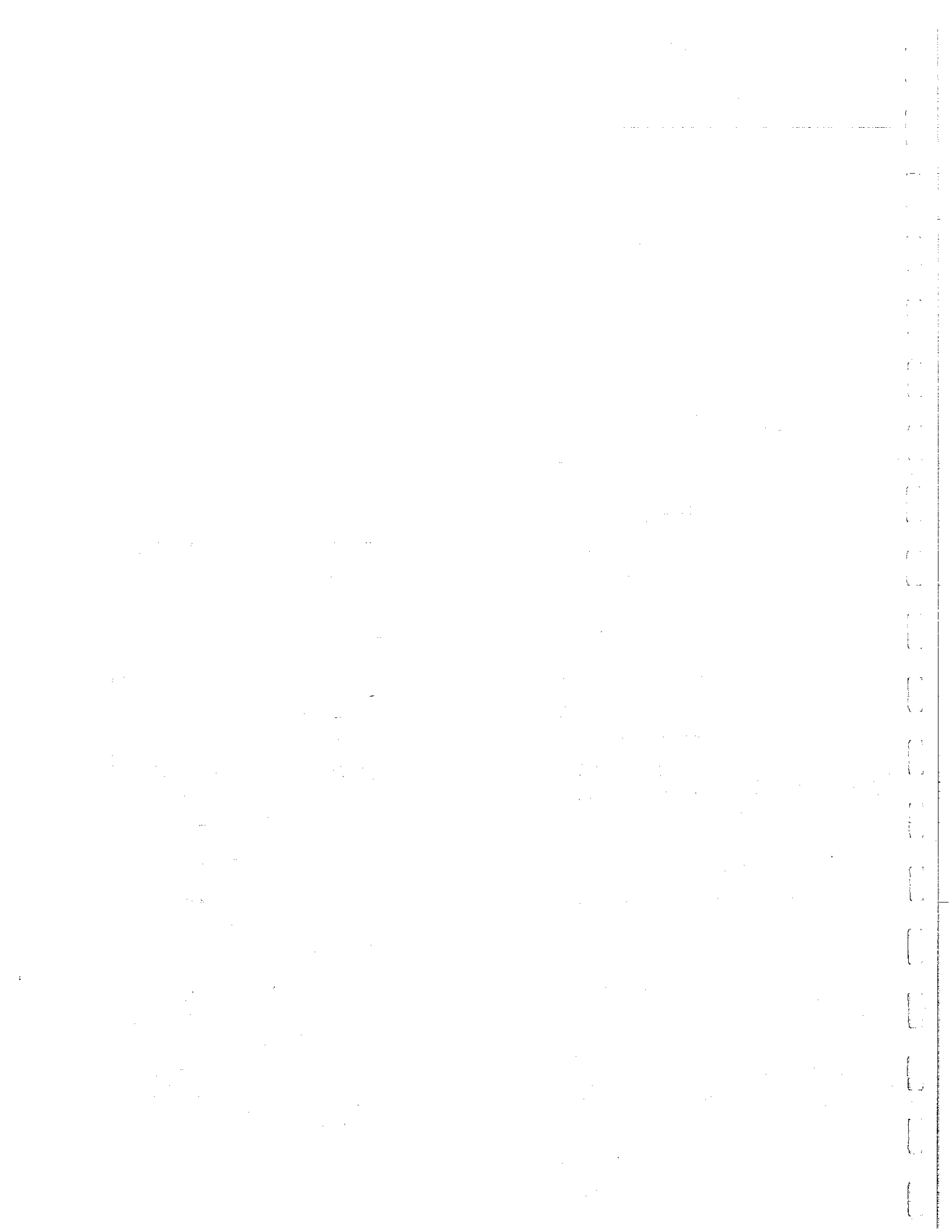
88.3	0.6	Wells College. On the right is Glen Park, built (1852), home of Henry Wells, founder of American Express (1850), Wells Fargo (1852) and Wells College (1868).
89.0	0.7	On left pass Popular Ridge Road , which leads to Moonshine Falls, an excellent exposure of Ledyard Shale and Centerfield Limestone.
95.9	6.9	Enter Village of King Ferry.
96.4	0.5	Turn right (south) onto Rte 34B.
99.2	2.8	Tompkins County Line.
99.9	0.7	Enter Village of Lake Ridge.
105.9	6.0	View south to Ithaca, with the Portage Escarpment in the distance.
106.4	0.5	Cross Salmon Creek, an excellent example of barbed drainage produced by stream diversion.
107.5	1.1	Enter Village of South Lansing
107.7	0.2	Turn right onto Portland Point Road. (at Cargill Salt sign). The salt mine is at the foot of the hill.
108.1	0.4	Turn left into Portland Point Quarry.

STOP 6 PORTLAND POINT QUARRY, MOSCOW/TULLY/GENESO

		Return to bus. Turn right onto Portland Point Road.
108.5	0.4	Turn right onto Rte 34B.
108.9	0.4	Turn right (south) onto Rte 34. On corner is Rogues Harbour Inn (1830).
		Return to Ithaca (6 miles) via Rte 34.
114.9	6.0	Park near the underpass of Rte 13 and walk to roadcut along Stewart Avenue access road to Rte 13N, above Lake Street(Rte 34).

STOP 7.(OPTIONAL), ROUTE 13 ROADCUT, RENWICK/ITHACA.

End of Field Trip



PALEOENVIRONMENTAL ANALYSIS OF THE FIDDLERS GREEN
FORMATION (LATE SILURIAN) IN NEW YORK STATE

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INTRODUCTION

The Late Silurian of New York has long been famous for its eurypterid fauna. A number of investigators have dealt with the occurrence of these unique arthropods such as Clarke and Ruedemann (1912), Ruedemann (1916), Kjellesvig-Waering (1958, 1963, 1964), and Ciurca (1973, 1975, 1978, 1982, 1986). An attempt to determine the habitat of the eurypterids by analyzing their geologic distribution was done by O'Connell (1913, 1916).

Recent works that have helped to refine the stratigraphy are those of Fisher (1960), and Rickard (1962, 1969, 1975). Within the last 15 years more attention has been directed toward the interpretation of environments of deposition. Treesh (1972), Ciurca (1973, 1978, 1982, 1986), Belak (1980), Hamell (1981, 1985, 1986), and Tollerton and Muskatt (1984) have proposed a number of paleoenvironmental models for the Late Silurian rocks of New York.

STRATIGRAPHY

The Late Silurian of New York have been generally described as a transgressive sequence. Examination of lithologies within the Bertie Group, however, indicates multiple oscillations of the Late Cayugan Sea. Lithofacies were deposited in elongate east-west belts arrayed subparallel to the present day outcrop belt. Following Walther's law, the vertical sequence of these facies is inferred to represent lateral north-south shifts of successive environmental belts. The result is a complex package of rocks representing sabkhal to subtidal deposition of carbonate sediments. Several lithofacies are traceable from eastern Ontario, Canada to Cedarville in eastern New York, a distance of approximately 250 miles.

The Bertie Group consists of five formations. In ascending order they are the Fort Hill, Oatka, Fiddlers Green, Scajaquada-Forge Hollow and the Williamsville. Recent terminology for the Bertie Group is shown in Figure 1. In the Syracuse region the Bertie Group has a maximum thickness of 90-100 feet and thins to about 50 feet to the east and west. The increased thickness in central New York is largely due to the occurrence of the gypsum beds of the Forge Hollow Formation. The interval roughly corresponds to the thinner Scajaquada Formation of western New York. Eastward, the entire Bertie Group grades into the gray to green pyritic shales of the Brayman Formation. Late Silurian rocks in western New York are shown in Figure 2.

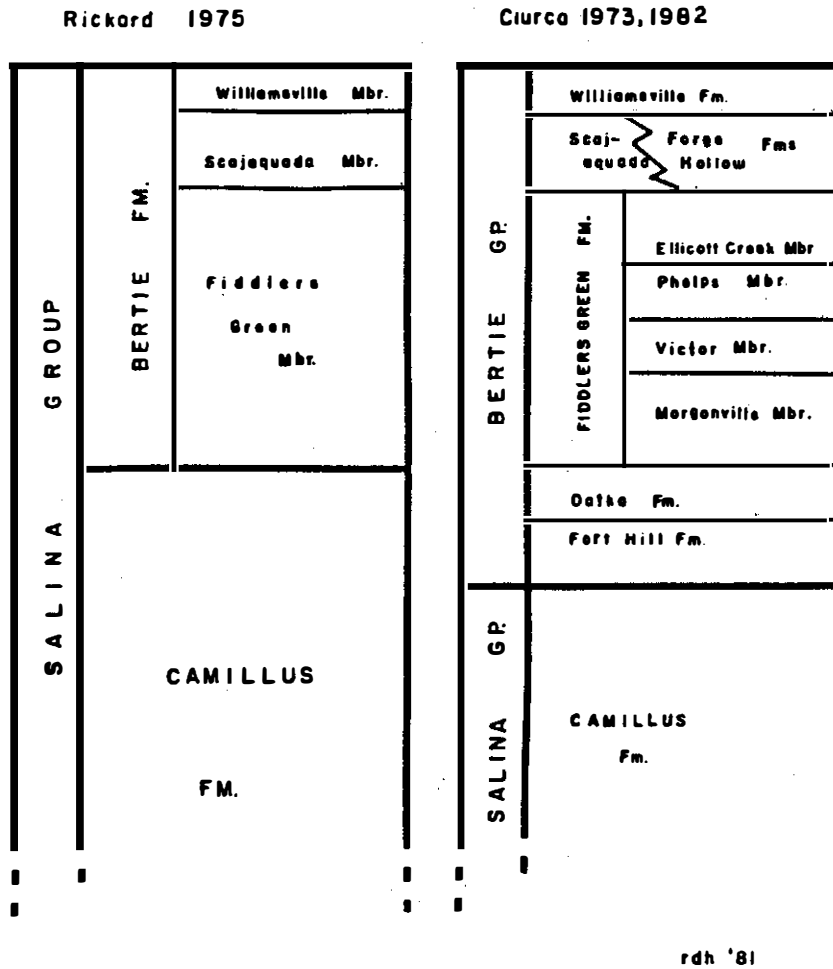


FIGURE 1. Stratigraphic terminology for part of the Late Silurian sequence in New York State.

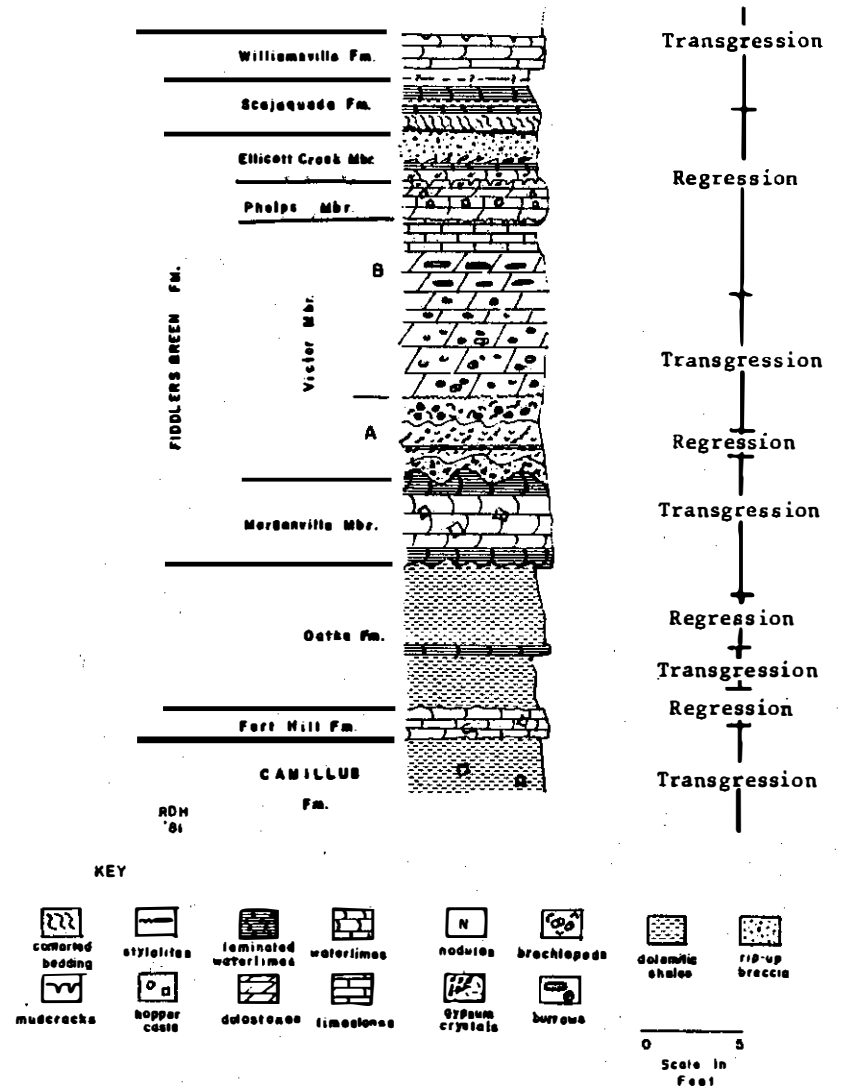


FIGURE 2. Generalized stratigraphic column for western New York State.

FORT HILL FORMATION

The basal one to two feet of the Bertie Group is the Fort Hill Formation. It overlies dolomitic shales of the Camillus Fm. (Salina Gp.) and is a very fine-grained, straticulate dolostone, characterized by small mineraliferous vugs of calcite, large salt hoppers, ostracods, and eurypterid fragments (Ciurca, 1973; p. D-3). The Fort Hill Waterlime is well developed in western New York, but has not been found east of Phelps.

Oatka Formation

The overlying Oatka Formation is about 10 feet thick in central-western New York. These dolomitic shales are easily fragmented into blocky chips having a silty texture. No fossils have been reported from this unit and the lithology represents a recurrence of Camillus-type deposition. Like the Fort Hill Waterlime, the Oatka Fm. is unknown east of Phelps and the interval may be represented by gypsum deposits that underlie the Fiddlers Green Formation in central New York.

Fiddlers Green Formation

The Fiddlers Green Fm. is divided into four members. In ascending order they are the Morganville Waterlime, the Victor Member (dolostones/limestones), the Phelps Waterlime and the Ellicott Creek Breccia. The formation is about 25 feet thick throughout its outcrop belt which extends from the Niagara Peninsula in Ontario, Canada east to Passage Gulf in eastern New York.

Morganville Waterlime Formation

The Morganville Waterlime, the lowest member of the Fiddlers Green Formation, is approximately five feet thick throughout central-western New York. The contact with the underlying Oatka Formation is sharp. The basal Morganville Waterlime is finely laminated and grades upward to a vaguely cross-laminated, thicker bedded, less straticulate dolostone that is typical of this unit. These features can be seen at the roadcuts along N.Y. Route 88 and the adjacent New York State Thruway, both north of Phelps (Stop 1). This locality is the reference section for the Fiddlers Green Formation in western New York.

Fossils are rare in the Morganville Waterlime; only fragmentary remains of Eurypterus have been reported from this unit at Marcellus Falls and Cayuga Junction (Ciurca, 1973, 1978). Salt hopper casts and the ostracod Leperditia have been found at Cayuga Junction and at other localities. The contact with the overlying Victor Mbr. is disconformable being marked by an undulatory surface (Fig. 3). This feature is well-displayed along the roadcuts on the New York Thruway and N.Y. Route 88.



FIGURE 3. Morganville Waterlime (massive unit) overlain by the Victor A Submember of the Fiddlers Green Formation. Note the Aglal (thrombolite?) structure left of the meter stick and the channel. Truncated surface marks the contact of the units.

Victor Member

The Victor Member extends from 2 miles west of Hagersville, in Ontario, Canada east across New York State to Passage Gulf near Cedarville. This unit is a 20 foot sequence of dolostones with some local limestone layers or lenses. The dolostones are coarser grained, usually mottled, possess mineraliferous vugs (Fig. 4), and low diversity fauna consisting of brachiopods (Whitfieldella) and ostracods (Leperditia). Eurypterid remains are rare and have been found only at Morganville and Passage Gulf. The upper portion of the Victor B Submember at Passage Gulf has yielded fragments of Eurypterus and Pterygotus along with gastropods and the inarticulate brachiopods Lingula. The eurypterid fauna of the Victor Member at Passage Gulf is not as rich as the overlying Phelps Waterlime Member. The limestones of the lower Victor A Submember along Route 88 have produced the conodont Spathognathodus remscheidensis remscheidensis, which is correlated with the eosteinhornesis Zone of Europe (Rickard, 1975). Barnett (1971) has suggested that S. remscheidensis ssp. may prove useful for intrabasinal correlation in the Appalachian Basin.

The basal contact of the Victor A Submember with the underlying Morganville is irregular. Cryptalgal structures have been observed draped over the elevated portions of the Morganville (Fig. 3). Burrows and intraclasts are common. This zone is less than one foot thick and grades upward into a 3 foot sequence of sublithographic limestones with abundant lath-shaped gypsum crystals (Fig. 5) up to 5 mm. in length (Hamell, 1981; p. 17). A concentration of these evaporites occurs in the central portion of the Victor A Submember. Upward the gypsum crystals are localized in wisp-shaped lenses together with disarticulated brachiopod shell fragments.

A slight undulatory surface marks the upper contact with the overlying Victor B Submember. This unit is characterized by lithologies, fauna, and sedimentary structures associated with non-restricted subtidal deposition. The basal 1.5 feet is a brachiopod-rich shell hash layer that is highly bioturbated and capped by a well-developed stylolitic horizon. The overlying 11 feet consist of limestone and dolostone beds that are dominated by the brachiopod Whitfieldella. Horizontal burrows are common in the top portion of this unit. A recurrence of Victor A lithology marks the uppermost 2.5 feet of Victor B and a pronounced stylolitic seam marks the contact with the overlying Phelps Member. The aforementioned lithologies of the Victor submembers occur primarily in the centrally located exposures, i.e. the Phelps to Cayuga Lake region.

Phelps Waterlime Member

The Phelps Waterlime Member is fairly uniform in thickness, usually 5 feet, except where truncated by the pre-Onondaga unconformity. It is a very fine-grained, locally straticulate dolostone that exhibits good conchoidal fracture. The top of the Phelps is marked by a mud-cracked horizon that can be traced from central New York (Phelps) to the easternmost locality at Deck (Cieurca, 1978). West of Phelps the mud-cracked horizon is not present. It is either absent due to erosion or is represented only a few inches of waterlime beneath the Ellicott Creek Breccia Member.



FIGURE 4. Mineralized vug in the upper part of Victor A (limestone). Vug diameter about 6 inches.



FIGURE 5. Small crystal molds (gypsum) in lower part of Victor A (limestone). Crystal molds about 0.5 cm. in length.

It is the Phelps Waterlime that has yielded the remarkable eurypterid fauna known as "The Herkimer Pool" for which the Herkimer County area has long been famous. In addition to three genera of eurypterids, the Phelps Waterlime has yielded at least three species of scorpions (Kjellesvig-Waering, 1986) and the aquatic vascular land plant Cooksonia. The latter two forms are found only at the easternmost exposures.

Sedimentary structures such as salt hoppers and salt-reticulate casts characterize the Phelps member, particularly in western-central New York. Specimens 8 inches on a side have been found at the Nied Road Quarry near LeRoy in western New York. To date only a single salt hopper has been reported from the Phelps Waterlime at Passage Gulf. Small scale cross-bedding has been observed at this locality as well as windrows of fossil fragments and complete eurypterid specimens. The Phelps Waterlime at the type section is shown in Figure 6.

Ellicott Creek Waterlime Member

The Ellicott Creek Waterlime is thickest (6-8 ft.) in the Niagara Peninsula of Ontario, Canada, and thins (1-2 ft.) eastward to Phelps, New York (Ciorca and Gartland, 1975, 1976). This unit contains an eurypterid fauna that is similar to the Phelps Waterlime eurypterid assemblage (Ciorca, 1982; p. 114). Chert and sphalerite nodules are found in the lower portion, overlying the mud-crack zone of the Phelps Member. Above this horizon laminae grade from contorted bedding to a zone of rip-up clasts suspended in a micrite matrix. The clasts are tabular to subrounded in shape and weather to a light buff color. Within this zone, sand-sized silica grains occur. Petrographic analysis suggests the intraclasts are partially silicified oolites and the mineralized nodules are replacements of evaporites (Hamell, 1981; p. 22). The contact with the overlying Scajaquada Formation is gradational.

Scajaquada-Forge Hollow Formations

The Forge Hollow Formation consists of approximately 60 feet of gypsum-bearing dolostones in the Syracuse area but thins eastward to about 30 feet near the town of Deck. No fossils have been reported from this interval. The Scajaquada Formation of western New York, usually less than 15 feet in thickness, grades eastward into the Forge Hollow Formation in the Auburn area.

Williamsville Formation

The Williamsville Formation of western New York is characterized by the presence of an Eurypterus fauna (Eurypterus remipes lacustris). Other fossils commonly encountered are the phyllocarid Ceratiocaris, the inarticulate brachiopod Lingula and the graptolite(?) Inocaulis. Cephalopods and gastropods are present but poorly preserved. In central New York the Eurypterus fauna, which is characteristic of western New York, is replaced by a Paracarcinosoma (formerly Eusarcus) fauna as reported by Ciorca (1978, 1982). East of Syracuse this interval is marked by a thin petroliferous waterlime that is unfossiliferous.



FIGURE 6. Upper Victor B Submember (flaggy beds) and Lower Phelps Waterlime Member. Contact just below the meter stick.

Throughout western New York the Williamsville contact with the overlying Cobleskill Formation is gradational. At Forge Hollow the contact is disconformable (Rickard, 1962). The Williamsville Formation grades eastward into the uppermost Brayman Formation.

PALEOENVIRONMENTAL INTERPRETATION

Based upon the geographic and vertical distribution of lithologies, sedimentary structures, petrographic and paleontological observations, eight depositional environments are inferred by comparison with modern depositional environments (Persian Gulf and elsewhere):

1. sabkha
2. lower supratidal
3. upper intertidal to lower intertidal
4. lower intertidal
5. restricted subtidal
6. non-restricted subtidal
7. hypersaline lake
8. semi-restricted lagoon-estuary

The environmental distribution of facies for a portion of Fiddlers Green time is shown in Figure 7. The paralic arrangement of Bertie Group facies is illustrated in Figure 8.

Sabkha

Sabkhal sedimentation is characterized by dolomitic muds and an absence of organisms. During dolomitization shells are usually leached out during flooding. Relict evaporite minerals such as gypsum-anhydrite and halite are generally preserved in sabkha-type sediments. Later diagenesis can replace the evaporite nodules with silica and/or other minerals. The Oatka and Scajaquada Formations are interpreted as representing sabkhal deposition as is the underlying Camillus Formation of the Salina Group.

Hypersaline Lakes

The environmental setting for the deposition of the Forge Hollow gypsum beds is analogous to present day sedimentation of gypsum occurring in shallow hypersaline lakes on Bonaire, in the Northern Antillies as reported by Lucia (1968). These restricted sabkhal-hypersaline lake deposits are completely separated from the ocean by a barrier composed of permeable sabkhal sediments. Precipitation of gypsum is maintained by a continuous influx of seawater through the permeable sediments. The deposits are bedded or laminated and form a thick sequence of evaporites. No organisms have been found associated with these lakes. The thick gypsum beds of the Forge Hollow indicate deposition in a restricted sabkhal-hypersaline lake.

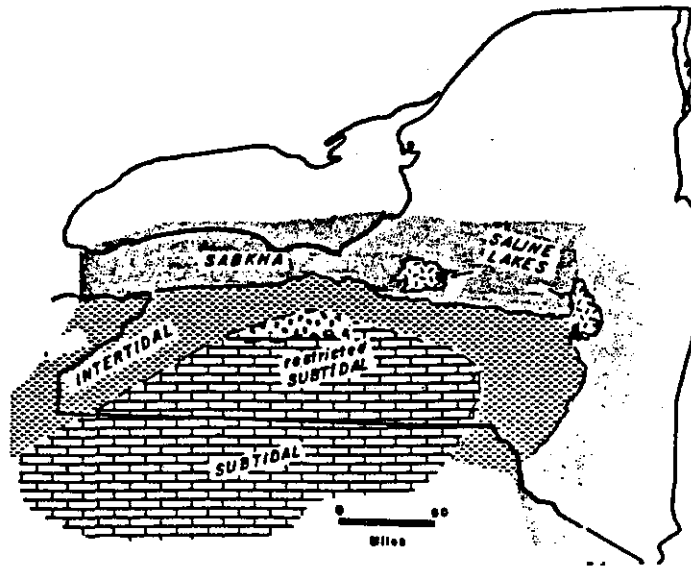


FIGURE 7. Paralic relationship of depositional environments during Fiddlers Green time. (From Hamell, 1981)

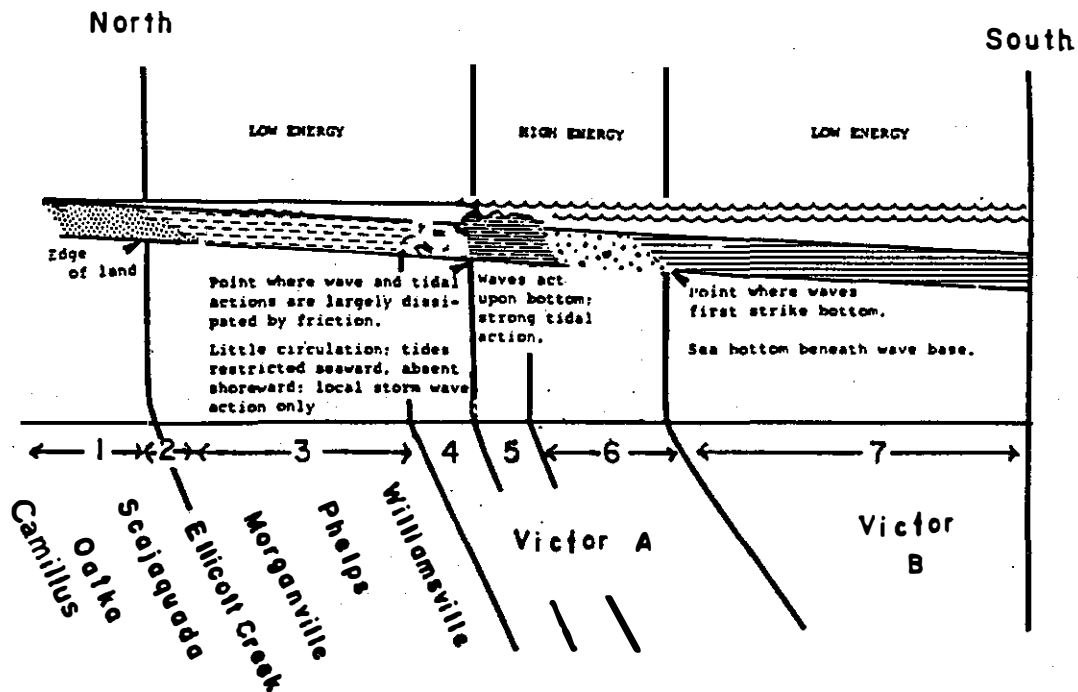


FIGURE 8. Cross section of Late Cayugan lithofacies indicating distribution of Bertie Group lithologies in accordance with Shaw-Irwin model of epeiric sea sedimentation and observations: 1) sabkha, 2) supratidal to intertidal (rip-up breccia), 3) intertidal, 4) restricted subtidal, 5) algal structures on eroded intertidal facies, 6) more normal marine subtidal, above wave base and 7) subtidal, below wave base (From Hamell, 1981).

Supratidal

The collapse and rip-up breccias of the Ellicott Creek Member (Fiddlers Green Fm.) were probably formed in the lower sabkhal to upper supratidal environments. This zone lies above the mean high tide and sedimentation occurs during excessively high spring tides or major storms. In the interim, the environment is subjected to long periods of subaerial exposure and evaporation. Collapse breccias are the result of dissolution of evaporite minerals and subsequent collapse of the sediment during periods of flooding.

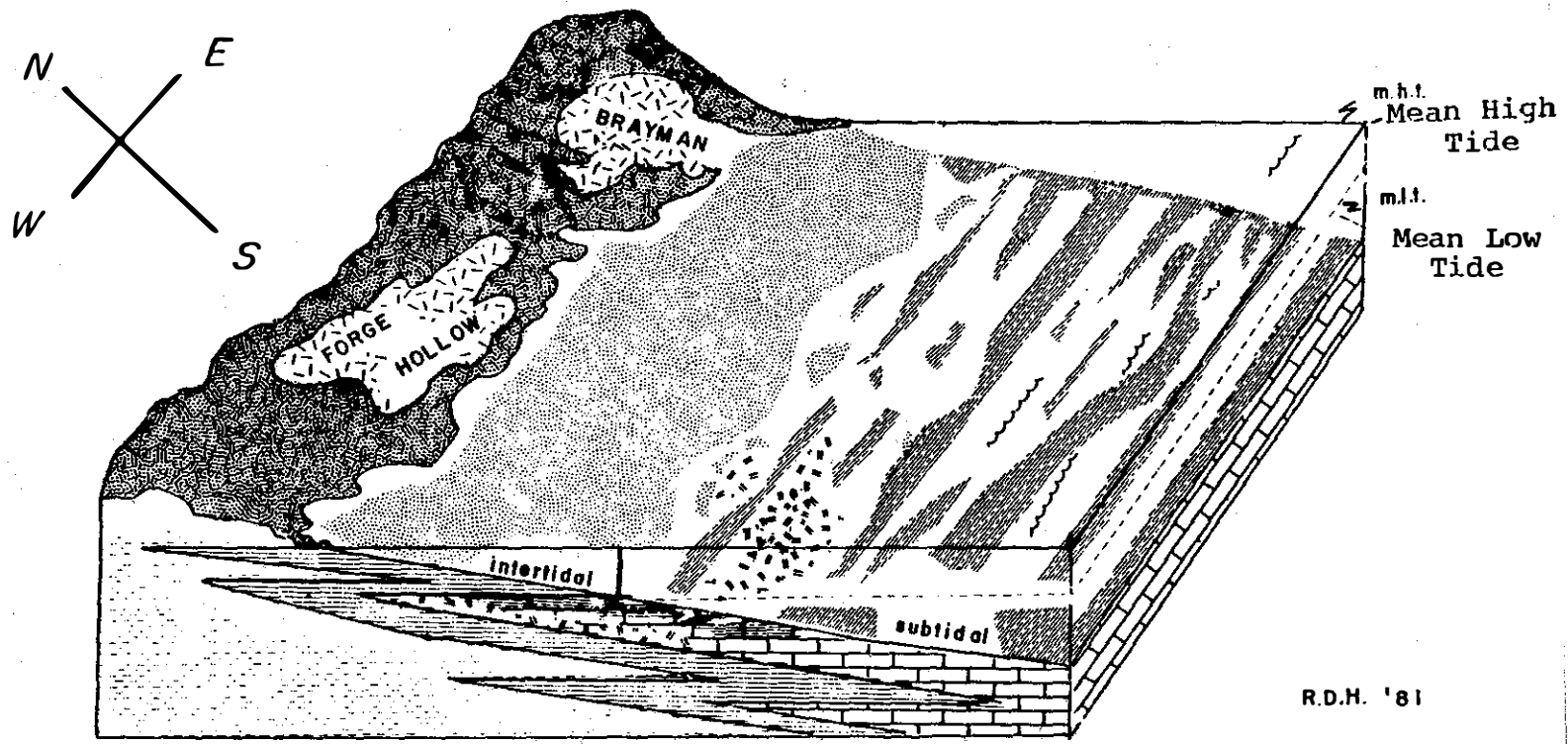
Intertidal

Treesh (1972) and Belak (1980) have implied that the Brayman Shale, Phelps and Williamsville Waterlimes were deposited in an intertidal environment. Their environmental interpretation of these rock units does not provide an adequate explanation for the high pyrite content of the Brayman, which is lacking in the Phelps and Williamsville. Therefore, the Brayman represents a different environmental setting. As noted by Fisher (1960), the slight dissimilarity of the Brayman with the Forge Hollow Formations is due to a slight facies change. The stratigraphic relationship of these two rock units indicates a similar environment of deposition. However, in the case of the Brayman Shale, gypsum must have been inhibited from precipitating or accumulating. Furthermore, the deposition of black organic rich mud was favored in the Brayman environment for there is a lack of significant pyrite in the Forge Hollow gypsum beds. Morris and Dickey (1957) described modern evaporite deposition associated with black muds occurring in a relict river channel in Peru which forms an estuary with an open connection to the Pacific Ocean. Sulfate-reducing bacteria inhibit the accumulation of gypsum in this environment. Based on these lithological similarities, such an environmental setting can be postulated for the deposition of the Brayman Shale Formation.

Intertidal sedimentation is represented by portions of the Fort Hill Formation, the Morganville and Phelps Members of the Fiddlers Green Formation as well as the Williamsville Formation. This environment is characterized by mud-cracks that mark the upward transition into a supratidal environment. Hypersaline conditions are supported by the presence of salt hoppers, reticulate halite structures and laminated sediments. Cryptalgal structures and an eurypterid fauna are the main fossil constituents.

Subtidal

The lithographic limestone in the basal Victor A Submember contains the brachiopod Whitfieldella. Conditions of hypersalinity are indicated by a one foot thick bed of packed gypsum crystals in the central portion of Victor A and in the upper 2.5 feet of the Victor B Submember. These lithologies are indicative of deposition in a restricted upper subtidal environment.









- 
SABKHAL
- 
**HYPERSALINE
'LAKES'**
- 
INTERTIDAL
- 
**UPPER
SUBTIDAL**
- 
**RESTRICTED
SUBTIDAL**
- 
**LOWER
SUBTIDAL**

FIGURE 9. Schematic reconstruction of the environment and deposition of the Bertie Group lithologies.

The primary features of the Victor B Submember are the wavy-laminated to mottled and thick-bedded dolostones having a strong petroliferous odor. The mottled nature of this facies is due to bioturbation and is typical of modern shallow subtidal zones. Complete specimens of Whitfieldella and Leperditia are common. The deeper subtidal facies of the Victor B Submember is recognized by well-developed horizontal burrows. Although this environment represents a more offshore setting, the small size of Whitfieldella, low faunal diversity and the absence of typical marine organisms, are suggestive of less than normal marine conditions.

CYCLICITY AND DEPOSITIONAL HISTORY

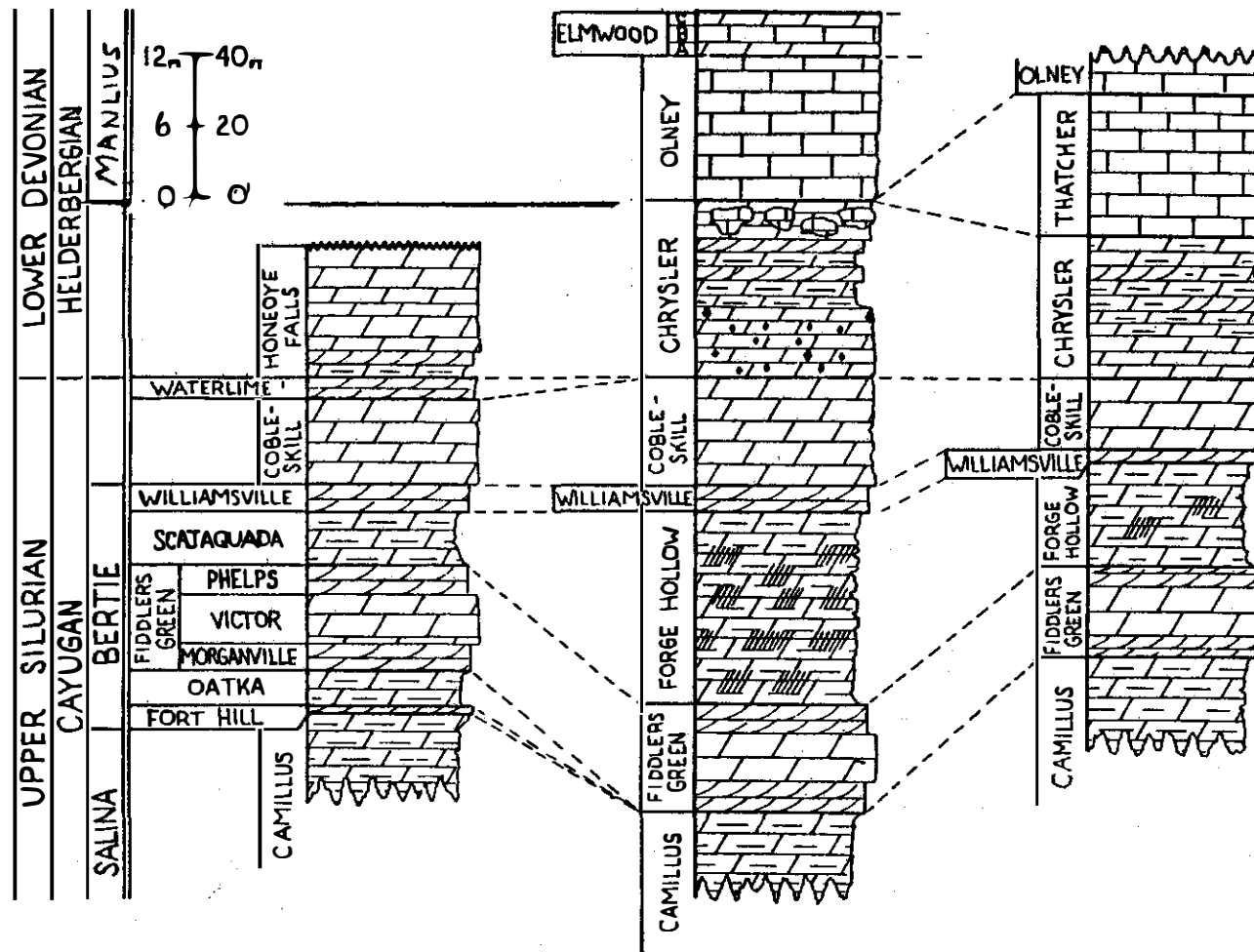
The cyclic nature of the Late Silurian strata in New York was indicated by Cieurca (1978). During deposition of the Bertie Group most of the lithofacies and biofacies were shifted geographically over a broad area. The waterlime facies, for example, disappears from a localized section only to recur repetitively at stratigraphically higher intervals. Concomitantly, the eurypterid faunas that are so well known from this facies also recur but are replaced by newly evolved or introduced species (Table 1).

The cyclic alternation of lithofacies and biofacies reflects the effects of salinity and sedimentation. Waterlimes formed mostly under hypersaline conditions but it is obvious that the thin limestones (Victor A Submember) were deposited in more nearly normal marine conditions. No normal marine limestones like those of the Cobleskill Formation (Stops 2, 3 and 7) are known in the outcropping of the Fiddlers Green Formation. Perhaps subsurface analysis of the Fiddlers Green Fm. is needed, especially to show how this formation merges with basinward deposits.

Whatever the underlying causes were for the repeated transgressions of the latest Silurian and Early Devonian, the stratigraphic column (Fig. 9) reflects the widespread shifting of shore to shallow restricted areas (sabkhal to intertidal) with the more offshore regimes of varying salinities (restricted to non-restricted subtidal).

TABLE 1

<u>LITHOLOGY</u>		<u>FORMATION or MEMBER</u>	<u>FAUNA</u>
shaly dolostone		Scajaquada Formation	barren
waterlime	Fiddlers Green Formation	Phelps Waterlime	eurypterids
crystalline dolostone & limestone		Victor Member	brachiopods, ostracods
waterlime		Morganville Waterlime	eurypterids
shaly dolostone		Oatka Formation	barren



Stratigraphic sections for Syracuse area (center column) and comparable sections to west, south of Rochester (left column), and to east in Stockbridge Valley area (right column).

FIGURE 10

(modified from Ciurca, 1978)

After two major cycles (Bertie through Cobleskill Fms.) most of New York State was overwhelmed by deposition of Helderbergian limestones (Fig. 10). These thick marine deposits were built up by biohermal and biostromal stromatoporoid and coral growth in conjunction with the accumulation of abundant debris of marine organisms. The type section of the Olney Limestone (Stop 4) exhibits some of the sedimentological features characteristic of this major transgression. Helderbergian facies and inferred environments have been discussed by Rickard (1962, p. 93-99; Fig. 26) and by Laporte (1969). See also Ciurca, 1978.

SUMMARY

The Bertie Group was deposited during multiple oscillations of the late Cayugan Sea. This resulted in a cyclic sequence of lithofacies reflecting sabkhal to subtidal depositional belts (Fig. 9). A minor oscillation from sabkhal to intertidal to sabkhal sedimentation is recorded by the upward transition from Camillus-Fort Hill and Oatka Formations.

The Fiddlers Green Formation records the major transgressive-regressive cycle of the Bertie Group. Victor B exhibits the farthest offshore facies of the Fiddlers Green Formation. The facies consists of fine-grained fossiliferous limestone parts of which contain well developed horizontal burrows and mottling due to bioturbation.

Supratidal conditions are indicated by the collapse and rip-up breccias of the Ellicott Creek Member. Supratidal to intertidal sedimentation is displayed by the Morganville and the Phelps Waterlime Members. Key indicators are mudcracks, cryptalgal structures (Fig. 3), and laminated sediments. Rare occurrences of ripplemarks, cross-bedding, and channel structures (Fig. 3) add to the support of this interpretation. The Victor Member, the thickest and most fossiliferous member of the Fiddlers Green Formation, records subtidal deposition under conditions approaching near normal salinity.

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216
ROADLOG

<u>Cumulative Mileage</u>	<u>Miles From Last Point</u>	<u>Route Description</u>
0	0	STOP #1. Roadlog starts at the roadcut along NY 88 near the NY Thruway (I-90) overpass, 0.5 mile north of NY 96 at Phelps.

STOP #1. The Onondaga Ls. is the highest unit present and contains a basal sandy layer (Springvale?) just above the unconformity with the Silurian Scajaquada Fm. Most of the Fiddlers Green Fm. is exposed along the highway. Camillus Fm. (Salina Gp.) is exposed north of the railroad overpass.

0.5	0.5	South to NY 96. Turn left (East).
5.7	5.2	Intersection of NY 14 turn north.
6.3	0.6	Intersection NY 318 turn right (east).
17.3	11.0	Junction U.S. 20. Turn left to Auburn.
19.4	2.1	Junction NY 90. Continue east on U.S. 20.
26.9	7.5	Turn left onto Clark Street.
27.0	0.1	Turn right onto Beech Street.
27.8	0.8	Turn left onto Canoga Junction.
28.3	0.5	Turn right into parking lot (before the bridge).

STOP #2. OWASCO CREEK.

STOP #2. South of the bridge is an exposure of the stromatoporoid biostromal facies of the Cobleskill Fm. (Late Silurian). Downstream are poor exposures of the Fiddlers Green Formation.

28.6	0.3	Continue east on Canoga Junction Road. Turn left onto Wadworth Street.
28.7	0.1	Turn left onto Aurelius.
28.8	0.1	Turn left onto Wall Street.
29.6	0.8	Junction of NY 38. Turn right. Follow signs to U.S. 20 East.
29.9	0.3	Junction of U.S. 20. Turn left (to Skaneateles).
37.3	7.4	Junction of NY 321 North in Skaneateles continue east on NY 20.
39.1	1.8	Junction of NY 175. Turn left (to Marcellus).

- 42.6 3.5 Junction of NY 174. Turn left.
- 44.0 1.4 Turn right then next left on North Street (NY 174).
Head toward Marcellus Falls.
- 45.3 1.3 STOP #3. THE COBLESKILL AND CHRYSLER FMS.

STOP #3. The Cobleskill and Chrysler Fms. are well displayed at this stop. A cherty facies is present in the Cobleskill Fm. and appears to be limited to the Syracuse area. Continue north on NY 174.

- 49.0 3.7 Junction of NY 5 at Camillus. Turn right onto NY
5 (Old Route to Syracuse).
- 52.9 3.9 Junction of NY 173. Turn right.
- 54.2 1.3 Junction of Split Rock Road. Turn right.
- 54.8 0.6 STOP #4. SPLIT ROCK QUARRY

STOP #4. There is a complete section of the Olney Fm. in the quarry. Above the Elmwood A is the Onondaga Ls. The uppermost Chrysler Fm. (E. Dev.) is exposed on the quarry floor near the entrance road.

- 55.4 0.6 Return to Junction NY 173. Turn right.
- 60.4 5.1 U.S. 11 Jct., turn right (So. Salina St.).
- 63.4 3.0 Junction of I-81, turn left (north) on I-81.
- 66.2 2.8 STOP #5. E. DEV. CARBONATES & ANTICLINE.

STOP #5. This E. Dev. carbonate sequence is one of the best exposures of the Manlius Gp. in central New York. Facies changes in the Olney Ls. (compared to the type section at Split Rock) are readily apparent. Stromatoporoid biostromes are common and become more important eastward.

- 66.7 0.5 I-481, Exit 16A north to Dewitt.
- 67.4 0.7 Exit to Rock Cut Road (Exit 1).
- 67.6 0.2 Turn left onto Rock Cut Road.
- 69.9 2.3 At the Stop Sign head south toward Jamesville.
- 70.2 0.3 Turn left. Stop at bridge over Butternut Creek.

STOP #6. TYPE SECTION-FIDDLERS GREEN FM.

STOP #6. Butternut Creek contains fine exposures of this unit. Uppermost Fiddlers Green (mudcrack zone) occurs just above the falls. No Forge Hollow Fm. has been observed here but just up the road (toward the quarry) is a good exposure of the Cobleskill Fm. (celestite facies).

- 70.3 0.1 Return to main road. Turn right.

71.0	0.7	Turn right onto Woodchuck Hill Road.
72.6	1.6	Turn left onto Old Quarry Road.
73.0	0.4	<u>STOP #7.</u> ABANDONED QUARRY-FORGE HOLLOW FM.

STOP #7. The stratigraphic relationships of the Bertie-Cobleskill rocks are clearly shown at this quarry. The Williamsville Fm. has an unique fauna including Paracarcinosoma, Orbiculoidea, and Lingula. Overlying this unit are stromatoporoid beds of the Cobleskill Formation.

END OF ROADLOG

KIMBERLITES OF THE FINGER LAKES REGION

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INTRODUCTION

Phlogopite-bearing ultramafic dikes in the Finger Lakes region were first reported in the literature by L. Vanuxem in 1837. Although not called kimberlites at the time, these were the first kimberlites described in the world (Dawson 1980). Over the years, numerous papers and theses have been written on their location (Table 1), age, field setting and mineralogy, but many questions regarding the dikes remain unanswered. The dikes have generated interest in each generation of geologists because they are the only manifestations of igneous activity in the Mesozoic in the region; and the timing of their intrusion cannot be easily related to the regional tectonics. These dikes and similar intrusions in Pennsylvania and to the south are the only windows into the lower crust and mantle in the Appalachian basin. The existence of such small ultramafic intrusions from deep within the mantle has caused interest in the mechanism of emplacement of these and similar intrusions. Not incidentally, the designation of the Finger Lakes dikes as kimberlites has generated interest in the possibility of diamonds being found in the region.

LOCATION AND AGE OF THE FINGER LAKES DIKES

Eighty-two dikes and two diatremes have been reported from the Central Finger Lakes region south and west of the city of Syracuse. The greatest concentration of dikes occurs in the glacially cut gorges in and immediately north of the city of Ithaca. Most of the localities are shown in Figure 1 and described in Table 1. Localities to be visited on this trip include Cascadilla Gorge (17), Williams Brook (10), Taughannock Creek (5), Poyer Orchard (8) and Portland Point (20).

The dikes are generally vertical and occur in the prominent north-south joint set in the Devonian sedimentary rocks (see stratigraphic section in introductory part of this guide). Most of the dikes are narrow with widths ranging from several centimeters to about 0.75 meter. Exceptions are dikes near Poyer Orchard and in Cascadilla Creek (localities 8 and 17, Table 1) which are close to a meter wide and the Williams Brook dike (locality 10) which is 3.5 meters wide. Even the thin dikes can extend vertically for great distances. For example, a dike that is only 3-4 cm. wide extends almost 100 meters up the gorge wall at Taughannock Falls (Table 1, locality 4; Foster 1970). Evidence for multiple injections in the same dike occurs in dikes at the Portland Point Quarry and at Taughannock Creek (Foster 1970).

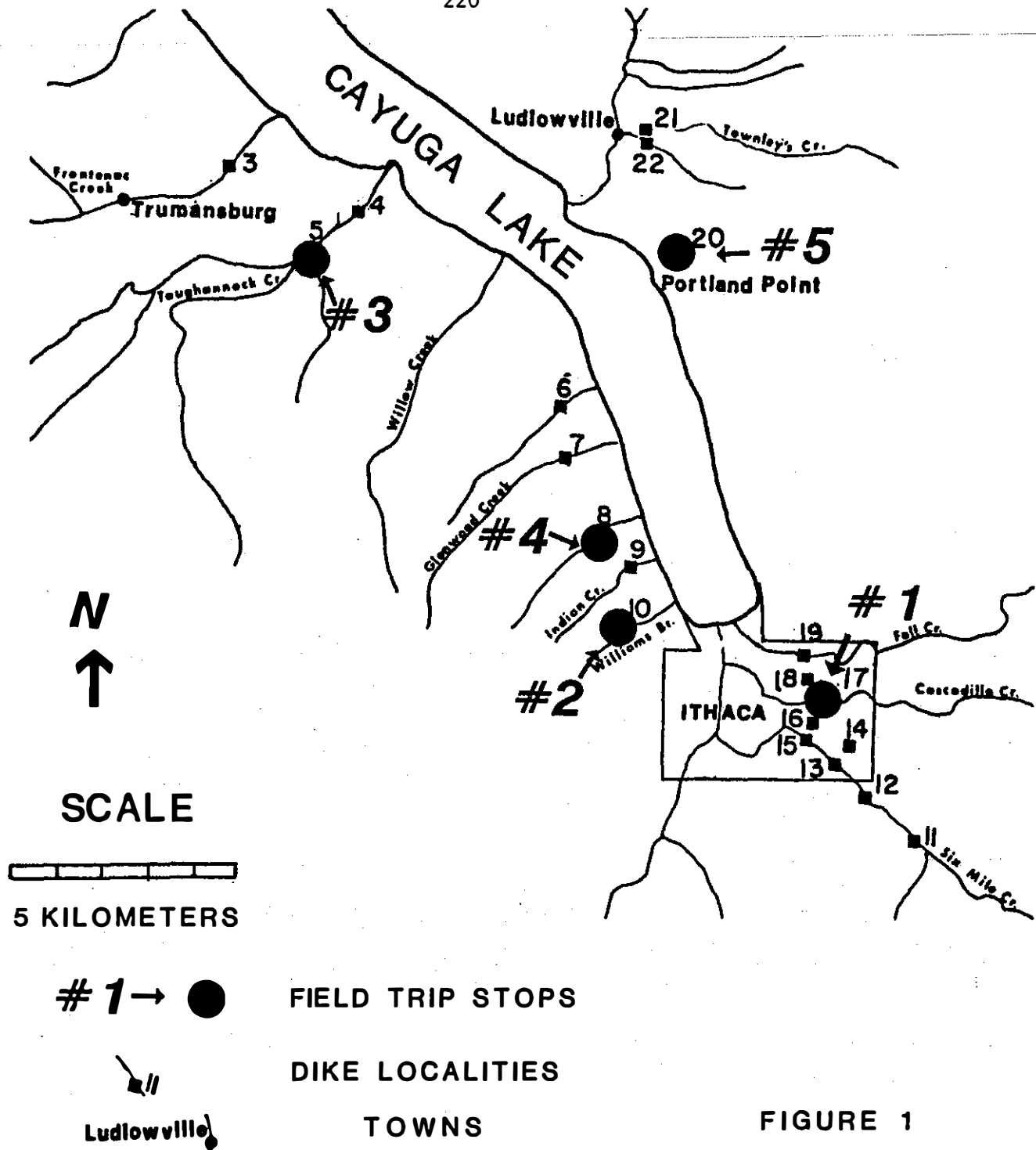


FIGURE 1

Map showing locations of most of the known kimberlite dikes and diatremes outcropping around Cayuga Lake near Ithaca, New York. Dike localities are keyed to localities described in Table 1. Field trip stop numbers refer to localities discussed in this guide. Figure modified from Foster (1970).

TABLE 1 - REPORTED LOCATIONS OF DIKES IN THE CENTRAL FINGER LAKES

(Further information and other references can be found in cited references, table modified from Foster 1970)

- 1 - Small quarry, 400 m. west of intersection of NY 96 and NY 336, one dike in quarry floor, maximum width is 0.4 m., strike N35W, (Wells 1961).
- 2 - Lively Run Creek, 425 m. east of NY 89. Four narrow dikes, 61 m. east of two dikes which are 12 m. apart, maximum width is 5 cm., strike N10W (Martens 1924).
- 3 - Frontenac Creek, 460 m. west of NY 89. Cluster of 3 dikes, the widest is 0.6 m. across. 150 m. upstream to west is a dike, 0.2 m. wide, all strike N10W (Foster 1970).
- 4 - Taughannock Creek, base of falls in Taughannock State Park. Five closely spaced dikes with a maximum width of 5 cm., strike N-S (Martens 1924).
- 5 - Taughannock Creek, 1 km. east NY 96 bridge (Stop 3). Ten dikes in creek over distance of 0.5 km. (Figure 2), maximum width is 0.25 m., strike N10W (Martens 1924, Foster 1970).
- 6 - Unnamed creek north of Glenwood Creek, where creek meets bend in Glenwood Hts. Rd. One dike, 0.45 m. wide (Filmer 1939), not located by Foster (1970).
- 7 - Glenwood Creek, 365 m. east of Duboise Rd. on south side of creek. One dike, 2.4 m. wide, strike N5E (Martens 1924).
- 8 - Ravine, 1.2 kilometers south of Glenwood Creek in Poyer Orchard.
 - a) Diatreme exposed for about 60 m. along ravine, strike N1-5W (Stop 3 and Figure 3).
 - b) Dike, 0.9 m. wide, 150 m north of diatreme.
 - c) Dike, 0.45 m. wide, 180 m. north of diatreme (Filmer 1939). Dikes 8b and 8c not located by Foster (1970).
- 9 - Indian Creek, north of old hospital along road to old heating plant. Three dikes, widest is 0.6 m. wide, strike N8W. Poor exposures (Martens 1924, Foster 1970).
- 10 - Williams Brook, 60 m. west of NY 96 (Stop 2). One dike, 3.7 m. wide on north side of creek, strike N3W (Filmer 1939).
- 11 - Six Mile Creek, 1.2 km. south of 30 foot dam. One dike, 5 cm. wide, strike N5E. (Martens 1924). Not located by Foster (1970).
- 12 - Six Mile Creek, 90 m. north of 30 foot dam. One dike, 5 cm. across, strike N2W (Martens 1924).
- 13 - Six Mile Creek, 80 m. south of pumping station. Two dikes, 25 cm. and 10 cm. wide, 9.5 m apart in stream bed, strike N1E (Martens 1924).

- 14 - Near Six Mile Creek, 60 m. up ravine near small lake (now covered by Ithaca reservoir), at base of falls. Weathered igneous mass, (diatreme ?), 3 m. by 2.4 m. (Filmer 1939). Not located by Foster (1970).
- 15 - Six Mile Creek, 180 m. north of old pumping station. Two dikes, 25 cm. and 10 cm. wide, 10 m. apart in stream bed, strike N3E (Martens 1924).
- 16 - Quarry on Brandon Place, near Six Mile Creek in southeast part of City of Ithaca. No longer accessible. Four dikes, widest is 20 cm., N2E (Martens 1924).
- 17 - Cascadilla Creek in gorge below Central Ave. on Cornell University campus (next to Snee Hall - Dept. of Geol. Sciences, Stop 1).
a). 15 m. east of bridge at foot of steps, one dike, about 1 m. across, N2E. b). 3 m. E of bridge, 7 cm. wide, not visible, c). 40 m. west of bridge below falls, two dikes, 5-25 cm. wide, N-S strike. d). North of Eddy Gate, two dikes, spaced 0.6 m. apart, widest is 25 cm., strike N10W (Martens 1924, Foster 1970).
- 18 - Ravine, Cornell University campus south of Willard Straight Union. One dike, 35 cm. wide (Sheldon 1927, Filmer 1939).
- 19 - Fall Creek gorge, South side in first deep notch east of Stewart Ave. One dike, 10 cm. across, strike N-S (Martens 1924).
- 20 - Portland Point at Cayuga Crushed Stone Quarry (Stop 5). South and east walls of quarry. Two dikes of varying width, widest is 0.7 m., strike N5W (Martens 1924, Foster 1970). A dike, 0-25 cm. wide, about 600 m. below in the Cargill salt mine may connect with one of these dikes, strike N14W, dip N75W (Broughton 1950).
- 21 - Townley's Creek east of Ludlowville. a). About 70 m. east of falls over Tully ls., two dikes with maximum width of 20 cm., strike N10E. b). Five dikes about 335 m. east of dikes in 20a, on second step of falls, maximum width is 18 cm., strike N2W (Martens 1924).
- 22 - First ravine south of Townley's Creek. a). 300 m. east of falls over Tully ls. Three dikes, maximum width is 5 cm., strike N10E. b). 400 m. east of dike 22a above single falls. Seven dikes, maximum width is 15 cm., strike N10E (Martens 1924).
- 23 - South of Moravia, Fillmore Glen State Park, 120 m. above Pinnacle lookout. One dike, 40 cm. wide, strike N12E (Wells 1961).
- 24 - Clintonville dikes, north of Otisco Lake, first stream north of US 20 that flows east to Nine Mile Creek, 366 m. upstream in south wall. Six dikes in two groups separated by 60 m., maximum width is 30 cm, strike N5E (Smith 1931).

Well-developed, closely spaced joints often occur in the country rocks parallel to the dikes. These joints are spaced from a few mm. to several cm. and the width of the jointed zones are usually about the width of the enclosed dike. They have been related to the emplacement of the dikes by Sheldon (1927).

The diatrema at Poyer Orchard (locality 8a, Table 1) crops out as a breccia for about 60 meters along a ravine, but is known to extend both to the north and south from magnetic surveys. The other possible diatrema was reported by Filmer (1939) near the present Ithaca Reservoir and may be covered by it. Kimberlite models (see Dawson 1980) suggest that diatrema facies are emplaced at shallower levels than dikes and that the diatremes have interacted with groundwater. This suggests that a limited amount of erosion has occurred from the top of the Paleozoic section in the Ithaca region since the emplacement of the diatremes.

The relation between the dikes and jointing and faulting in the Finger Lakes region was first studied by Sheldon (1927). Comparison of the dike orientations with recent work on jointing in the region (Engelder and Geiser 1980) suggests the dikes are in the north-south 1A joint set. This makes sense as the 1A joints are the earliest and most through-going of the joint sets and were tectonically induced by high pore pressures. The concentration of the dikes in the 1A joints suggests that the north-south 1B and east-west joint sets were not open at the time of dike emplacement. The emplacement of dikes in the north-south joints gives little tectonic or age information as it is consistent with the predominantly east-west extensional stresses that have existed in the Finger Lakes region since the Paleozoic (north-south compression during Appalachian folding, east-west extension associated with the opening of the Atlantic).

Isotopic dating indicates a lower Cretaceous age for the dikes. An age close to 140 m.y. is suggested by a Rb-Sr date of 136 ± 8 m.y. on the large mica fraction and a 145 ± 7 m.y. K-Ar date on the small mica fraction from the western dike at Portland Point (Table 1, locality 20) (Zartman et al. 1967) and by whole-rock K-Ar ages (Basu et al. 1984) for the Williams Brook (locality 10) dike (139 ± 7 m.y.), the Frontenac Creek (locality 3) dike (140 ± 8 m.y.) and the Cascadilla Gorge (locality 17) dike (146 ± 8 m.y.). Younger ages reported by Basu et al. (1984) on dikes from Taughannock Creek (121 ± 23 m.y.) (locality 5) and Portland Point (113 ± 11 m.y.) (locality 20) have large error bars, but could suggest a second period of intrusion.

An early Cretaceous age is also consistent with pole positions inferred from magnetic polarities (DeJournett 1970). At least two periods of intrusion were postulated by DeJournett (1970) as both normal and reversed magnetic polarities were observed in the dikes. Normal polarities (62.2° inclination and 346.9° declination) were found for dikes from Williams Brook, Taughannock Creek and Cascadilla Gorge and a reverse polarity (inclination -50.4° and declination 167.5°) was measured for a dike from Portland Point. Later work by DeJournett (personal communication to A. Bloom, 1971) indicates that both normal and reverse polarities occur in multiply intruded dikes at Portland Point. Detailed timing information has not been extracted from this data due to the large number of magnetic reversals which occurred in the lower Cretaceous.

TECTONIC FACTORS CONTROLLING THE EMPLACEMENT OF THE DIKES

Emplacement of the Finger Lakes dikes may be related to a northeast-southwest trending zone of crustal weakness inferred from geophysical data (Snedden 1983, Parish and Lavin 1982). As discussed by Snedden (1983), the zone is indicated by changes in Bouguer gravity (see also Parish and Lavin 1982) and aeromagnetic signatures across the region and the location of structural lineaments inferred from satellite imagery. This lineament is not the prominent New York-Alabama lineament of King and Zeitz (see Dennison 1983) which is east of this region. Dennison (1983) suggests that the Finger Lakes dikes as well as a string of dike localities that extend from Norris Lake, Tennessee to Quebec are controlled by the axis of the deepest part of the mid-late Paleozoic Appalachian Basin. The basin axis may be controlled by a preexisting basement feature.

Parish and Lavin (1982) further suggest that the dikes are concentrated in zones of crustal weakness where the regional northeast-southwest lineament intersects northwest-southeast trending fractures. However, as pointed out by Dennison (1983), some of the proposed cross-trending basement faults in Pennsylvania are not convincing. Furthermore, similar cross-trending features are not apparent in association with the Finger Lake dikes. Snedden (1983) points out that dikes are absent where granitic bodies are inferred in the basement and suggests that these granites control where the dikes reach the surface.

The mechanism that triggered the eruption of the Finger Lakes dikes remains unclear. Crough (1981) suggested that regional uplift and kimberlite emplacement occurred when the Finger Lakes region passed within 5 degrees of the Great Meteor hotspot. Although the timing fits, the distance is great and the model does not explain the occurrence of similar dikes in Pennsylvania. Likewise, a hotspot track has been suggested from the New England Sea Mounts to Quebec (most recently by Foland et al. 1986 who obtained a 124 m.y. age on the Quebec intrusions). The Finger Lakes are even farther from this proposed hot spot track. Several authors (i.e., Taylor and Hunter 1982, Parish and Lavin 1982) have related the intrusion of the dikes to the opening of the Atlantic ocean. The correlation is not straight-forward as the Finger Lakes dikes were intruded long after rifting initiated, during a period of steady-state spreading of the central North Atlantic south of Newfoundland (160-135 m.y. ago) (see synthesis of McHone and Butler 1984). The explanation of the triggering mechanism is tied to understanding the cause of the widespread, post-rifting alkaline and ultramafic igneous activity in the eastern US.

MECHANISM OF DIKE EMPLACEMENT

Foster (1970) and Reitan et al. (1970) concluded that the dikes could not have been emplaced as magmas. This conclusion is supported by the sharp contacts and lack of visible metamorphism in the adjacent sediments, the textures of the dikes and the vertical extent of very thin dikes. Based on determination of Curie temperatures, DeJournett (1970) estimated temperatures of emplacement to be much in excess of 525°C (Curie temperature of dike) for the normally polarized dikes and between 490°C (Curie temperature of dike) and 580°C (Curie temperature of shale) for the reversely polarized dikes.

The probable mechanism of emplacement of the Finger Lake dikes is by fluidization which occurs when fragmental material is transported in a fast-moving gas stream (Reitan et al. 1970). Such a mechanism has also been suggested for other kimberlites (see Dawson 1980). Entrained oriented country rock xenoliths in the dikes and the closely spaced joints in the country rocks along the margins of some dikes support the fluidization mechanism.

THE FINGER LAKES INTRUSIONS AS KIMBERLITES

A question that has been asked about the Ithaca dikes is "Are they really kimberlites?" The question revolves around the evolving definition of kimberlite (see Dawson 1980) and whether or not, the mineral melilite occurs in the dikes, in which case they are more properly designated alnoites.

Melilite was first reported in the Finger lakes region in the Syracuse dikes by Smyth (1902) and was subsequently reported in the Ithaca dikes by Martens (1924) and Foster (1970). Reexamination of the samples studied by Martens and Foster has failed to confirm the presence of melilite. Basu et al. (1984) also were unable to find melilite and point out that the major element compositions of the New York kimberlites are essentially similar to South African kimberlites and dissimilar to olivine melilites. Until melilite is positively identified by microprobe analyses, the occurrence of melilite in the Ithaca kimberlites is an open question.

Mineral components in the Ithaca dikes consistent with their classification as kimberlites are olivine, phlogopite, diopside, opaque oxides, perovskite, sphene and calcite. Unstrained olivine grains with compositions near F088 are interpreted as phenocrysts, while strained olivine grains with higher FO contents (F091.5) are interpreted as xenocrysts (Basu et al. 1984). Most phlogopite compositions are like those in normal kimberlite groundmass micas and clear diopside rims overgrown on corroded clinopyroxene xenocrysts have compositions similar to groundmass diopside in other kimberlites (Kay et al. 1983). Both phlogopite and diopside compositions are slightly more Al-rich than those in the South African kimberlites (see Dawson 1980 for references). Opaque oxides are ilmenite and titanomagnetite which have reacted with the melt to form perovskite (Jackson et al. 1982; Basu et al. 1984). High Mg-ilmenite, a phase common in many kimberlites, is absent in the Ithaca dikes.

Basu et al. (1984) report that the Finger Lakes dikes have a range of Nd isotopic compositions ($\epsilon_{Nd} = +1.2 - +4.2$; Williams Brook and Portland Point, $\epsilon_{Nd} +4.2$; Cascadilla Gorge, $\epsilon_{Nd} = +1.2$) similar to that in 90 m.y. kimberlites from South Africa ($\epsilon_{Nd} = +0.8 - +2.1$). This range of Nd isotopic composition does not overlap that of other known groups of volcanic rocks and supports the interpretation of the Finger Lakes dikes as kimberlites. The range of values in the Finger Lakes region is interpreted as reflecting mantle heterogeneity (Basu et al. 1984).

Using a model proposed by Wyllie (1980) as a basis, Kay et al. (1983) proposed that the Ithaca kimberlites represent the early stages of kimberlite eruption and never achieved the eruptive maturity of larger kimber-

lites such as those in South Africa. This suggestion was based on the narrow, one or two-stage dikes typical of the region and the relatively shallow mantle origin of the fragments contained in the dikes (see below).

On the other hand, as summarized by Dawson (1980), kimberlites in circum-cratonic orogenic belts are generally diamond-free and tend to be associated with other rare, high-volatile ultra-alkaline rock types of limited volume, while those in cratons tend to be diamond-bearing. The Finger Lakes dikes are of the former type petrologically and lie on the western margin of the Appalachian fold belt. The correlation between kimberlite type and location suggests the type of kimberlite erupted is dependent on the age, composition and thickness of the lithosphere, as well as the volume of the kimberlite.

THE LOWER CRETACEOUS UPPER MANTLE AND CRUST IN THE ITHACA REGION

Xenocrysts and rare crystalline xenoliths in the Ithaca dikes are consistent with derivation from depths of less than 150 km. (Snedden 1983, Kay et al. 1983). Reconstructed mantle assemblages include spinel and garnet peridotites and garnet pyroxenites (Kay et al. 1983, Schultze et al. 1978). Eclogite xenoliths have been reported from the Taughannock dikes by Jackson et al. (1982) and Basu et al. (1984). Secondary pargasitic amphibole associated with the peridotite assemblages are attributed to mantle metasomatism (Kay et al. 1983). The Cr-rich assemblages which include sub-calcic diopsides common in many kimberlites (see Dawson 1980) have not been observed.

The spinel peridotite inferred from the high-Al mineral assemblage is the type commonly found in alkali basalts and associated with high geothermal gradients or undepleted mantle. No other geologic evidence suggests that heat flow has been high in the Ithaca region since the PreCambrian. As discussed by Kay et al. (1983), several interpretations are possible. First, the spinel peridotite could represent unequilibrated upper mantle of Pre-Cambrian Grenville age. Second, if the mantle below Ithaca is very undepleted in basalt-producing elements, the whole rock composition-dependent spinel to garnet peridotite transition may be at a greater depth than assumed in determining the geothermal gradient. In this case, a high geothermal gradient is not necessarily implied by the xenocrysts. Third, the implied geothermal gradient could be a transient geotherm related to the rise of the diapirs and therefore, not reflect conditions in the surrounding mantle. With the information available, there is no easy way to choose between these alternatives.

Crystalline fragments of presumed crustal origin include a suite of Mg-rich minerals which may belong to a granulite carbonate assemblage, small mafic syenite xenoliths and clinopyroxene and garnet xenocrysts similar to minerals found in basic to intermediate composition granulite facies rocks (Kay et al. 1983). These fragments are all compatible with a granulite facies metamorphic basement similar to the Grenville terrain in the Adirondacks beneath the Paleozoic section and a granulite facies lower crust in the Ithaca region. The high-Mg assemblage could also be from hydrated ultramafic rocks in the upper mantle or lower crust, but the proximity of the Grenville marbles to the north in the Adirondacks and the

presence of marbles in basement drill holes (see introductory chapter of this guide) lend support to the metamorphic carbonate interpretation.

The most abundant xenoliths in the dikes and diatremes are from the Paleozoic section described in the introductory part of the guidebook.

COMMENT ON THE PROBABILITY OF FINDING DIAMONDS

The absence in the Ithaca dikes of the suite of Cr-rich xenoliths and xenocrysts associated with mantle depths below the diamond-graphite transition and usually found in diamond-bearing kimberlites, makes the prospect of finding diamonds in the Ithaca dikes unlikely. The recognition of graphite in several dikes (Martens 1924) also supports this conclusion. The rotting remains of the unsuccessful sieving operation in the ravine at Poyer Orchard diatreme attests to the lack of success in finding diamonds in the Ithaca region.

ACKNOWLEDGEMENTS

We would like to thank the people at the Cayuga Crushed Stone Quarry for their generous access to the quarry and their continued interest in the scientific value of geologic features in the quarry. We also thank Daniel Karig and Robert Kay of the Department of Geological Sciences, Cornell, for discussion of the Portland Point outcrop and the owners of the Poyer Orchard for access to their property.

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Ten dikes crop out in Taughannock Creek (Figure 2) over a distance of 0.8 kilometer upstream from the road bridge above the high falls. You are even with dikes 2-5 at the parking area. These dikes outcrop best on the ledge on the south side of the creek. Dikes 4 and 5 are 1.2 meters apart. Dike 3 is about 16 meters east of dike 4 and dike 2 is about 3.5 meters east of dike 3. Dikes 6-8 crop out above the stream in a closely spaced cluster, a short distance to the west of the parking area. Dikes 9-10 are upstream about 30 meters east of a stress-release "pop-up" in the Devonian shale in the middle of the stream. Dike 10 is exposed under a tree on the south bank of the river. The outcrop has been considerably reduced by sampling in recent years. Dike 9 is very difficult to find.

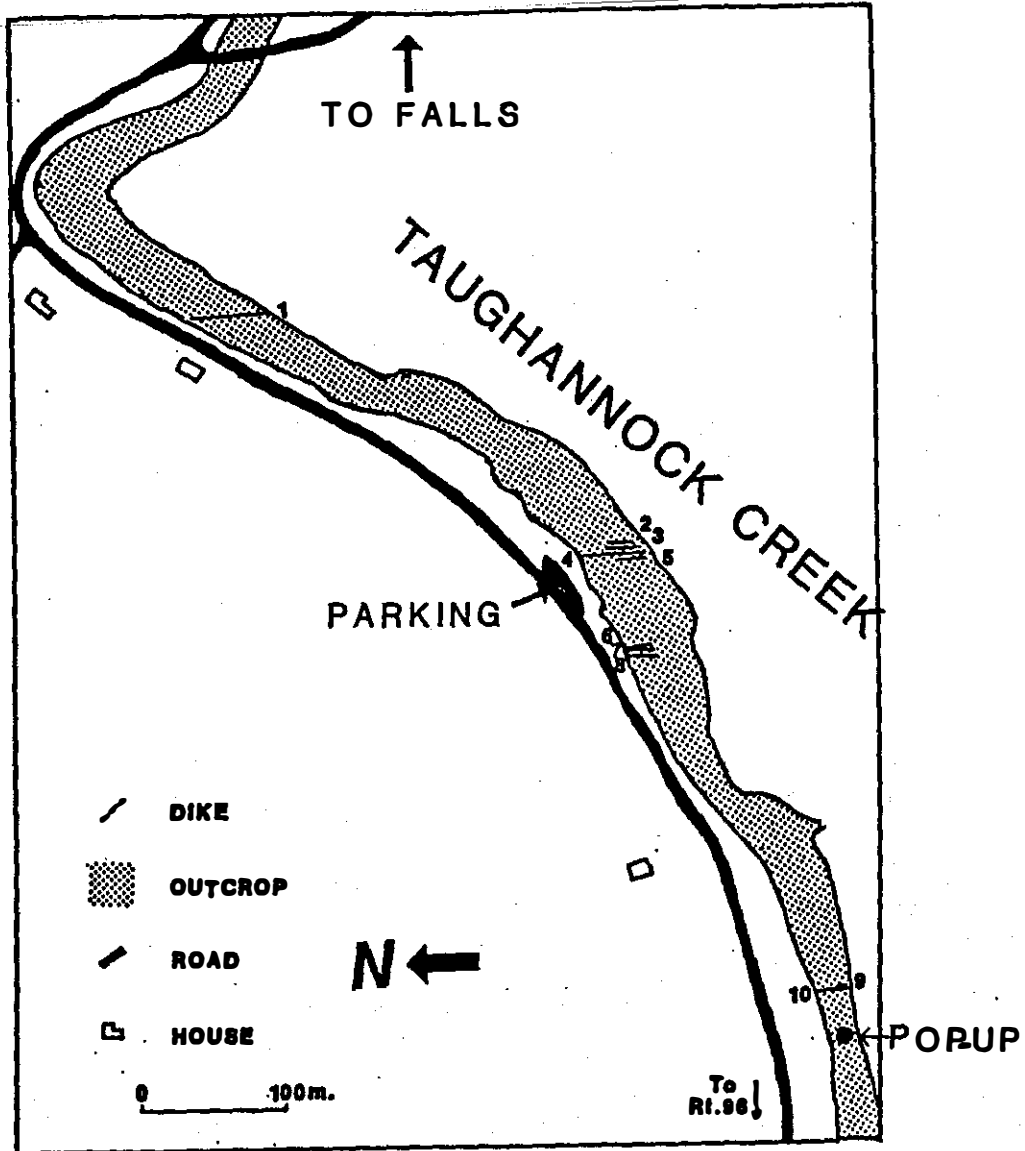
The control of the dike orientation by the prominent N-S joints is particularly well exhibited at this locality. In addition, the closely spaced joints that typically occur in the country rocks parallel to the dikes are well developed along several of the dikes. Dikes 1, 4 and 10 extend across the exposed bedrock outcrop while the other dikes pinch out rapidly in both directions. The occurrence of the dikes in clusters (1; 2-3; 4-5; 6-8; and 9-10) suggests branching by one or more feeder dikes near the present surface. A large shale xenolith in the center of dike 8 suggests how this divergence occurs. The composite width of each dike group appears to remain constant across the outcrop width as illustrated by the northward thickening of dike 4 and the southward thinning of dike 5. Dike 10 shows (or did show) evidence for three possible injection episodes in which material was deposited along the contacts and not completely removed by later events.

Petrographic differences occur among the dikes at this locality. The general mineralogy of all the dikes is similar to that in the Cascadilla Gorge dike (Stop 1), but the modal percentages are variable (Foster 1970). Dike number 10 is the widest and freshest dike at this locality and has 37% serpentine, 7% phlogopite, 5% magnetite and perovskite and 51% groundmass (Foster 1970). Most of the other dikes show considerably more alteration. All of the dikes have high concentrations of calcite in the ground mass (dike 1 - 35% calcite, 2 - 25%, 4 - 36%, 7 - 23%, 10 - 23%) (Foster 1970).

The xenocryst assemblage in the Taughannock dikes consists of pyrope garnet, green chrome diopside, augite and spinel. Dikes 6, 7 and 8 are anomalous in that they are very calcite rich (up to 30%), yet contain abundant xenocrysts of unaltered olivine, Cr-diopside, augite and garnets (Foster 1970). Garnet is particularly abundant in dike 8. Small eclogite xenoliths have been reported by Basu et al (1985) and Taylor and Hunter (1982). Mineral compositions for xenocrysts in these dikes are reported by Kay et al. (1983). Martens (1924) reports finding graphite in dike 10.

DeJournett (1970) reports that dike number 10 has a normal magnetic polarity with a pole position consistent with a lower Cretaceous age. Basu et al. (1984) obtained a whole rock K-Ar age of 121 ± 23 m.y on one of the dikes from this locality.

- | | | |
|------|-----|----------------------------------|
| 13.6 | 0.0 | Turn around and return to NY 96. |
| 14.5 | 0.9 | Turn left (south) on NY 96. |



Sketch map of the dikes occurring in Taughannock Creek upstream (east) of the falls (Stop 3). Stippled pattern shows outcrop of the Devonian shales. The width of the river is not indicated. Dike locations are approximate. Dikes 2 and 3 are farther from dikes 4 and 5 than suggested by map (see description in discussion of Stop 3). The maximum widths and strikes of the mapped dikes are shown below. Data and map is from Foster (1970). Pop-up refers to Devonian bed in stream that has been uplifted and cracked as a result of stress release due to former quarrying of the flagstones in the creek.

	Maximum Width	Strike
1	4	N 10 W
2	2	N 10 W
3	7	N 9 W
4	10	N 9 W
5	8	N 9 W
6	6	N 19 W
7	6	N 10 W
8	6	N 14 W
9	1	N 15 W
10	10	N 15 W

FIGURE 2

- | | | |
|------|-----|---|
| 21.7 | 7.2 | Turn left (east) on Duboise Road. Note sign to Poyer Orchard. |
| 22.5 | 0.8 | Poyer Orchard sales office. Ask permission to enter orchard. Permission may be denied due to insurance risks. |
| 22.6 | 0.1 | Return west on Duboise Road to Orchard entrance. Park. |

STOP 4. POYER ORCHARD DIATREME (Locality 8a, Table 1)

Walk about 500 meters along the orchard road to the old part of the orchard. Head north north-east to the ravine about 50 meters beyond the orchard boundary. The diatreme is exposed in the bottom of the ravine where it is particularly wide due to the preferential erosion of the diatreme relative to the surrounding shales. Remnants of the sluicing operation are visible near the diatreme.

The amount of visible exposure depends on the intensity of recent storms. In the original report on the locality, Barnett (1905) was uncertain whether two dikes (7.6 and 4.0 meters across respectively) or a single intrusion (46 meters across) occurred. After an exceptionally large storm in the late thirties, Filmer (1939) described the intrusion as a breccia pipe or a swelling in a single large dike that extended 58 meters in an east-west direction and 24 meters in a north-south direction.

In subsequent literature (i.e. Foster 1970), the outcrop has been called a diatreme. The petrology of the unit is consistent with its classification as a tuff breccia of the kimberlite diatreme facies (see Dawson 1980). The outcrop is composed of extremely abundant xenolithic fragments surrounded by a very altered groundmass consisting of hydrothermal-type minerals. The groundmass has not been studied in detail. The outcrop is cut by several irregular tuffaceous dikes about 15 cm. across. An x-ray diffraction scan of the tuffaceous dike material by Foster (1970) indicated the presence of microcrystalline calcite and phlogopite.

Most of the xenoliths are angular pieces of sedimentary rock derived from the enclosing Devonian shale. These inclusions average about 5 cms. in diameter, but some as large as a half meter in diameter have been observed (Foster 1970). Fossils in inclusions reported by Filmer (1939) have not been studied. Other sedimentary inclusions are rounded sandstone and quartz pebbles. Rounded pieces of kimberlite similar to those found in nearby localities also occur as inclusions. Several syenite xenoliths, approximately 12-15 cm. across, were found during the sluicing operation (Wells, 1977, personal communication) and similar smaller xenoliths were analyzed by Snedden (1983). A fragment containing actinolite, chlorite, potassium feldspar and sphene reported by Kay et al. (1983) suggests that lower grade Precambrian rocks overlie the granulite facies in the basement in this locality. The heavy mineral fraction of the diatreme contains relatively abundant garnet, spinel and pyroxene xenocrysts (Foster 1970, Snedden 1983, Kay et al. 1983).

A magnetic anomaly map of the locality (Figure 3) produced by Angerani and Hanguis (1978) as a term project in a geophysical methods course at Cornell gives additional information on the extent of the body. The map shows that the body is elongate north-south and continuous along the length of the survey (150 meters). The most intense magnetic anomaly occurs just to the north of the stream and a second smaller anomaly occurs parallel to the main anomaly about 20 m. to the east. Angerani and Hanguis modeled the anomalies as vertical dikes and suggested that the width of the main dike varies from 1.4-4.0 meters and that its top is buried from 1.6 to 5.1 meters below the surface. In the model, the maximum depth and widest part of the intrusion occur at the maximum anomaly. Their model predicts the second dike is 4.1 meters wide and 11.1 meters deep.

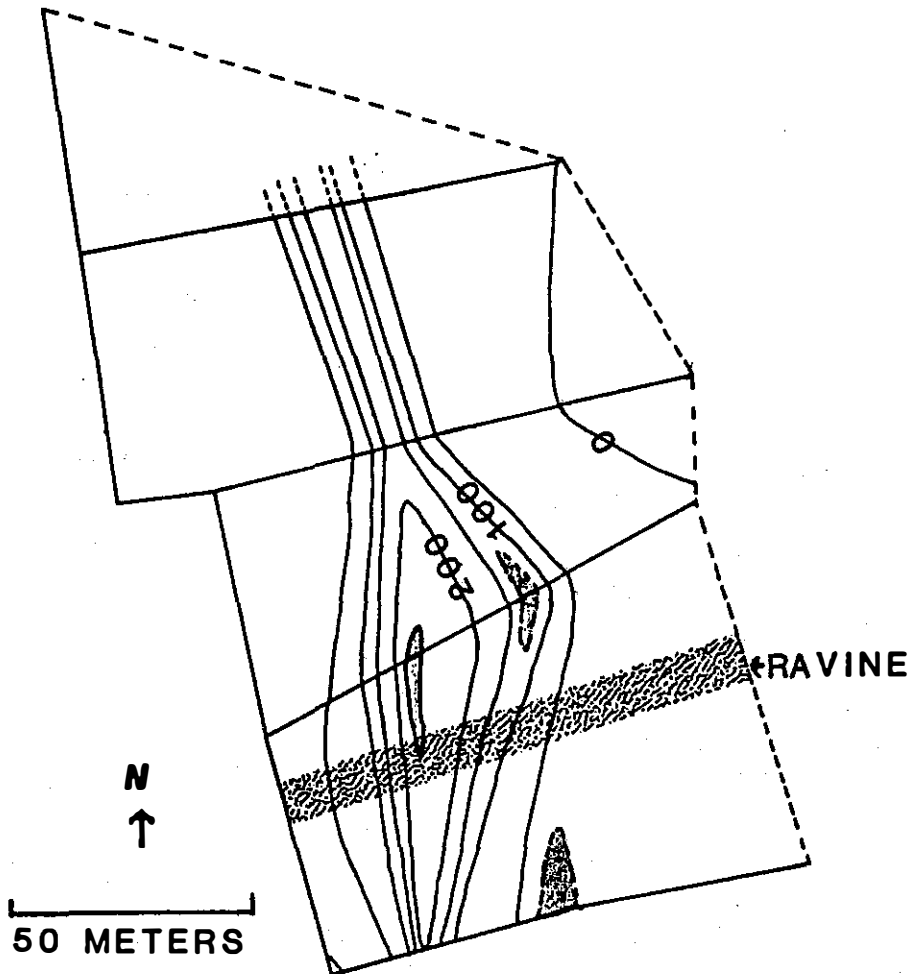
The magnetic model does not take into account that the stream exposure is in the diatreme facies. Combining the magnetic information with the petrography of the outcrop suggests that the diatreme represents a wide spot in a dike which along strike is similar to other larger dikes in the region (i.e. those in the same ravine, see localities 8b and 8c in Table 1). Although the magnetic data may indicate that two dikes occur, the discontinuous nature of the smaller anomaly is consistent with the magnetic pattern indicating the irregular configuration of the diatreme. The magnetic modeling in the narrow part of the anomaly north of the stream (Figure 3) is consistent with a dike 1.4 meters wide, striking N5W (Angerani and Hanguis 1978).

As reviewed by Dawson (1980), diatreme facies kimberlites are, in general, interpreted to result from explosions resulting from the rapid cooling by groundwater of fluidized mixtures rich in exsolved high pressure gas. This suggests that the Poyer Orchard diatreme formed as part of the dike encountered groundwater and an explosion occurred. This interpretation is supported by the abundance of angular fragments from the surrounding country rock. The tuffaceous dikes would then result from the explosion. If the diatreme formed by this mechanism, the present erosion surface at Poyer Orchard was within the groundwater table in the lower Cretaceous.

22.6	0.0	Return to NY 96 via Duboise Road.
23.3	0.7	Turn left (south on NY 96) and return to the city of Ithaca.
26.2	2.9	Stoplight. Continue on NY 96 through light and cross bridge over Inlet Creek. Continue straight ahead to intersection with NY 13 and NY 34.
26.6	0.4	Turn left (north) on NY 13 and NY 34 (N. Meadow Street). Continue north on NY 13 and NY 34.
28.2	1.6	Follow ramp on right to NY 34 North.
28.4	0.2	Turn left (west) at stop sign onto NY 34 North. Continue north on NY 34 to South Lansing.

MAGNETIC MAP

POYER ORCHARD DIATREME



CONTOUR INTERVAL - 50 GAMMAS

Magnetic map of kimberlitic intrusion in Poyer orchard (locality 8, Figure 1; locality 8a, Table 1) modified from Angerani and Hangu (1978). See discussion for stop 4. Survey was made with two portable proton precession magnetometers. Measurements were taken at 2 meter intervals along the solid straight control lines and corrected for lateral and diurnal variations in the earth's magnetic field measured at 3 minute intervals. The zero contour was arbitrarily chosen as that for the initial reference value (56882 gammas). Shaded area on left is region of maximum anomaly (266 gammas). Shaded regions in dashed areas on right are minor anomalies above the smooth gradient (52 and 127 gammas respectively).

FIGURE 3

34.0	5.6	Junction with of NY 34 with NY 34B at Rogues Harbour Inn. Turn left (west) towards King Ferry on NY 34B.
34.7	0.7	Turn left (west) on Portland Point Road.
35.1	0.4	Main entrance to Portland Point Quarry.

STOP 5. PORTLAND POINT QUARRY (Locality 20, Table 1)

THIS IS PRIVATE PROPERTY. YOU MUST OBTAIN PERMISSION TO ENTER. Collecting is allowed. Fossil collecting is also good in the quarry (see guides for other trips). The quarry is complex and constantly changing. Ask quarry personnel for best places to view kimberlite dikes.

Walk south 1 kilometer along the main quarry road to the east-west trending southern quarry wall (parallel to the north side of Falls Gulf Creek). In August 1986, this was the southern part of the main pit. Quarrying was complete and the region was being backfilled. The slag heaps near this area of the pit had blocks of kimberlites.

Two segments of north-south striking dike are exposed in the east-west trending wall of the quarry, above and below the nearly horizontal trace of a thrust fault which strikes east-west and has a shallow dip to the south (third dimension of fault is exposed in adjacent north-south trending quarry wall). The fault, which is marked by a slickensided surface above a brecciated zone, is associated with the crest of the Firtree anticline, a late Paleozoic Alleghenian fold (i.e. Engelder and Geiser 1979). Slickensides dipping 20° to the south can be seen on the east-facing dike surface adjacent to the country rock indicating some post emplacement differential slip between the dike and the wall rock.

As discussed earlier, the dike segments have an early Cretaceous age (Zartman et al. 1967, DeJournett 1970, Basu et al. 1984). They are not offset by the fault, which apparently constituted a discontinuity in the north-south joint system across which the dike was diverted. The upper dike segment, which is about 1.5 meters to the east of the lower segment, is uniformly about 20 cm. wide and terminates in the fault zone. The lower segment tapers from a maximum width of about 20 cms. to about 12 cms. at the fault. As it passes through the rubble zone below the fault, the dike bends and shows at least one minor offset of about 20 cm. to the east. The lower dike segment also tapers to about 20 cm. in width near the bottom of the exposure that outcropped in August.

Another exposure in the quarry where dike segments are bounded by a fault zone was described by Sheldon (1927) and may have been along strike of the present exposure. On the basis of the geometry of pinching and swelling of dikes in that exposure, Sheldon argued that dike intrusion was contemporaneous with the latest stages of faulting. The relationship of the dike segments to the fault zone in the modern exposure, and the observation that pinching and swelling occur in discontinuous dikes in other localities where evidence of faulting is not observed (Stop 3 and below), strongly suggest that the dikes observed by Sheldon were also not offset by the fault.

The exposed dikes may be along strike of the locality where DeJournett (letter to A. Bloom 1971) mapped dikes extending about 120 meters across the quarry floor in 1970-71. He indicated that these dikes showed normal magnetic polarity in the interiors and reverse magnetic polarity on the exteriors. He also stated that the dikes showed textural evidence consistent with two episodes of intrusion and that these textures correlated with the magnetically distinct zones. His map clearly shows that the dikes pinch and swell and are discontinuous. These features were not visible in 1986.

Leave the main pit and head north and east along road to the older inactive part of the quarry. The road to the bottom of the pit of interest is blocked by a small slag heap. This pit is in the eastern part of the quarry, southeast and across a large slag heap from the quarry lake that is visible from the hill above the pit (see Ludlowville 7.5 minute topographic map). In August 1986, this pit had much less mud and grass on the floor than surrounding pits.

In this pit, a N-S dike about 5-7 cm. wide can be followed across the quarry floor. The exposure has deteriorated considerably in recent years due to sedimentation on the quarry floor and the weathering of the dike. Blocks of the kimberlite that have sedimentary xenoliths and show textural inhomogeneities can be found in the surrounding slag heaps, particularly along the northwest wall of the pit. Some of these blocks show possible evidence for multiple intrusion.

Return to quarry entrance.

35.1	0.0	Main entrance to Portland Point Quarry. Turn around and head east on Portland Point Road to NY 34B.
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END OF ROADLOG

ICE READVANCE IN THE VALLEY OF FIVEMILE CREEK, STEUBEN COUNTY, NEW YORK
AND THE EFFECT OF THE RESULTING DEPOSITS ON POSTGLACIAL STREAMFLOW

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INTRODUCTION

Fivemile Creek rises near Jubertown swamp in the town of Jerusalem, Yates County, New York and flows southwestward to join the Cohocton River near Kanona in the town of Bath, Steuben County (fig. 1). Stratified drift deposited during the main Wisconsin deglaciation of Fivemile Creek valley is generally mantled by till or related diamicts attributed to a readvance of the ice. Younger deposits, laid down during and after the decay of the readvanced ice, include large areas of clay-silt rhythmites and of organic-rich muck. Most broad valleys of the Susquehanna River basin are floored predominantly by sand and gravel, and the presence of till, clayey silt, and muck at land surface over large areas of the floor of Fivemile Creek valley is chiefly responsible for the remarkably small discharge per square mile from this watershed during periods of low flow.

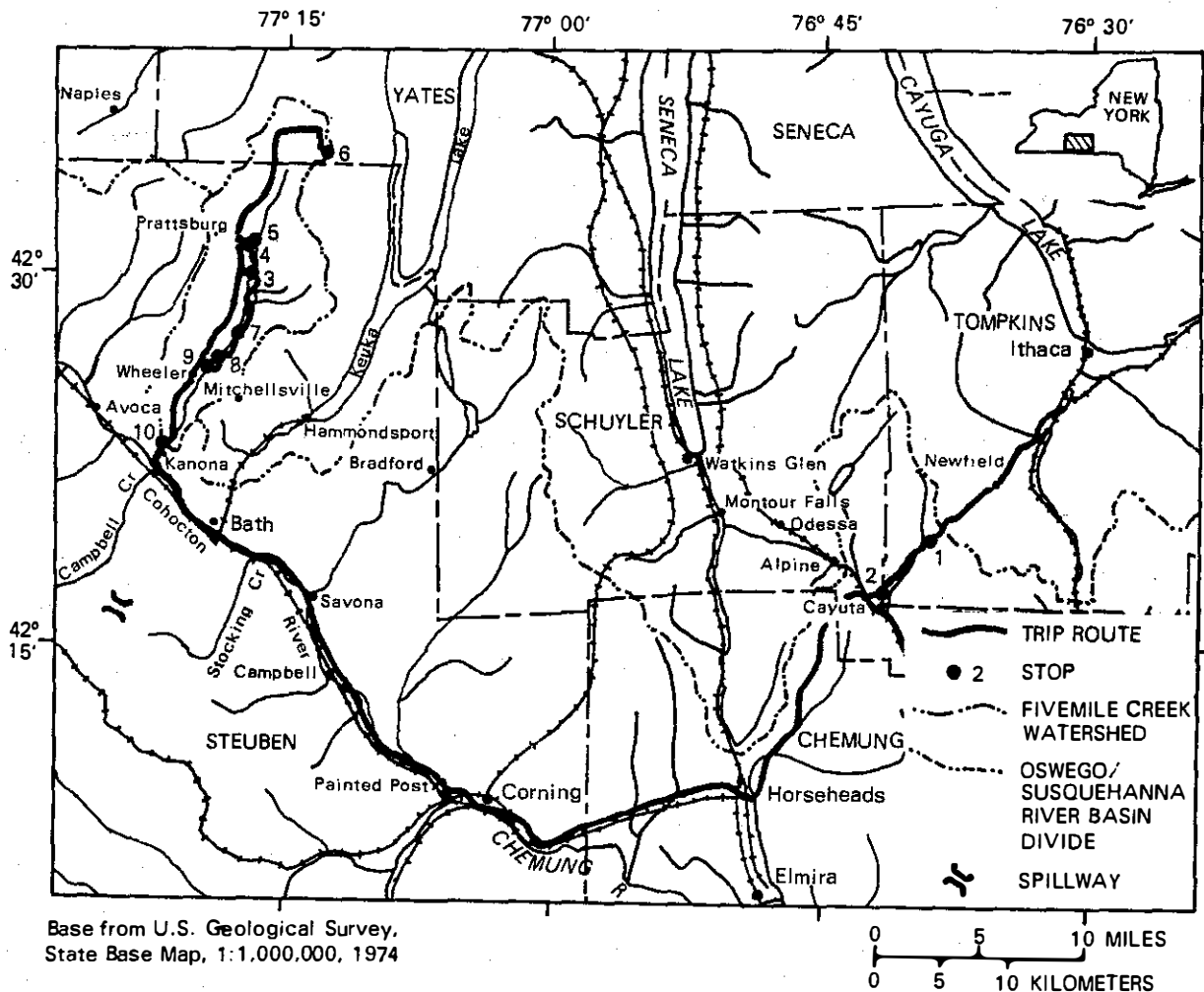


Figure 1.--Location of Fivemile Creek and route of field excursion.

This article can be read as an example of how knowledge of geology can help us understand surface-water hydrology, or of how knowledge of stream-flow can require reinterpretation of geology. It was prepared as an offshoot of studies of the interaction of aquifers and streams in the glaciated Northeastern States under the Regional Aquifer Systems Analysis program of the U.S. Geological Survey. It includes an interpretation of surficial geology and deglacial history of Fivemile Creek valley, a summary of evidence supporting that interpretation, and an explanation of the influence of surficial geology on flow of Fivemile Creek and other streams in the Susquehanna River basin. It concludes with an outline of a field excursion designed not only to display the Pleistocene stratigraphy but also to show how streamflow data, depth to water, well records, auger holes, and soils maps can be used to supplement examination of natural and artificial exposures in deciphering surficial geology. The study did not include test drilling, geophysics, regional correlations, nor determination of till fabric or provenance.

GLACIAL DRIFT IN THE SUSQUEHANNA RIVER BASIN

The surface drift over most of the Appalachian Plateau of New York, including the Susquehanna River basin, is generally considered to be the product of the ice advance that built the Wisconsin terminal moraine in Pennsylvania. This drift, referred to as "Olean," has been correlated as Altonian by Muller (1977) and as Woodfordian by Crowl and Sevon (1980). As the ice sheet melted, thinned, and gradually disappeared from south to north, upland areas were left with a mantle of till (deposited from the base of the ice as it melted) and related diamicts (emplaced by mass movements that redistributed unstable till and superglacial debris). Meltwater left few deposits in the uplands; apparently the distal few miles of the retreating ice sheet were too crevassed and porous to retain ponded water at high levels, so erosion predominated in the uplands, then as now. In the broad valleys, however, lakes formed in reach after reach as soon as the ice melted down to the level at which water was ponded behind drift previously deposited down-valley. Sediments accumulated in these lakes, typically in three facies (fig. 2). The earliest of these facies, deposited when the valley was still largely choked with ice, is coarse, heterogenous, and commonly silty. Later, as melting exceeded deposition, large expanses of open water formed, in which fine-grained sediment settled on the lake bottom. Much of the fine-grained sediment was eventually capped by deltaic outwash or inwash that filled the shoaling lakes, or by alluvial gravels spread by postglacial streams across their fans and floodplains. Where lakes remained into postglacial time, fine-grained sediment rich in organic matter accumulated. Many factors, including buried ice melting at varying rates and non-synchronous development of facies in successive reaches of broad valleys as the ice margin retreated, resulted in a complex stratigraphy (Denny and Lyford, 1963; Fleisher, 1977, 1986; MacNish and Randall, 1982; Randall, 1978). At nearly all sites, however, sand or gravel caps the valley fill.

GLACIAL DRIFT IN FIVEMILE CREEK VALLEY

Overview

Much of the drift in Fivemile Creek valley was probably deposited during the retreat of the "Olean" ice sheet across the region. Ice-contact, lacustrine, and outwash facies can be recognized in exposures and

records of wells. Next, however, ice apparently readvanced at least as far south as Dineharts (fig. 3B), and deposited atop stratified drift a layer of compact stony sandy clayey silt that resembles the till that covers the bordering uplands. South of Dineharts, a distinctive sparsely pebbly non-bedded silty clay occupies the same stratigraphic position; it may have been deposited in ponded water beneath continually or episodically floating ice.

Decay of the readvanced ice resulted in a new array of stratified deposits in Fivemile Creek valley. Outwash or inwash covers most of the valley floor between Renchans and Marshalls (fig. 3A), and mantles part of the readvance till and pebbly silt between Stickneys (fig. 3B) and Renchans; hummocky topography north of Renchans is indicative of deposition against ice. Later, an extensive lake developed east and north of Stickneys, and near Beans Station (fig. 3B) a delta was built partway across the lake before the flow of meltwater and sediment ceased. At the same time or slightly later, sand and gravel was deposited amid stagnant ice 7 miles to the north near the Yates County line and Jubertown swamp (fig. 3D). These deposits are bordered by large depressions in which fine-grained sediment continued to accumulate after deposition of sandy outwash ended.

The wide extent of surficial till or related diamict in Fivemile Creek valley is unusual. A few investigators have postulated ice readvances several miles southward into the Susquehanna basin (Muller, 1966; Fleisher and Cadwell, 1984). They cited as evidence ice-contact landforms, a few scattered till exposures, and the logs of wells that penetrated gravel atop extensive lake silt, but they neither claimed nor cited evidence that a layer of till is widespread atop or within the valley fill. A layer of till at or near the top of the valley fill has been reported in several through valleys near the northern divide of the Susquehanna River basin (Randall and others, in press). In these valleys, however, the till layer is not known to extend more than 1 or 2 miles south of the divide. Surficial silts and clays are likewise rare. Silts or clays were deposited

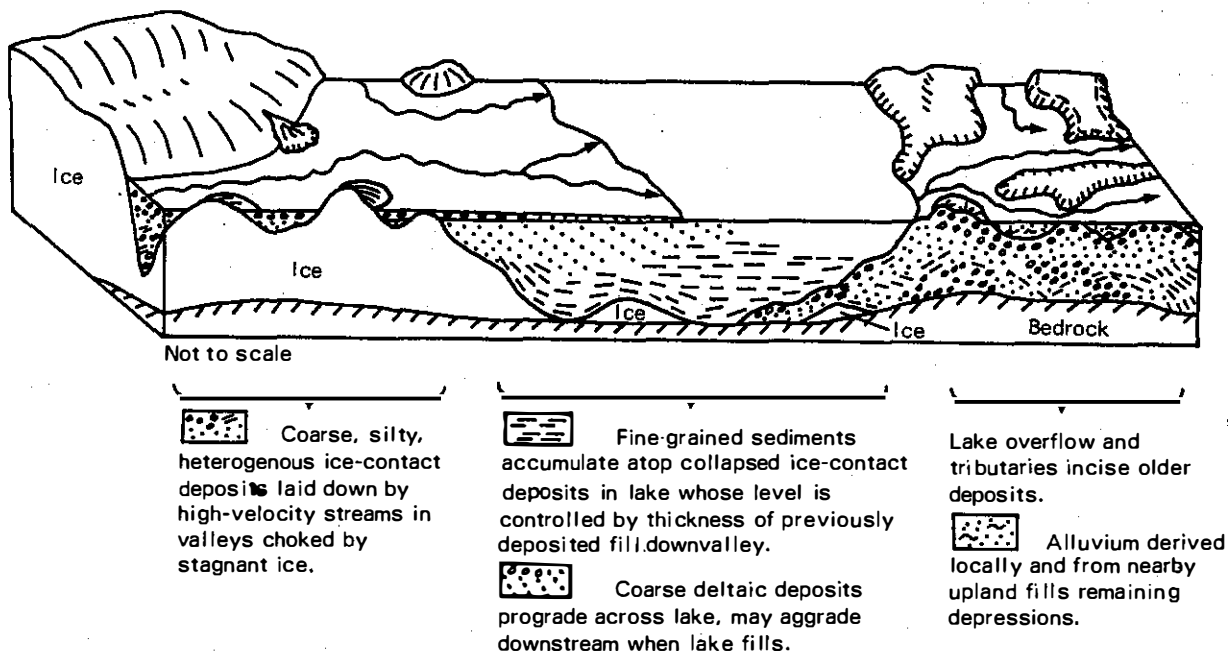
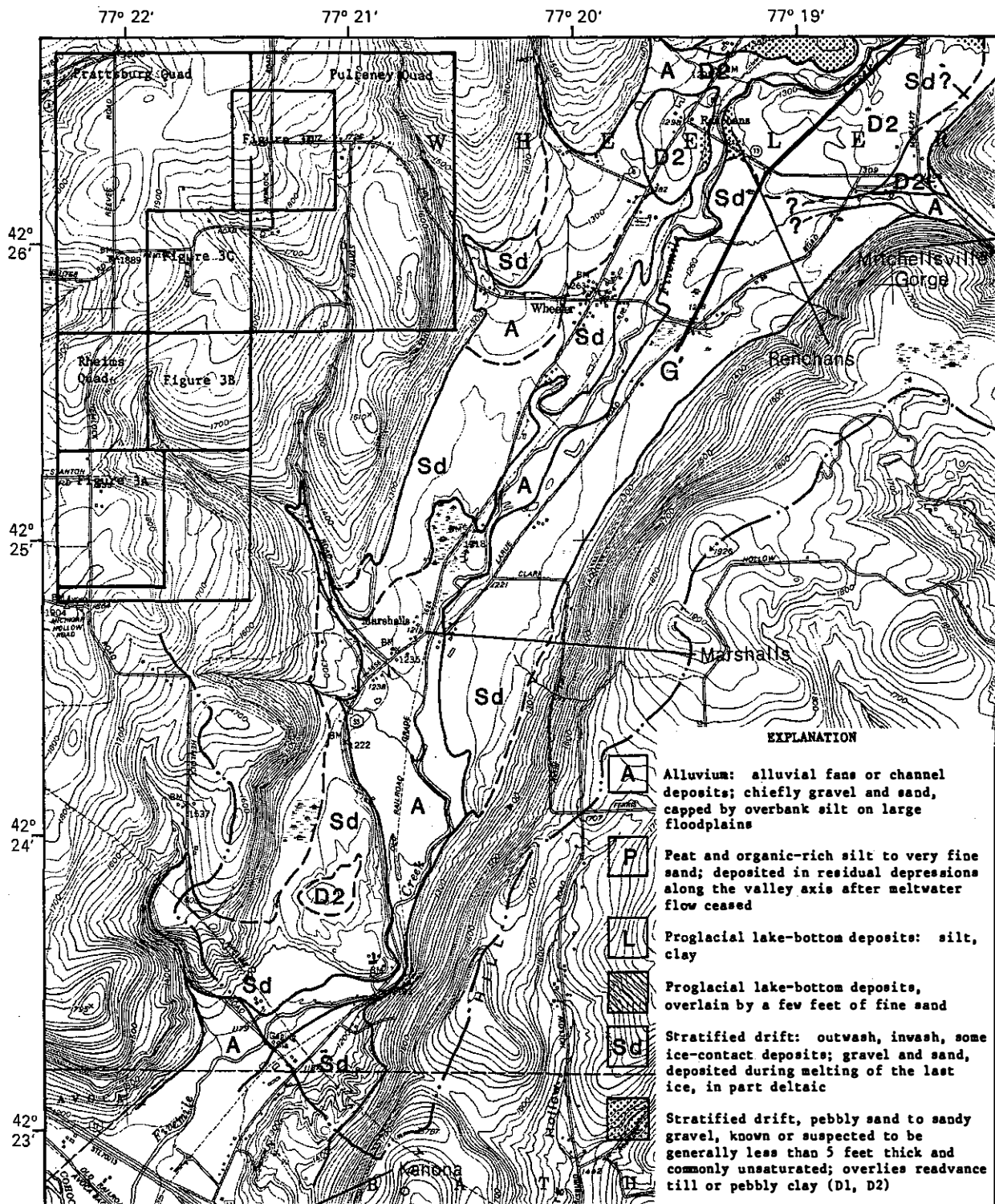


Figure 2.--Idealized diagram of typical broad valley during deglaciation.



Base from U.S. Geological Survey
Rheims, N.Y., 1953, 1:24,000

FIGURE 3A

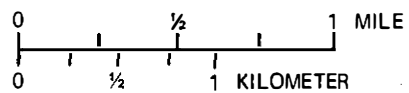
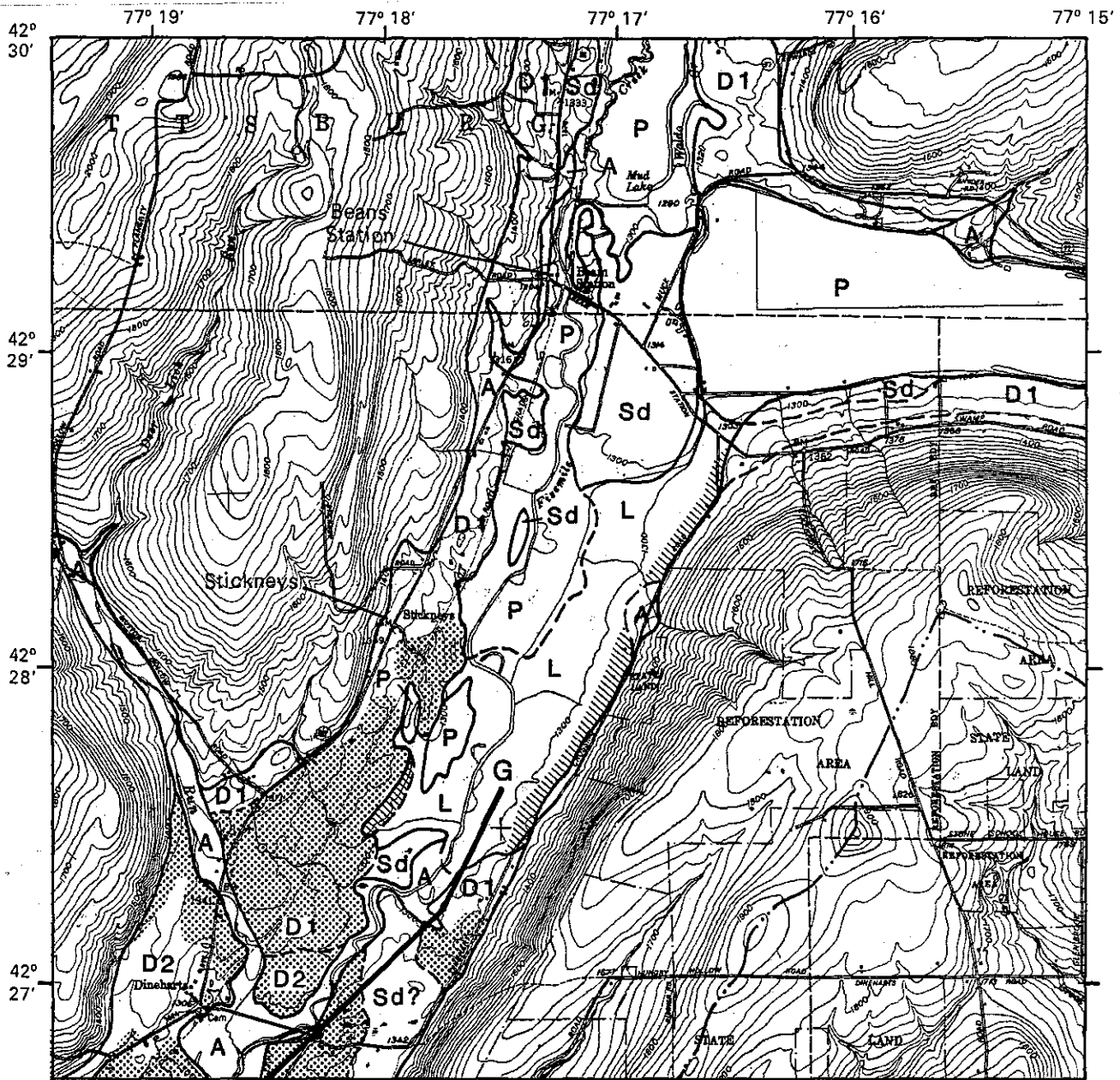
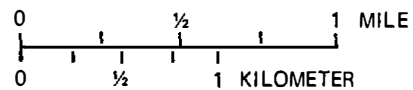


Figure 3.--Surficial geology of Fivemile Creek valley. Mapping is based on examination of earth materials at 110 points, on 70 well and testhole records, and on soils survey (French and others, 1978).



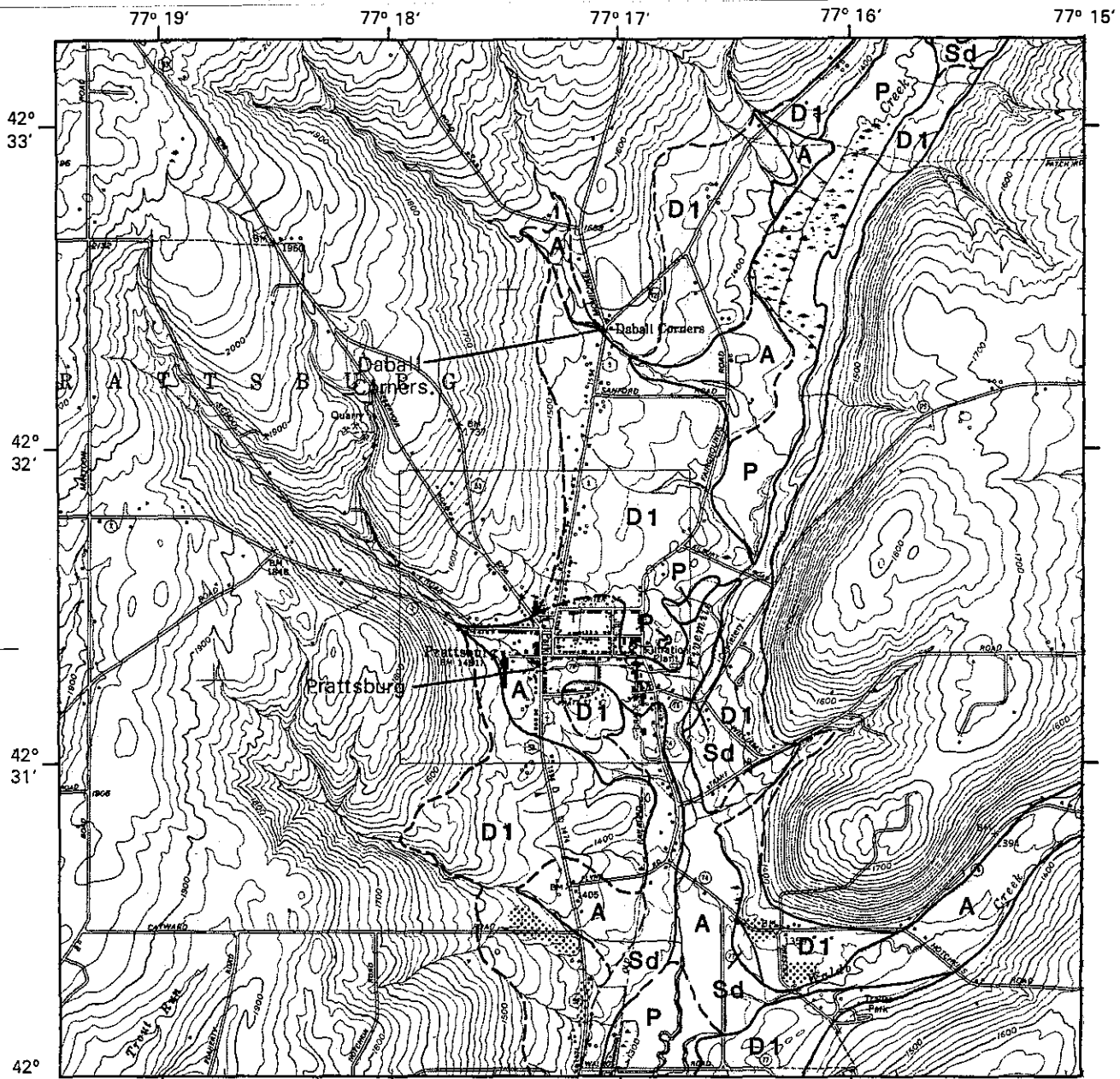
Base from U.S. Geological Survey
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FIGURE 3B



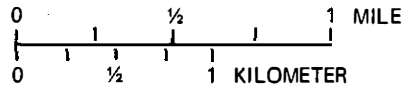
EXPLANATION (continued)

- Till, overlying bedrock in upland areas; all or lower part predates all other glacial drift.
- D1 Till and related diamicts, at or near land surface in areas of constructional topography, generally overlying ice-contact deposits; generally a compact stony sandy clayey silt containing some rounded stones; locally angular and rounded stones in a loose fine-sandy silt matrix; deposited during an ice readvance.
- D2 Pebbly silty clay; generally nonbedded, locally contains deformed wisps of silt
- Contact between geologic units, dashed where approximate
- Drainage divide, Fivemile Creek watershed
- G — G'** Line of geologic section (fig. 4)



Base from U.S. Geological Survey
Prattburg, N. Y., 1942. 1:24,000

FIGURE 3C



EXPLANATION



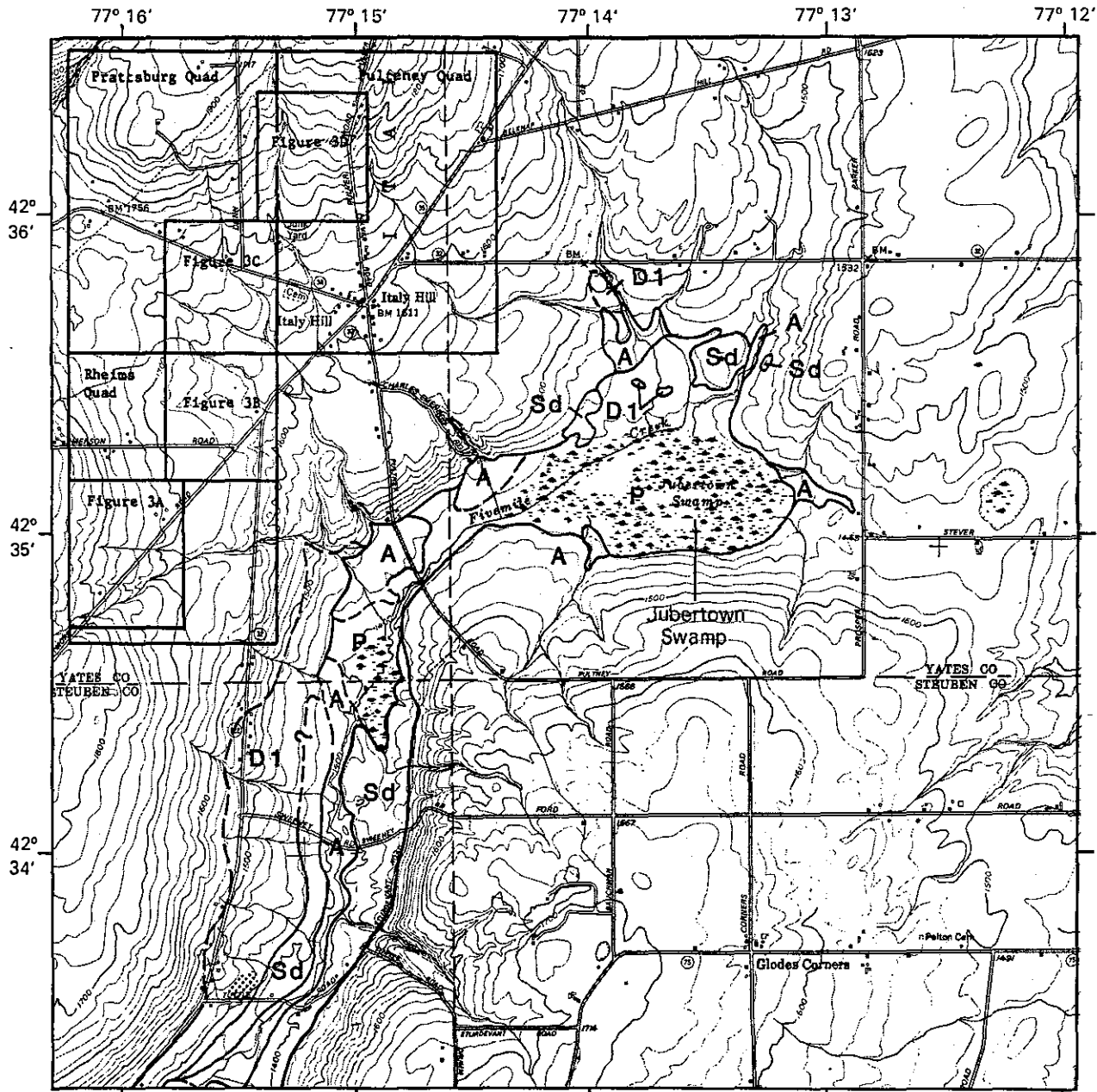
- A** Alluvium: alluvial fans or channel deposits; chiefly gravel and sand, capped by overbank silt on large floodplains
- P** Peat and organic-rich silt to very fine sand; deposited in residual depressions along the valley axis after meltwater flow ceased
- L** Proglacial lake-bottom deposits: silt, clay
-  Proglacial lake-bottom deposits, overlain by a few feet of fine sand
- Sd** Stratified drift: outwash, inwash, some ice-contact deposits; gravel and sand, deposited during melting of the last ice, in part deltaic
-  Stratified drift, pebbly sand to sandy gravel, known or suspected to be generally less than 5 feet thick and commonly unsaturated; overlies readvance till or pebbly clay (D1, D2)

Figure 3 (continued)



Base from U.S. Geological Survey
 Pulteney, N.Y., 1942, 1:24,000 and
 Prattsburg, N.Y., 1942, 1:24,000

FIGURE 3D



EXPLANATION (continued)

- Till; overlying bedrock in upland areas; all or lower part predates all other glacial drift.
- D1 Till and related diamicts, at or near land surface in areas of constructional topography, generally overlying ice-contact deposits; generally a compact stony sandy clayey silt containing some rounded stones; locally angular and rounded stones in a loess fine-sandy silt matrix, deposited during an ice readvance.
- D2 Pebbly silty clay; generally nonbedded, locally contains deformed wisps of silt
- Contact between geologic units, dashed where approximate

in many proglacial lakes throughout New York, and in the Lake Ontario lowland they constitute the uppermost unit over large areas, but in the Susquehanna River basin they are generally mantled by sand or gravel. Because the surficial deposits of ice readvance in Fivemile Creek valley are atypical, they are documented in the next section.

Deposits Resulting from Readvance

The lower sides and floor of Fivemile Creek valley are characterized by constructional topography and by slopes much less steep than those of the upper valley walls. Small terraces are found locally, but more commonly land surface is hummocky or irregular. Hummocky terrain is common on the floor and the lower sides of valleys in the Susquehanna River basin and is generally associated with ice-contact stratified drift. Nevertheless, in Fivemile Creek valley many shallow exposures and drainage features indicate that a diamict layer lies at or within a few feet of land surface all across the valley except for the lowest terraces and the floodplain.

Till

The surficial diamict north of Dineharts is interpreted as a readvance till because of its lithology and its ubiquitous presence atop older stratified drift. It is generally similar to upland till in that it has a matrix of clayey silt, although in several places it is more stony or contains more rounded stones than typical upland till. A few knolls consist of angular to rounded stones in a plentiful matrix of loose silty fine sand; this material may be redeposited ablation debris. Many shallow depressions are filled with water or swampy vegetation even though they are well above the nearest stream, which would not be true unless poorly permeable material such as till lay at shallow depth. East of the highway on the western side of the valley between Daball Corners (fig. 3C) and the Yates County line, till was exposed in several places on the lower valley walls, wet spots were common in April 1984, and small streams carried flow all the way to Fivemile Creek. (By contrast, on the eastern side of the valley several equally small streams went dry about 150 feet from where they began to cross a low stratified-drift terrace. Infiltration would not be so rapid if till were immediately below the stream channels, so the terrace is inferred to cap a post-readvance gravel deposit that is at least several feet thick.) Sites at which exposures or wells completed above bedrock demonstrate that sand or gravel underlie a near-surface till layer are listed in table 1.

Much of the area interpreted in figure 3 as readvance till overlying ice-contact deposits is mapped on the most recent Steuben County soils survey (French and others, 1978) as "Howard-Madrid complex". Although French and others describe this mapping unit as about half Howard soils (well-drained, formed on high-lime gravel) and half Madrid soils (well drained, formed on till), in the present study till was observed much more commonly than gravel at land surface. Furthermore, according to French and others (1978, p. 26), the till that constitutes the C-horizon of Madrid soils is generally underlain by gravel where the topography is "undulating," which includes most areas of these soils on the valley floor. Thus the soils survey provides some support for the concept of readvance in this valley.

Possible Subaqueous Diamict

From Dineharts south to Wheeler, diamict is widely observed at or very close to land surface (fig. 3A) but its lithology is rather different from that exposed north of Dineharts (fig. 3B). The stratigraphy shown in figure 4 is inferred from exposures in the bluffs along Fivemile Creek between Dineharts and Wheeler and from several well records, including some not shown in the figure.

The diamict (unit 3 in fig. 4) may be described more fully as follows: The predominant lithology is massive silty clay containing scattered angular to rounded pebbles, granules, and coarse sand that generally constitute from 1 to 5 percent by volume but locally from 10 to 15 percent. A few small lenses or masses of silty clay contain from 25 to 40 percent pebbles. Locally, the silty clay is nearly free of pebbles but contains blebs or severely deformed wisps of silt or clayey silt. No bedding can ordinarily be recognized, except for rare thin layers of sand or gravel. This diamict, although generally containing much less sand and stones than typical till, occupies the same stratigraphic interval as the readvance till north of Dineharts. The areas shown as pebbly silty clay (D2) in figure 3 correspond generally to areas mapped by French and others (1978) as Niagara and Collamer soils. These soils are silt loams and are described by French and others (1978, p. 20, p. 34) as formed on lake-laid silt, clay, and very fine sand.

The gravel layer termed "older outwash" (unit 2) in figure 4 is at least 20 feet thick in the bluff along Fivemile Creek, and is tapped by wells 0.5 mile to the east, near the intersection of Mitchellsville and

Table 1.--Evidence for till over stratified drift in Fivemile Creek valley

Location				Land- surface altitude (feet)	Description
Latitude	Longitude	Latitude	Longitude		
4227 25	7718 30+	1,345		Gravel soil; small pond and swamp. Well 40 feet deep, reported to penetrate hardpan over gravel; large yield.	
4228 30	7713 06	1,370		Roadcut is till; two wells reported to obtain water from gravel at 23 feet.	
4230 05	7716 33	1,335		Eight-foot roadcut is till, probably overlies silty gravel near base. Well 40 feet deep, ends in gravel. Earlier well reported to penetrate fine sand 60+ to 130+ feet; abandoned.	
4230 27	7716 09	1,355		Exposures 1,000 feet south, west: a few feet of gravel over till. House here: large rocks reported in cellar hole, poor drainage; well 30 feet deep, small yield from gravel.	
4230 33	7717 12	1,405		Ditch: alluvial gravel 0-5 feet. Dug well nearby, poor yield. Drilled well reported to penetrate gray gravel and clay, no water 0-35 feet (mostly till?); small yield from gravel or broken shale 35-38 feet, water level 13 feet.	
4231 16	7716 26	1,500		Sidehill pit: till and very stony diamicts 0-20 feet, cobble gravel 20-30 feet, sand and fine gravel 30-60 feet, tough unoxidized till at 60 feet.	
4231 17	7716 41	1,362		Till exposed 400 feet west and beyond. Two industrial and one municipal well, all screened in gravel between 60 and 100 feet in depth, each yields >100 gallons per minute.	
4232 31	7716 41	1,445		Till soils; drilled well ends in gravel at 23 feet.	
4232 51	7716 18	1,400		Streambank 17 feet high: till at 9 feet, sandy silty gravel at 16 feet.	
4233 06	7716 11	1,425		Till soils; stones and dirt (hard at 8 feet) and poor drainage reported in house foundation; well ends in gravel at 40 feet, large yield.	

Gardner Roads, where its thickness exceeds 36 feet. As suggested in figure 4, it may correlate with a unit described as "pebbly gumbo with water" in a well east of Renchans (4226 15 7719 09). On the other hand, wells 1000 feet east and west of Fivemile Creek near Renchans (4226 15 7719 09, 4226 22 7719 34) obtain water from gravel layers 3 to 4 feet thick, apparently within the pebbly clay diamict (unit 3). Both of these gravel layers lie at altitudes of about 1260 feet, the same as the top of the "older outwash" in the most downstream exposure along Fivemile Creek. Perhaps it is these thin gravel layers that should be interpreted as the southward continuation of unit 2. In either case, unit 2 apparently pinches out a short distance south of Renchans. Records of several wells near the intersection of Wheeler and LaRue Roads, 0.4 mile east of Wheeler (fig. 3A), indicate that about 25 feet of surficial gravel are underlain by 125 feet or more of fine-grained sediments.

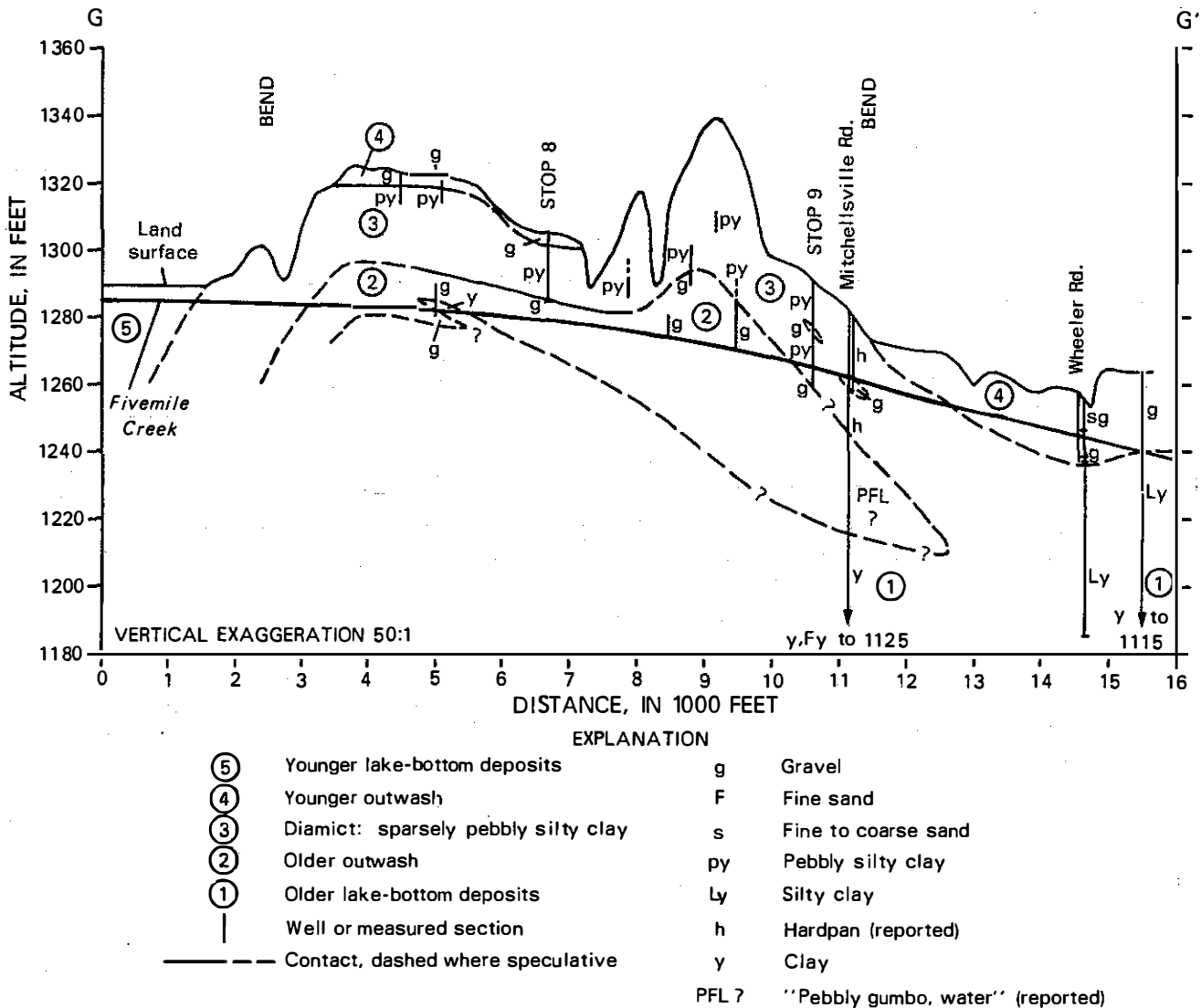


Figure 4. Geologic section near Renchans in Fivemile Creek valley. (Location of section shown in figure 3.)

The silty-clay diamict (unit 3) is interpreted to be the product of ice readvance into standing water south of Dineharts. Presumably the ice terminus was at least partly floating; perhaps the pebbles and sand could have been deposited from calved icebergs as well as from an attached floating ice tongue. This interpretation is based primarily on analogy with till in Buttermilk Creek valley in western New York, where the ice must have advanced into standing water and much of the till has a low pebble content and wisps of silt (LaFleur, 1979, 1980; Randall, 1980). Similar interpretations have been suggested by others for pebbly silty clays observed on field trips. However, P. J. Fleisher (State University of New York, written commun., 1986) reports that exposures of similar diamicts in the eastern part of the Susquehanna River basin are numerous and so widely distributed that interpreting them to be subaqueous readvance tills would unreasonably complicate deglacial history.

Ponded water in Fivemile Creek valley is easily explained by the geometry of nearby through valleys. The Hammondsport valley extends from Keuka Lake through Hammondsport to Bath, nearly parallel to Fivemile Creek valley (fig. 1). Because Hammondsport valley is much broader and, especially to the north, much deeper to bedrock than Fivemile Creek valley, presumably the tongue of active ice in Hammondsport valley would have advanced more rapidly. Once it reached the west wall of the Cohocton valley at Bath (fig. 1), it would have impeded or blocked the natural southward drainage of all tributaries north of Bath, including Fivemile Creek. If so, the resulting lake level must have been controlled either by a spillway now at 1,450 feet altitude between Campbell and Stocking Creeks (fig. 1), 8 miles south of the confluence of Fivemile Creek and Cohocton River at Kanona, or by the ice tongue at Bath, whichever was lower. Deltaic sand and gravel high on the valley wall near Bath and Kanona might date from such a lake ponded during readvance rather than during the previous retreat; beds commonly dip north or northwest at altitudes of 1,380 to 1,440 feet and are not known to be mantled by the pebbly silty clay or other lake-bottom deposits. Several feet of compact silty diamict containing about 20 percent pebbles, many of them rounded, are exposed atop gravel in a pit cut into a terrace at Mitchellsville, 3 miles southeast of Renchans. The form and altitude of the terrace suggests a delta deposited by meltwater flowing north from the Hammondsport valley through the Mitchellsville gorge (figure 3) but if so the surficial diamict implies that ice subsequently readvanced in Hammondsport valley at least far enough to block the Mitchellsville gorge.

Before readvance, a lake probably occupied the low-lying central and eastern parts of Fivemile Creek valley north of Dineharts, where older till-mantled stratified drift is absent or buried by younger sediments. The altitude of the highest part of the pebbly clay diamict exposed near Dineharts is 1,340 feet, not far below the spillway into Stocking Creek south of Kanona even if some allowance for postglacial rebound were added. Perhaps careful evaluation of plausible lake depths and lithology of the pebbly clay south of Dineharts might lead to a conclusion that incorporation of lacustrine sediment into grounded ice as it readvanced is a more plausible explanation of the origin of this diamict.

Denny and Lyford (1963, p. 14-17) observed diamicts overlying stratified drift at several sites in the Southern Tier of New York and in adjacent

Pennsylvania. These diamicts ranged in texture from rubble to silty clay loam that resembled upland till, were thickest near the valley wall, and were interpreted to be colluvium emplaced by mass movement. Colluvium may be found locally in Fivemile Creek valley, but the complex hummocky topography in some diamict-mantled areas, the absence of surficial diamict on younger terraces, and the thin fluvial gravel capping the diamict in some localities all require that emplacement of diamict ceased before ice and meltwater disappeared from the basin, an unlikely history for mass movement.

Lacustrine Deposits

The valley floor near Beans Station includes three principal geomorphic elements: a broad sandy terrace at an altitude of about 1,310 feet immediately southeast of Beans Station, a gently sloping plain extending southward another 1.5 miles, and the "Prattsburg muck," a huge drained swampland extending east about 2 miles. The Beans Station terrace is capped by pebbly coarse sand. A well near the center of the terrace (4228 55 7716 51) penetrated 300 feet of fine sand to silt that was interbedded with some medium-to-coarse sand to a depth of about 60 feet. West, south, and southeast of low scarps that bound the mapped extent of outwash, several holes drilled to depths of 70 to 110 feet penetrated only fine sand, silt, and clay. Near the west side of the valley floor a well (4229 01 7717 17) penetrated through fine-grained alluvium into sand that may be equivalent to the sand that forms the Beans Station terrace. All this information suggests the Beans Station terrace is an outwash delta, whose south and southeast margins (near the 1300 foot contour) may be the open-water delta front. While the delta was being deposited, ice blocks occupied parts of the valley floor to the east (the present Prattsburg muck), to the north (near Mud Lake and several small kettles), and probably to the west. The delta and the lake-bottom deposits overlap the older ice-contact deposits and readvance till along the west side of Fivemile Creek valley. Along the east side of the valley lake-bottom clay-silt rhythmites overlap till. Within 250 feet of the base of the till-covered slope the rhythmites are overlain by a few feet of fine to very fine sand, presumably a beach or shoreline deposit.

In most broad valleys of the Susquehanna River basin, low outwash or fluvial terraces of pebbly coarse sand or gravel mantle late-deglacial lacustrine deposits. The lack of late outwash in Fivemile Creek valley is a direct result of the fact that the valley is not quite a through valley: both the main valley (at Jubertown Swamp) and the principal spur (at Elmbois, east of Beans Station) head at till-covered saddles that were doubtless lowered by glacial erosion but nevertheless begin high on the side of the much deeper Hammondsport valley. Once ice ceased to flow into Fivemile Creek valley through these saddles or across the adjacent upland, meltwater from the north quickly was diverted to the deeper Hammondsport valley. Because the flow of meltwater was cut off so quickly, lake-bottom deposits south of Beans Station were never mantled with outwash, nor was there much sediment to fill the large kettleholes that developed as remaining stagnant ice blocks melted. The result was unusually abundant and extensive accumulations of peat (in Jubertown Swamp, a swamp northeast of Daball Corners, the Prattsburg muck east of Beans Station, and smaller swamps) and organic silt (along Fivemile Creek north and south of Beans Station). A lack of outwash and an abundance of large or coalesced ice-block depressions are also evident in broad non-through valleys in the

eastern part of the Susquehanna River basin (Fleisher and Cadwell, 1984, p. 194) and have also been attributed to glacier thinning over the divide and subsequent stagnation downvalley (Fleisher, 1986).

Regional Considerations

The readvance in Fivemile Creek valley proposed in this article seems consistent with drift borders in this region postulated by Connally (1964). He inferred, primarily on the basis of heavy mineral separations and distribution of constructional topography, that the surface drift in the northwestern part of the Chemung River basin was younger than the widespread "Olean" drift to the south. His younger drift was distinguished by predominance of purple over red garnet, many exotic pebbles, stream-rounded stones in till in the valleys, and sparsely stony till overlying contorted lacustrine deposits at "many" unspecified valley locations. Connally thought that the younger drift was probably of Kent age, and that its southern boundary lay somewhere near Kanona but "may prove to be correlative with one of the moraines in the Prattsburg (Fivemile Creek) valley". He described these moraines as located at Renchans and south of Prattsburg.

Multiple layers of till, attributed to repeated oscillations or readvances of the ice, have been recognized in two valleys immediately north of the Appalachian Plateau, near Dryden (T. Miller, U.S. Geological Survey, oral commun., 1985) and Herkimer (Ridge and others, 1984). In Fivemile Creek valley, logs of wells and test borings near Prattsburg (4231 17 7716 41, table 1) might be interpreted as having penetrated multiple layers of till. However, exposures and well records near Waldo Creek (4230 05 7716 33, table 1) are suggestive of a single layer of till capping an earlier delta. Generally, subsurface data in Fivemile Creek valley are inadequate as a basis for deciding whether more than two till layers are present.

INFLUENCE OF GLACIAL DRIFT ON POSTGLACIAL STREAMFLOW

The areal extent of surficial sand and gravel in a watershed is strongly correlated with streamflow draining from the watershed during the periods of minimum flow that recur in summer or early autumn each year and recur with greater severity in occasional drought years, as shown by studies in Connecticut (Thomas, 1966; Randall and others, 1966; Cervione and others, 1972; Cervione and others, 1982) and New York (Ku and others, 1975; Barnes, in press). The powerful influence of surficial sand and gravel on low flow is illustrated in figure 5, which represents 73 watersheds in the Susquehanna River basin. The percentage of watershed area underlain by sand and gravel is compared to a statistical index of low flow; namely, the 7-day low flow having a 10-year recurrence interval. The term "7-day low flow" refers to the mean flow averaged over the 7 consecutive days of lowest flow in a year. To state that a particular 7-day low flow has a 10-year recurrence interval means that over several decades, in 1 year out of 10 the mean flow for the 7 days of lowest flow in that year would be equal to or less than the value stated.

Regression analysis of the set of data plotted in figure 5 demonstrated that percentage of watershed area underlain by sand and gravel explains 86 percent of the variation in low flow per square mile of streams in the Susquehanna River basin (Ku and others, 1975, table 4). At least three

the greatest percentage reduction would be at low flow. However, three wells near the gaging station penetrated mostly poorly permeable fine-grained sediment below a depth of 20 feet. Furthermore, consistent gains in streamflow along both Fivemile Creek and the adjacent reach of the Cohocton River proportional to the area of sand and gravel were recorded by streamflow measurements in August 1982 and September 1983 (U.S. Geological Survey, 1983, 1984). Thus, no evidence of unusually large underflow near the Fivemile Creek gaging station has been obtained.

2) Irrigation pumpage. In the western part of the Susquehanna River basin, some farmers whose lands abut major streams have dug pits in or adjacent to stream channels and used portable pumps to occasionally withdraw water for irrigation. If substantial amounts of water had been pumped regularly from Fivemile Creek for irrigation over many years, the statistical indices of flow during the irrigation season would presumably have been less than natural flow by an amount equal to the typical rate of pumpage. However, interviews with current and former owners and managers of farms revealed that annual irrigation pumpage in Fivemile Creek valley was substantial only during 1966-69, and took place almost exclusively from June through mid-August, well before the annual minimum flow in nearly every year. Thus, irrigation pumpage seems to have had little or no effect on the 7-day 10-year low flow of Fivemile Creek.

3) Surficial geology. The area of sand and gravel within the watershed of Fivemile Creek was interpreted by Ku and others (1975) from a brief reconnaissance of the surficial geology in 1965 and an earlier soil survey (Pearson and others, 1931). Till and fine-grained lake-bottom sediments seemed to be more abundant in Fivemile Creek valley than in most others, and were presumed to be interbedded within the stratified drift, but no method was devised to represent this possibility in regression analysis. The geologic reexamination described in this article resulted in the hypothesis that for purposes of regression analysis all parts of the valley fill where fine-grained lake deposits or diamict are at or very close to land surface should perhaps be classified with till-mantled upland rather than with sand and gravel. These poorly permeable units are commonly underlain by sand or gravel that is tapped by wells and that comes in contact with streambeds southeast of Dineharts and perhaps elsewhere. However, water stored above stream grade in sand and gravel is the principal source of ground-water discharge at low flow, and is probably less abundant where shallow diamicts limit recharge than where sand and gravel extend from land surface to below stream grade. The diamicts are overlain by a few feet of outwash gravel in some places, but much of this thin mantle seems to be drained and unsaturated during low flow. Perhaps such areas would yield as little water at low flow as areas of till-mantled upland.

The area of surficial sand and gravel in Fivemile Creek watershed was recalculated in accordance with the foregoing hypothesis. Areas of surficial till, silt, and clay were excluded. The recalculation brought low flow predicted by regression analysis into much closer agreement with low flow calculated from streamflow records (table 2). Several areas mapped as sand and gravel in 1984 were not examined in the field, or no indication of surficial gravel thickness was obtained. Perhaps some of these areas should have been excluded, inasmuch as predicted low flow is still greater than observed (table 2).

Some confirmation of the hypothesis that till-mantled stratified drift should be treated as till in predicting low flow comes from other parts of the Susquehanna River basin. Along the west side of the West Branch Tioughnioga River valley near Homer, and along its tributaries Factory Brook and Cold Brook, rounded benches a few tens of feet above the valley floor seem to be capped by till, but gravel is exposed in a few deep excavations. The county soil survey (Sery, 1961) generally classified these benches as "Bath-Howard soils" or some similar hybrid of till and gravel soils. The surficial diamict may have been deposited atop older kame terraces during a readvance, as in Fivemile Creek valley, or may be the product of mass movements as described by Denny and Lyford (1963). These areas have been excluded from the stratified drift in regression analyses to date, and including them would increase the error in the prediction of low flows; hence, exclusion seems proper.

Table 2.--Effects of revised geologic interpretation on prediction of low streamflow, Fivemile Creek valley

Date of mapping	Drainage area (square miles)	Extent of sand and gravel			7-day 10-year low flow (cubic feet per second)	
		Basis	Area (square miles)	Percent of drainage area	observed ^{a/}	predicted ^{b/}
1968	66.8	Maximum	>14.8	>22.2	0.6	>5.1
	66.8	Average	14.4	21.6	0.6	4.9
	66.8	Minimum ^{c/}	14.0	21.	0.6	4.7
1984	66.8	Maximum	7.9	11.8	0.6	2.2
	66.8	Minimum ^{d/}	6.7	10.0	0.6	1.6

^{a/} For period of record through 1975; 0.7 for 1931-60.

^{b/} Predicted by best-fit provisional regression equation applied to set of data assembled by Ku and others (1975) and revised in 1983.

^{c/} Excludes areas where identification as sand or gravel seemed questionable.

^{d/} Excludes areas where thin surficial sand or gravel, thought to be less than 5 ft thick and unsaturated at times of low flow, overlies till or lake-bottom silt and clay. Both minimum and maximum exclude areas of surficial till, lake-bottom silt and clay, and postglacial organic-rich silt to very fine sand regardless of what may underlie them.

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LOG OF FIELD EXCURSION

Stops 1 and 2 show conditions in a typical valley of the Susquehanna River basin, to be contrasted later with conditions in Fivemile Creek basin. Stops 3-5 present evidence for till overlying stratified drift. Stop 6 and nearby hesitation stops overlook the divide at the head of Fivemile Creek. Stops 7-9 show diamict and other deposits in proglacial lakes during and after the inferred ice readvance. Stop 10 considers the hydrologic implications. The road log begins at the entrance to Buttermilk Falls Park along Route 13 just south of Ithaca.

Total Miles from
miles last point

- | | | |
|------|-----|--|
| 0.0 | 0.0 | Proceed south on Route 13. The route follows the valley of Cayuga Inlet, a trough greatly overdeepened by glacial erosion. Bedrock is more than 400 ft below the valley floor and is below sea level. |
| 2.0 | 2.0 | Junction with Route 96; continue south on Route 13. |
| 2.4 | .4 | View to left of continuation of the Cayuga Inlet trough. |
| 3.8 | 1.4 | Views to left across Cayuga Inlet trough, next 0.5 mile. |
| 5.0+ | 1.2 | Entering valley of West Branch Cayuga Inlet, a tributary valley also enlarged by glacial erosion, but not as deeply. |
| 7.5 | 2.5 | Turn right on paved road at sign "Sebring Rd Tavern" |
| 8.0 | .5 | Turn left at T junction; views left across valley next 1.1 mile. |
| 9.1 | 1.1 | HESITATION STOP, near crest of hill: view to left of hummoky valley fill; note absence of stream channel. |
| 9.5 | .4 | Park along road near barn bearing sign "Chuck's Marine Service".
STOP 1: VIEW OF THROUGH VALLEY. To the right, valley floor is relatively smooth, slopes gently southwest, and is underlain by outwash gravel. To the left, valley floor is hummoky, slopes irregularly northeast, and is underlain by the same outwash gravel, deposited in part over buried ice that did not melt until after the flow of meltwater and sediment cased. Most large streams in the Susquehanna River basin head in one or more through valleys like this: broad valleys that transect the drainage divide and served as a channel for ice during glaciation and for meltwater during |

deglaciation. Alluvial fans of small tributaries are visible at the base of the far valley wall; these streams flow southwest on the valley floor but lose water by seepage and are usually dry. The trees that form a line across the valley floor to the southwest border Carter Creek, a larger tributary that also loses water by seepage; the ground-water mound thus created near Carter Creek constitutes the ground-water divide in this valley.

- 9.6 .1 Junction with Route 13, continue straight ahead on Route 13.
 10.3 .7 Cross Carter Creek. For next 3 miles the route descends a valley train of outwash augmented by alluvial fans of tributaries.
 12.2 1.9 Cross Pony Hollow Creek.
 12.7 .5 Cayuta Road on left.
 12.9 .2 Cross dry channel leading from Hendershot Gulf, a meltwater spillway and incipient through valley.
 13.0 .3 Turn left on dirt road into gravel pit.
 13.3 .3 STOP 2: COARSE OUTWASH. The lower terraces in most broad valleys in the Susquehanna River basin are underlain by outwash gravel. The gravel exposed here is coarser and richer in clasts of local shale bedrock than most outwash, because it lies immediately downgradient from a remarkable 4-mile-long spillway cut in bedrock, from which many of the clasts were doubtless eroded.
 13.6 .3 Leave pit, turn left (south) on Route 13.
 14.8 1.2 Junction Routes 13 and 224. Proceed to Kanona, N.Y. following Route 13 south to junction with Route 17 near Elmira, then Route 17 west to Corning, Bath, and Kanona (about 53 miles).

Road log resumes at junction of Routes 17 and 53 at Kanona, N.Y.

- 0.0 0.0 Turn right (north) on Route 53; enter Fivemile Creek valley. For the next 0.8 mile, ice-contact stratified drift mantles the lower valley walls (both sides), up to more than 200 feet above road level.
 1.0 1.0 Cross Fivemile Creek.
 1.2 .2 Road is cut through "valley choker" moraine; test borings (Randall, 1972) suggest that till constitutes more than 50 percent of its mass; road borders north edge for the next mile. Fivemile Creek is briefly incised in bedrock at east side of valley.
 4.0 2.8 Wheeler; Route 53 is on outwash, overlapped by alluvial fans from the west; views to the left of fans next 0.3 mile.
 4.6 .6 Avoca-Wheeler Rd on left. For next 1.3 miles land surface is mostly a pebbly clay diamict (to be examined at stops 8 and 9)
 6.4 1.8 Large barn on left; shallow pond on right. Near pond, a well was reported to penetrate "hardpan" (till), then obtain a large yield from gravel at a depth of 40 feet.
 7.0 .6 Wetland to right of road.
 9.7 2.7 Turn right on Waldo Road.
 10.2 .5 Road intersection; continue 400 feet ahead and park.
 STOP 3: TILL-MANTLED STRATIFIED DRIFT. Topography, as shown on topographic map and as visible from this point, is suggestive of a kame terrace; a moderately level surface, hummoky in detail, 40 feet above the floodplain. At the trailer to the north, a well is finished in gravel at a

depth of 40 feet, after another well was abandoned at about 130 feet after penetrating many feet of fine sand. However, roadcuts 400 to 800 feet east and 400 feet north of the intersection reveal till.

- 10.2 .0 Turn left (north) at intersection.
- 10.8 .6 Turn left at T-junction onto Steuben County 74, Pultney Road
- 11.0 .2 Cross Fivemile Creek, bear right.
- 11.8 .8 Intersection; creamery visible ahead; park 100 ft beyond intersection. Walk 300 ft down driveway to the right (east). Small depressions north of driveway are kettles floored with peat.

STOP 4: TILL-MANTLED STATIFIED DRIFT. Till is exposed in a small excavation here. Wells in the two small buildings on the valley floor immediately to the southeast supply 150,000 gallons per day to the creamery; the village of Prattsburg obtains most of its water from a well 250 feet further south. Drillers' logs of these wells and nearby test holes differ in detail but all indicate that silty or clayey gravels and stony clay (or hardpan) predominate. The well nearest the creek reportedly penetrated

- 0-13 feet soft clay
- 13-32 feet boulders, clay hardpan
- 32-66 feet mostly gravel, silty, yields some water
- 66-72 feet gravel, yields more water
- 72-89 feet hardpan.

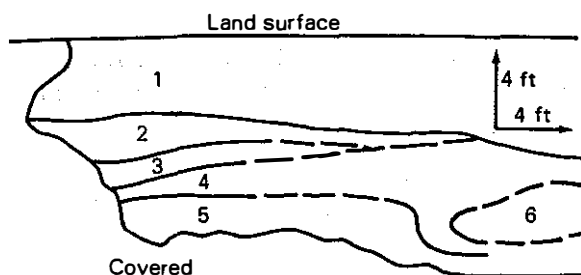
Walk 300 feet downstream along Fivemile Creek to view exposure of post-till gravel on opposite bank. Return to main road.

- 11.8 .0 Turn around, head south on Mill St. (Pultney Road).
 - 12.0 .2 Turn left on paved road; pass Narcissa Prentiss house.
 - 12.2 .2 Cross Fivemile Creek, then turn left. Cemetery on left.
 - 12.45 .25 Turn right into gravel driveway.
 - 12.6 .15 Prattsburg Town Highway Dept. garage.
- STOP 5: TILL-MANTLED STATIFIED DRIFT. High excavation to east was described in 1984 as follows:
- 1500-1480 feet altitude: includes definite till and much problematic diamict (>50% stones in a till-like matrix).
 - 1480-1470 Cobble gravel.
 - 1470- Layered sands, clean to gummy from interstitial clay, and fine gravels.
 - 1440- Till, tough, unoxidized.
 - 1400 Base of slope, garage level
- Top of cut is 150 feet west of base of steep valley wall.
- 12.6 .0 Turn around, head out driveway.
 - 12.75 .15 Turn left on paved road. Note swamp on left at corner.
 - 13.0 .25 Turn right at intersection. Note swamps and pond on left and ahead.
 - 13.2 .2 Turn right on Mill St.
 - 13.4 .2 Creamery and stop 4; turn left on Mechanic St.
 - 13.7 .3 Turn right on North Main St., pass Prattsburg village green on left.
 - 15.3 1.6 Paved road to right. HESITATION STOP: View to right across upper Fivemile Creek Valley. Exposure of till over gravel in woods ahead, 1000 ft east of road. Flat valley floor is underlain by peat.

- 17.9 2.6 Enter Yates County, town of Italy.
- 18.2 0.3 HESITATION STOP: View to right across Jubertown Swamp (head of Fivemile Creek). Skyline beyond swamp is not marked by deeply incised saddle or through valley.
- 19.6 1.4 20 mph turn, Blue Eagle tavern; bear right.
- 21.3 1.7 Turn right on Prosser Road
- 21.5 .2 Crest of hill. HESITATION STOP: Views to right (northwest) and to the rear (north) of gentle saddles on basin divide.
- 22.3 .8 Turn left on Stever Hill Road; follow this road as it crosses saddle on basin divide.
- 22.9 .6 Crest of saddle.
STOP 6: VIEW OF BASIN DIVIDE: This saddle, the lowest on the Fivemile Creek divide, shows no incision or deposition due to meltwater flow. The slope ahead (east) descends to Keuka Lake, 730 feet below this saddle. The deeper valley presumably captured most ice and meltwater flow when the ice surface was above this saddle, and certainly when it was below. Turn around and retrace route to Prattsburg.
- 32.3 9.4 Prattsburg Village Green; turn right, then turn left on Route 53.
- 33.1 .8 Swampy depression beside road on right.
- 33.3 .2 Side road on left. For next 0.2 mile, land surface is underlain by 5 feet of flatstone gravel, an alluvial fan of a brook.
- 33.5 .2 Cayward Hill Road on right. Till exposed at base of knoll on right, overlain by 10 feet of silty gravel, fine sand, and sandy gravel, presumably late-glacial inwash.
- 34.0 .5 Turn left on Waldo Road.
- 34.25 .25 Cross Fivemile Creek. Floodplain downstream from here is underlain by silt, clay, peat, and organic-rich muck that extends at least 5 to 15 feet below stream grade.
- 34.5 .25 Road intersection and stop 3; continue straight.
- 35.2 .7 Turn 160° right.
- 35.55 .35 HESITATION STOP: View of "Prattsburg Muck". This large ice-block depression, underlain by many feet of peat, is drained by buried tiles to perimeter ditches, which drain (or, if necessary, are pumped) into Mud Lake and thence into Fivemile Creek. Crops of lettuce and onions are grown here. The muckland occupies most of a broad tributary valley that heads at a gentle saddle overlooking the Keuka Lake-Hammondsport valley, like the saddle at stop 6.
- 36.2 .55 Turn left (southeast) at T-junction. Road crosses the BS outwash delta. Surficial sediment is pebbly medium to coarse sand. A well drilled in 1949 near the buildings south of the junction was abandoned at a depth of 320 feet, after penetrating mostly "quicksand" (silt to very fine sand) and clay.
- 36.6 .4 Turn right on gravel road. A well drilled here in 1983 was abandoned at 70 feet depth after penetrating only clay and silt.
- 38.1 1.5 Cleared cropland to right.
STOP 7: LAKE-BOTTOM DEPOSITS. Auger holes 270 and 820 feet west of road penetrated layered silt and clay, capped (near the road) by very fine sand.
- 39.2 1.1 Road bends sharply right.
- 39.3 .1 Pond and swamps on left.

- 39.5 .2 Turn left and stop. Walk along far side of the field west of the road to a small exposure in the bluff above Fivemile Creek.

STOP 8: PEBBLY CLAY DIAMICT. This exposure is plotted on figure 4. The sketch shows its appearance in 1983.



1. Pebble-cobble gravel; sandy streaks, silty at base
2. Silty very fine sand.
3. Fine to very fine sand, interbedded with medium to very coarse sand.
4. Oxidized massive silty clay with sparse pebbles.
5. Unoxidized massive silty clay with sparse pebbles.
6. Lens of silty pebble gravel. Pebbles numerous in underlying clay.

- 40.3 .8 Turn right at T-junction on Mitchellville Road. View left, at turn, of outlet of former spillway from Mitchellville.

- 40.5 .2 Wells nearby are finished in gravel beneath surficial pebbly clay.

- 41.05 .55 Approaching Fivemile Creek. Park along road, walk 1000 feet upstream (right) along edge of field to exposure in bluff.

STOP 9: PEBBLY CLAY DIAMICT. This exposure is plotted in figure 4. It was described in 1986 as follows

0-12 ft Clay or silty clay, containing 1 to rarely 5 percent rounded to sharply angular pebbles and coarse sand; no regular bedding recognized, but a few deformed streaks or blebs of coarse silt observed; oxidized.

12-15 ft Pebble-cobble gravel, rounded, a few exotics, generally very silty, appears to grade northward into very pebbly clay.

15-20 ft (exposed) 20-30 ft (augered) Sparsely pebbly clay, like that above but unoxidized, plastic.

30 ft Gravel; could not penetrate.

Property owner reports accelerated recession of this bluff 1984-86 since channel of tributary entering from the west was excavated for flood control, which owner believes has caused deflection of the flow of Fivemile Creek against this bank.

- 41.25 .2 Turn left on Route 53.

- 45.1 3.85 Cross Fivemile Creek.

- 45.5 .4 John Walsh Sales on right; turn right on gravel road.

- 45.7 .2 Approaching Fivemile Creek.

STOP 10: FIVEMILE CREEK GAGING STATION. Records of stage and flow have been collected here since February 1937. The operation of the station will be explained, and the influence of surficial geology on the minimum flow measured here will be discussed.

- 45.9 .2 Return to Route 53, turn right.

- 46.5 .6 Intersection with Route 17. End of Road Log.

GEOMORPHOLOGY OF THE CAYUGA LAKE BASIN

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GEOMORPHOLOGY

The Finger Lakes Region of Central New York is justly famous for two aspects of its geology: the Devonian stratigraphy and the Quaternary geomorphology. Not as well appreciated is the fact that the Quaternary landscape is the result of 360 million years of post-Devonian erosional history, for which no stratigraphic record is available in the region. An enormous unconformity everywhere separates glacial drift of late, or at the oldest, middle Pleistocene age, from the underlying lithified and mildly deformed Devonian marine strata. Volumes of sediment on the North Atlantic continental shelf and rise imply at least 2 km of regional denudation in the Cenozoic Era (Mathews, 1975). The remaining Paleozoic section in the Finger Lakes region is little more than 2 km thick, so at least half of the depositional section is gone. With a regional southward dip of about 1 per cent, 2 km of vertical denudation involved 200 km of homoclinal shifting of the present north-facing escarpments. The landscape viewed from the hill tops around Ithaca has a grand story to tell, when we can learn to listen.

The key to the geomorphology of the Finger Lakes Region lies in the geometry of deposition, deformation, and denudation. The Upper Devonian sedimentary facies were deposited in an epicontinental sea with a rising source area to the east; facies boundaries trend northeast to southwest, with Catskill facies alluvial-plain redbeds to the east, nearshore marine sandstones next to the west, and these grading westward into shales. Post-depositional regional deformation gently folded the Devonian rocks along fold axes trending N70°E, significantly shortened the sedimentary pile in a N-S direction (Engelder, this volume, and references therein), and regionally tilted the pile southward about one-half degree (1 per cent, 10 m/km, or 50 ft/mi).

The upper Paleozoic rocks of New York State record no significant source area to the north. The exposure of the Canadian Shield and Adirondack Highlands is therefore post-Devonian, although the former northward extent of Devonian sedimentation onto the continental platform is unknown. By Cenozoic time, the rivers of eastern North America were probably adjusted to structure on a regional scale. An intercueta lowland (now Lake Ontario) should have followed the Ordovician shale belt between the shield and the Lockport dolomite escarpment. An ancestral Hudson River probably was eroding headward along the Hudson Valley shale belt at the base of the Shawangunk and Catskill escarpments, and at its head ancestral Mohawk and Lake George lowlands probably were forming on shale belts broadly concentric to the Adirondack highlands. If the present is a key to the past, regional erosional denudation was

accomplished primarily by homoclinal shifting of subsequent (strike-oriented) rivers along shale lowlands.

On the southern dip slopes of Upper Devonian shales, siltstones, and sandstones in central New York, regional dendritic south-flowing rivers can be reasonably inferred. River gradients should have been significantly less than the regional dip, so dipping strata would be truncated by erosion in progressively younger belts from north to south. Kindle (1909, Fig. 21) noted long ago that the highest area between Cayuga Lake and Seneca Lake (Connecticut Hill in Enfield, height 2099 ft) is a synclinal ridge, demonstrating topographic inversion of relief even on the gentle structures of the region.

Several fossil peneplains have been proposed for the uplands of the Finger Lakes Region (Cole, 1938, 1941; Fridley, 1929). The evidence is fragmentary, but a reductionist logic still permits: (1) a structure-beveling ancient surface of low relief (and possibly near sea-level) now represented by the numerous summits in the region that range in height between 1800 and 2000 feet above sea level; (2) a lower structurally controlled surface primarily on shales at about 1000-1200 feet above sea level, and (3) various erosional levels within valleys below the two regionally correlative upland surfaces. Ignoring several generations of scholarly research, we can suppose that the gentle regional uplift that created south-draining dip-slope river systems also initiated the subsequent river systems along shale belts and started a long history of homoclinal shifting and lateral migration of divides toward the south. Ever since an early stage of regional erosion, north-flowing escarpment streams would have had steeper gradients than their south-flowing dip-slope counterparts, and the cuestas should have been migrating southward.

Probably by late Cenozoic time, the region had evolved to a "broad valley" stage, with valley floors graded to a regional level now 900-1000 feet above sea level. Uplift (Cenozoic tectonism in New York?) rejuvenated the rivers, and a "deep stage" of entrenched inner valleys resulted, especially along the axes of the north-flowing escarpment streams (von Engel, 1961, p. 15). These valleys became the precursors of the Finger Lakes. Presumably the rejuvenation caused more rapid divide migration toward the south. It has long been noted that barbed tributaries are common in the Finger Lakes, with acute junction angles of drainage systems pointing south but now draining north. Near Ithaca, both Salmon Creek and Fall Creek show this pattern; Keuka Lake is a Y-shaped glacially eroded lake basin that outlines a pair of southward-merging ancestral river valleys, although it now drains north to the St. Lawrence along with all the other Finger Lakes.

Glacial erosion and deposition has massively modified the ancestral fluvial systems that drain northward, converting north-trending valleys into glacial troughs or rock basins, many of which contain finger lakes. However, the glacial lowering of summit levels was minor (Muller, 1963, p. 236), and the position of the regional St. Lawrence-Susquehanna divide may not have shifted very much by glacial processes. The evidence is weak, but the zone of highest summits across New York south of the Finger Lakes usually crosses the intervening valleys within a few miles of the present

divide in each valley. The divide regions were reamed out by glacial erosion and associated proglacial and ice-marginal meltwater drainage, creating the famous "through valleys" that are the distinctive geomorphic feature of the region, much more so than the trough lakes, which are common in glaciated regions (Tarr, 1905, p. 233). One brief published report concluded that the massive Valley Heads moraine at Tully, New York, which forms the surface-water divide between Onondaga Creek to the north and the Tioughnioga system to the south by a barrier with 600 feet of relief on its northern, proximal, face, may rest on a bedrock sill at a depth of only about 220 feet (Durham, 1958). The rivers that drain south from the divide flow on valley trains that are not deep over bedrock. Their valleys were straightened by glacial erosion, but not much deepened. As most Valley Heads moraines in central New York crest at about 1200-1400 feet above sea level, the inferred bedrock sill beneath them may well be at about the 1000-ft. level that characterized the preglacial heads of the south-flowing drainage systems.

The hanging valleys and gorgeous gorges of the Finger Lakes Region are the result of glacial overdeepening of the main north-draining valleys that had notched into the northern edge of the Allegheny Plateau in preglacial time. The lateral (east-west) tributaries lay athwart the dominant direction of ice flow and were subject more to glacial and glacio-fluvial deposition than to glacial erosion. The series of valleys that converge at Ithaca (Fig. 1) make the point very nicely: north-oriented Inlet Valley and the Cayuga Lake trough are eroded down to below sea level; northwest-trending Sixmile Creek Valley and Cascadilla Creek Valley ("Ellis Hollow") have deep glacial-lacustrine fills on their floors but expose bedrock in places at about 950 feet above sea level; Fall Creek, which flows west, has a few areas of exposed bedrock at about 900-950 feet above sea level but masses of glacial drift fill most of the preglacial valley. The depth of glacial erosion seems directly related to the degree that each valley funneled the advancing tongues of successive continental ice sheets.

How many glaciers have covered Ithaca? From the deep sea oxygen-isotope record, one can safely infer 20 ice ages of similar intensity to the last one, each lasting about 100,000 years. The classic 4-fold midwestern glacial subdivision of the Pleistocene is as obsolete as pre-plate tectonics land bridges and Atlantis, but the Finger Lakes Region has revealed little of its glacial past. A pre-Wisconsin valley fill at "Fernbank" on the west side of Cayuga Lake demonstrates that a gorge had been eroded previously, probably by a minor tributary that had incised the valley side after an earlier glacial event had deepened Cayuga trough (Bloom, 1967). Therefore, at least two and possibly three glacial events can be inferred, depending on some assumptions. "Old", noncalcareous glacial drift is exposed at times along Sixmile Creek and at the head of Beebe Lake. If this drift is pre-Wisconsin, then the valleys around Ithaca were already as big as they are at the present prior to two ice ages ago. To the extent that their shape is due to glacial erosion, at least a previous third ice age must be inferred. The multi-reflecting subbottom seismic profile of Seneca Lake (Fig. 2) also implies that a long history of glacial erosion and deposition shaped the present catenary cross section of the Seneca Lake basin.

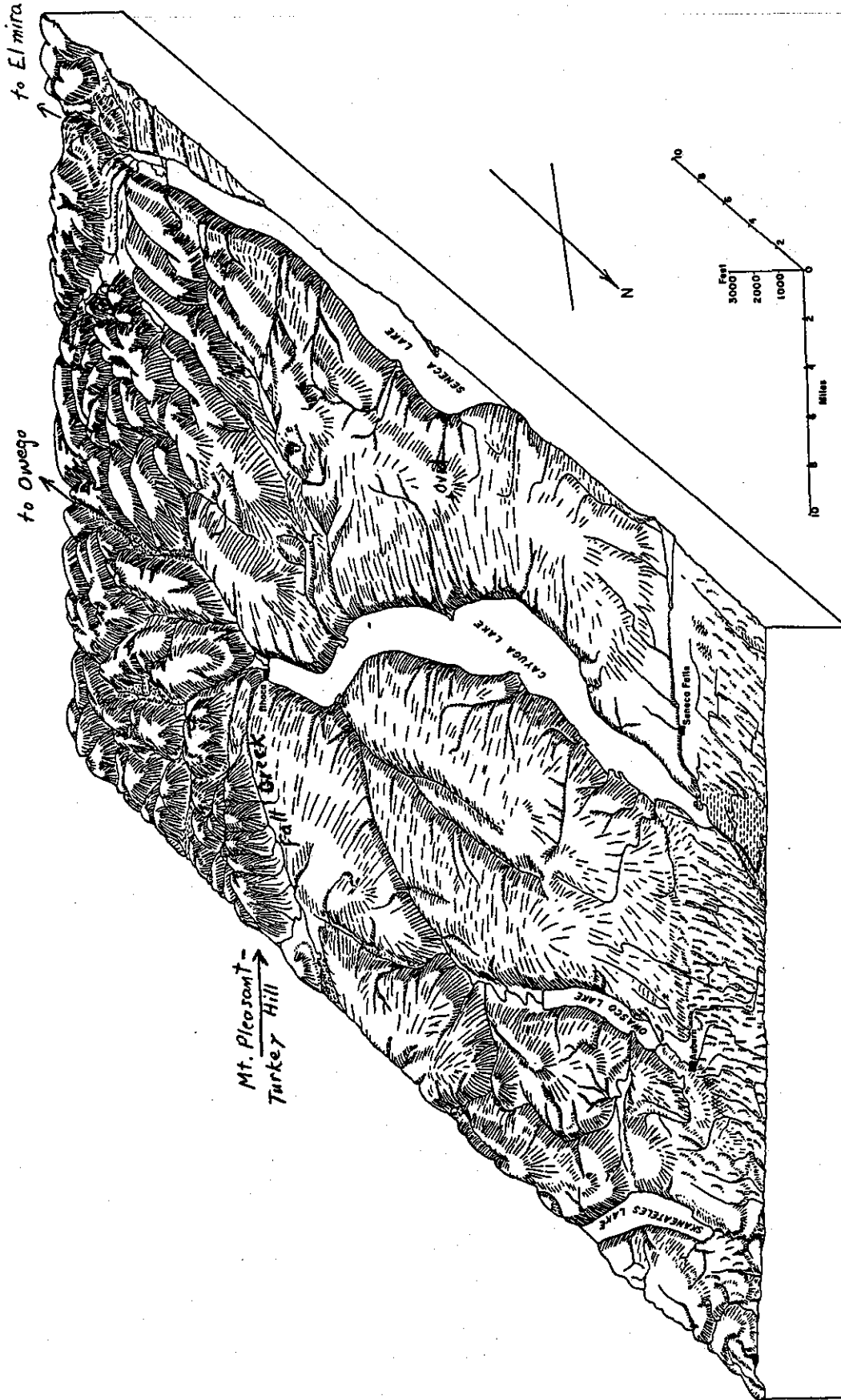


Figure 1. Physiographic diagram, Cayuga Lake basin (source unknown).

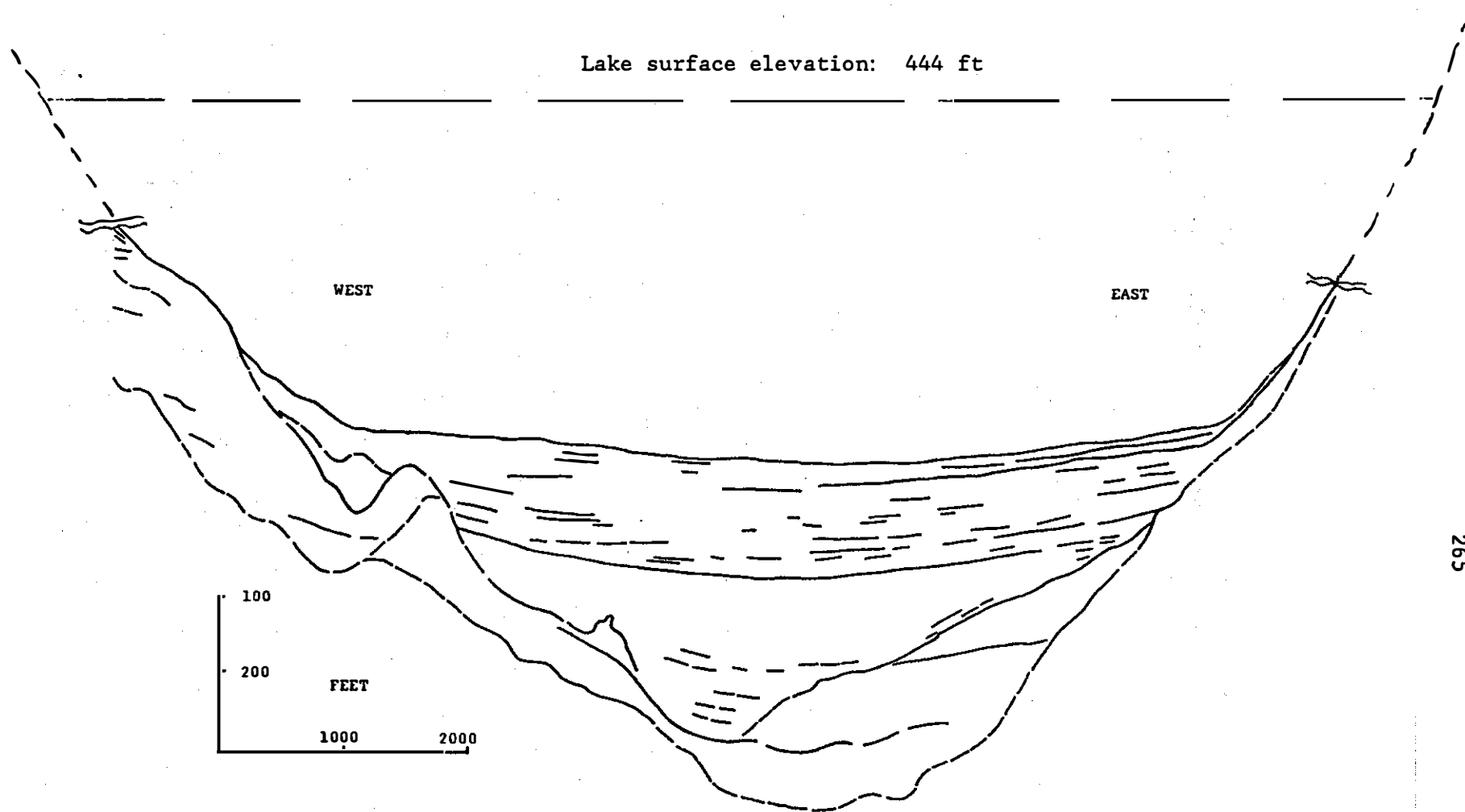


Figure 2. A seismic reflection profile across Seneca Lake based on data from a study for the Naval Research Laboratory. Maximum depth of the lake is 475 ft along this profile. Deep reflections are from sediment layers at least 500 ft below sea level. Vertical exaggeration X7 (adapted from a diagram by C. Windisch, in Woodrow, et al., 1969).

As we approach the present, the record of geomorphic evolution of the Finger Lakes Region becomes more readable. Most of the tributary valleys of Cayuga Lake near Ithaca were aggraded by a variety of glacio-fluvial and glacial-lacustrine deposits as the Wisconsin ice sheet advanced southward against the regional drainage. Rising proglacial lakes overflowed to the south across the divides, but most of the sediment remained within these great settling basins. Schmidt (1947) counted approximately 1000 annual couplets (varves) in Sixmile Creek Valley, grouped in four series that progressively thicken upward and show an increase in the thickness of the winter layer. The height of these deposits above the Cayuga Valley floor suggests that an ice lobe completely blocked the north end of the lake; the 1000 layers, if annual, suggest that the ice advanced about 40 miles (65 km) in 1000 years or about 65 m/yr. A finite radiocarbon date of 41,900 years from near the base of Schmidt's varve sequence has been correlated with a post-Plum Point Interstadial ice advance by Dreimanis and Goldthwait (1975). This date and several other "dead" radiocarbon analyses confirm that at least one mid-Wisconsin ice advance reached or affected the Finger Lakes Region. Little more can be said, now.

The final push of Wisconsin ice crossed Ithaca and moved south to the vicinity of Williamsport, PA. The age of that terminal position is still debated, but 17,000 to 19,000 years will be my educated guess until my southern colleagues agree on each other's evidence. The much favored relationship for the gradient at the edge of ice sheets, $h=4.7/d$ (h = ice thickness in meters, d = distance into the ice sheet from its margin, also in meters) yields a thickness of 1880 m of ice over Ithaca when the ice margin was 160 km to the south near Williamsport. Certainly, at this stage all the topography around Ithaca (maximum relief 2000 ft or 600 m) was deeply buried by ice.

By 13,000-14,000 years ago the ice margin had retreated to the regional divide south of the Finger Lakes. Here the ice edge paused or fluctuated, fed by a thickness of 1000 m or more of ice moving uphill toward the divide from the north, but unable to sustain glacier tongues down the Susquehanna system valleys. Here it built the Valley Heads moraine system, a series of massive morainic debris piles that choke the floors of the through valleys and with few exceptions determine the modern surface divide between the Susquehanna and St. Lawrence Rivers. With the progressive retreat of the ice margin northward from the Valley Heads moraine systems, the ice margin became more and more lobate and confined to the valleys. The South Cortland moraine, for example, was made by a tributary sublobe of the Cayuga Lake lobe that flowed northeastward for at least 12 miles from its base at Ithaca. The outwash from the South Cortland moraine flowed northeast and east to Cortland, where it merged with the valley train down the Tioghnioaga Valley from the Tully moraine. Near-contemporaneity of several of the Valley Heads moraines in adjacent valleys can be demonstrated by similar relationships.

Once north of the Valley Heads position, the retreating lobate ice margin terminated in proglacial lakes in all the valleys. All but the suspended fine silt and clay-sized sediment was trapped in the lakes. Notable around Ithaca is the pink or brown colors of the post Valley Heads

lacustrine beds. The color is derived from the Silurian red shale belt across the north end of Cayuga Lake, fully 40 miles away. The reddish sediments contrast sharply with the local sediment derived from gray-colored source rocks, so the abandonment of a south-flowing meltwater channel in favor of a lower western or eastern outlet, such as the rock gorges at Syracuse, is marked in the Ithaca region by an abrupt change from reddish to grayish sediments. The change is usually at a shallow depth, sometimes almost within the modern soil profile.

Deglaciation of the Cayuga Basin was even more rapid than earlier generations of geologists suspected (Fullerton, 1980). If the Valley Heads moraines are 13,000-14,000 years old and the St. Lawrence lowland was ice-free and open to the Champlain Sea by 12000 years ago, less than 2000 years were required to deglaciate the northern two-thirds of New York State! Some 10 or 15 names have been given to the sequence of hanging deltas in the Cayuga basin that record the episodic drop of proglacial lake levels, so each delta must have been built within a century or two. While the internal structure of the hanging deltas, with thick-bedded foreset gravel layers at angle of repose and only minor bottomset and subaerial topset beds, confirms rapid progradational growth, probably no geologist of the pre-radiocarbon era would have attributed their formation to a century time scale.

The river terraces and abandoned meanders along Fall Creek near Ithaca tell a similar story of rapid evolution. The oldest and highest surfaces record deposition or erosion in a variety of ice marginal streams or lakes, but with the integration of local lakes into glacial Lake Ithaca (outlet elevation 980 ft), lacustrine sediments and related hanging deltas and river terraces become reasonably correlative. A series of terraces near Varna record the dissection of the floor of local Freeville-Dryden Lake (elevation 1060 ft.) by Fall Creek as the creek cut down to the level of glacial Lake Ithaca and lower levels. The terraces should project down-valley into the surfaces of the appropriate hanging deltas, but the uncertainty of the channel gradients and the rapidity of changing levels make detailed correlation unlikely. The most significant fact is that all the terraces and hanging deltas were cut or built within one or two thousand years, at least 12,000 years ago. Soil profiles on the surfaces should differ more because of parent material, vegetation, and slope than because of climate or time. All the terraces' surfaces around Ithaca probably had soils forming through the late-glacial and postglacial times of the tundra (?), spruce, pine, and hardwood forest succession.

As the late-glacial gorge entrenchment encountered buried bedrock spurs and former valley floors, further downcutting was severely inhibited. Postglacial gorge cutting has proceeded downward for a few tens of meters and headward for some hundreds of meters, but most of our landscape has probably not changed much for 12,000 years. Even the impressive headward retreat of Fall Creek at Ithaca Falls, Enfield Creek at Lucifer Falls, or Taughannock Creek at Taughannock Falls may have been accomplished primarily by the re-excavation of preexisting valley fills left by earlier glaciations. The landscape we see on excursions around Ithaca was shaped in a brief interval of intense activity and rapid change about 14,000-12,000 years ago. Mastodons and Paleo-Indians walked on the same landscape that we now stroll.

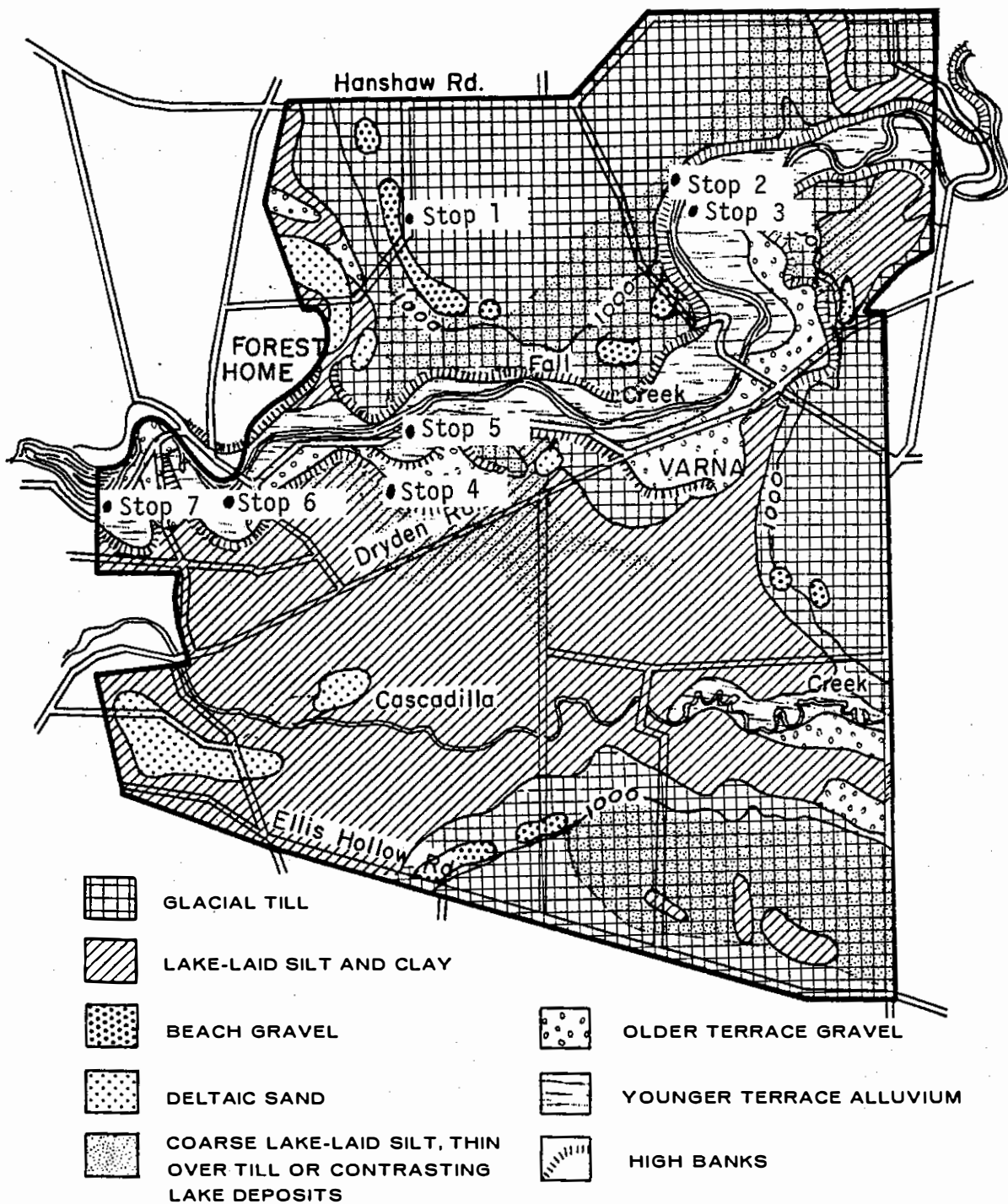


Figure 3. Locations of stops 1-7, and the parent materials of soils east of the Cornell campus (Cline and Bloom, 1965, Fig. 1).

LOCAL DESCRIPTIONS (SEE FOLLOWING ROAD LOG FOR ROUTES)

Stop 1. Lake Ithaca beach ridges near the Equine Research Laboratory.

In the course of making a detailed soils map of the Cornell campus area at a scale of 6.7 inches to 1 mile (-1:9500), Professor M.G. Cline and his students defined several areas with narrow ridges of well-sorted pebble gravel that are uphill from typical lacustrine silt-clay units of glacial lake Ithaca, and are downhill from non-lacustrine stony till soils (Cline and Bloom, 1965). The site along Bluegrass Road (Fig. 3) is typical. The pebbly soils are well drained, fertile, and warm, and have been allocated to small garden plots for Cornell staff and students. Downslope in the Cornell fields is a poorly drained belt of wet, cold soils, and across Bluegrass Road to the east is typical thin upland till, probably only a meter or less deep to bedrock. The gravel ridge was interpreted as a beach ridge of glacial Lake Ithaca, correlative with the 980-ft overflow south of Brooktondale at Belle School Road (Stop 9). A similar setting and soils sequence crosses Ellis Hollow Road and Game Farm Road along the base of Snyder Hill.

The beach ridge here is at an elevation of about 1020 ft; the similar ones in Ellis Hollow are about 1000 ft above sea level. The two beach ridge localities are 8 miles and 6 miles north of the 980-foot overflow, so the present gradient of the Lake Ithaca shoreline rises northward at 3 to 5 ft/mile. Possible causes include: (1) downcutting of the outlet during the life of glacial Lake Ithaca; (2) hydraulic gradient southward through the outlet (unlikely for a large, deep lake); (3) gravitational attraction of the water body northward toward the still extant ice mass; or (4) postglacial isostatic recovery increasing to the north. The tilt was aggressively debated by earlier geologists. Professor O.D. von Engeln gently referred to "a cult among geologists interested in the high levels of proglacial lakes which adheres to the concept that a postglacial northward uplift of the land made the original shore lines of the lakes rise toward the north" (von Engeln, 1961, p. 93). He went on to argue correctly that the correlation of delta tops from one tributary valley to another is much too indefinite to justify calculating a tilt. Perhaps he would soften his view if he could see the evidence we see today (or perhaps he would not!).

Stop 2. Top of the Varna high bank.

Fall Creek valley at this cross section is filled by a variety of Pleistocene sediments to be viewed in better perspective from stop 3. During the detailed soil mapping of the Cornell properties, Prof. Cline and his students noted the silt enrichment of the soil profiles in the fields north of this locality, and tentatively attributed it to an influx of eolian silt and sand in late glacial time. To their credit, the soils mappers noted that their eolian hypothesis did not fit the observation that the silt enrichment was found only below the 1060-ft contour. As long ago as 1934, H.L. Fairchild had named a shallow ice marginal lake called Freeville-Dryden Lake in the Fall Creek valley, which overflowed south around the west end of Mt. Pleasant into Cascadilla Lake and through the southeast end of Ellis Hollow (along Thomas Road, southeast corner of the Ithaca East topographic map.) The overflow is just above 1060 ft

(Fairchild's contour maps were in error), so Freeville-Dryden Lake must have been quite shallow and muddy, and it probably provided a minimal gradient for the ancestral Fall Creek when an ice wall to the west finally calved and drained Freeville-Dryden Lake down from 1060 ft into Lake Ithaca (in this region, at or below 1020 ft). The lowering of the local base level probably initiated a headward-migrating nick point across the exposed floor of Freeville-Dryden Lake and initiated the postglacial evolution of Fall Creek valley.

Stop 3. Base of the Varna high bank.

The high banks along Fall Creek north of Varna show the glacial stratigraphy. The lower half of the undercut bank over 100 feet in height exposes poorly sorted, crudely stratified sand and gravel. About 90 percent of the pebbles in the gravel are sandstone and shale of local derivation, and about 10 percent are limestone and crystalline erratics from the north. The sand, silt and clay matrix of the gravel is strongly calcareous. This stratified sand and gravel records the damming of lower Fall Creek by ice spreading eastward out of the Cayuga trough, while the headwaters of the creek were still ice-free.

Overlying the sand and gravel is about 40 feet of compact till that records the advance of ice up or across Fall Creek valley. Only about 70 percent of the till pebbles are of local origin, and most of the remaining 30 percent are limestone or dolomite. The tough, blue-gray matrix of the till is strongly calcareous. A thin layer of lacustrine sediments caps the oxidized upper part of the till.

As the succession of proglacial lakes in the Cayuga trough gradually fell to the level of present Cayuga Lake, Fall Creek has energetically re-excavated its interglacial valley. The thin cap of silt from Freeville-Dryden Lake was cut through while Lake Ithaca still drained through Willseyville Creek. Subsequently, Fall Creek established its postglacial course down the side of the Cayuga trough, soon to become superposed across buried rock spurs to give the succession of gorges and falls along the north edge of the campus. North of Varna, Fall Creek has not yet exposed its former rock floor.

Most of the cross-sectional area of the Fall Creek valley fill seems to be stratified sand and gravel. Roughly, the cross-section area of the modern valley on this transect is one-half as wide and about one-half as deep as the interglacial valley. Further downcutting has been inhibited by superposition on buried rock farther downstream (stop 5), but lateral migration has been extensive. The "Varna moraine" of Tarr (1909b, p. 151) and von Engel after him, is only a cut bank of the valley fill. The "proximal" face of the "moraine" exposes a layered stratigraphy similar to that in the high bank.

In Fall Creek valley near Beebe Lake, and at several places along Six-Mile Creek, are exposures of non-calcareous, clay rich diamictons in which the few remaining granite gneiss pebbles and cobbles are totally "rotten". It seems clear that most of the valley cutting in bedrock around Ithaca predates at least two glaciations, although most of the earlier drift filling was removed and replaced by calcareous drift of

probable Wisconsin age. An interesting applied aspect of the layered valley fill is that any dams across Fall Creek or other rivers in the Cayuga basin would involve a high risk of leaking through the coarse alluvial members of the glacial drift in the ancestral valleys.

Stop 4. Water tanks, east edge of campus in Newman Arboretum

From this stop at the 970 ft contour, the continued late-glacial and postglacial evolution of Fall Creek valley is easily viewed. The water tanks are on the highest point of the campus, below the water level on a fan-delta that Fall Creek built into glacial Lake Ithaca. A soil pit in the vicinity show a sandy loam, probably subaqueous. Other slightly higher parts of the surface to the east may be underlain by subaerial topset beds. Broad, shallow channels radiate down the delta slope from the apex.

Far on the southwestern skyline is a notch which is said to have puzzled Professor R.S. Tarr and others since. Is it a windgap recording the ancestral course of Fall Creek to the southwest, prior to capture by a north-draining escarpment stream? In part it is an optical illusion formed by a low ridge profiled against adjacent higher hills, but the location is intriguing.

Northeast of the water tanks is the Newman Arboretum, being developed as a major regional collection of native trees and shrubs. The Arboretum crosses the slopes of an abandoned meander of Fall Creek with a floor at about 900 feet, the regional level of glacial Lake Newberry. Lake Newberry drained past the south end of Seneca Lake toward Elmira, and was a very extensive ice-marginal lake across central New York. A smaller meander terrace 30 ft lower, in the shrub collection and test garden area, still shows an abandoned meander channel at its outer bend. This lower meander scar must be very close to a bedrock floor.

Stop 5. Flat Rock

This section of Fall Creek, slightly eroded into shale and siltstone, is the local and temporary base level for upper Fall Creek valley. No bedrock is now exposed in the stream bed until north of Maclean, a distance of at least 10 miles. In reexcavating its interglacial valley Fall Creek here encountered a bedrock spur and became superposed across it, thus greatly deterring further downcutting and permitting the broad lateral swings of the meander belt upstream. This popular local swimming hole was almost obliterated by a nasty overnight rainstorm in late October 1981. Some of the large slabs of siltstone in the stream bed were observed to be flipping edge-over-edge during the flood, which reached almost to road level.

Several hundred meters downstream, Fall Creek drops over a low waterfall and loses its bedrock floor, which appears again upstream of the Forest Home Drive bridge at the Wildflower Garden. An ancient buried channel of Fall Creek apparently crosses beneath the modern channel in the interval of no rock exposures.

Stop 6. Mundy Wildflower Garden, Cornell Plantations.

The Mundy Wildflower Garden is established on the floor of Fall Creek valley in another segment of a buried interglacial valley. A short distance downstream from the parking lot, Fall Creek again drops over a low waterfall into a segment a few hundred meters long in which no bedrock is exposed. A low, swampy floodplain in a nearly abandoned meander scar creates a natural habitat for many spring wildflowers. An oxbow swamp is in a well defined channel at the base of the cut bank of the meander scar. The Mundy Wildflower Garden is flooded by extreme high water, but the flora seems to thrive on the aperiodic disturbances.

Stop 7. The "Forest Home runaround" and Beebe Lake.

A gravel-floored abandoned meander followed by Plantations Road sweeps cleanly around a meander core (or Umlaufberg) on which the Cornell Plantations headquarters building and the rhododendron garden are located. Where Plantations Road and the "runaround" intersect Forest Home Drive along Beebe Lake, no bedrock is exposed. Beebe Lake, like several valley segments immediately upstream, is in a segment of an interglacial valley with a floor below the modern channel of Fall Creek. At the stone bridge over the upstream end of Beebe Lake, the rock walls of the buried valley are exposed, as they also are at the dam. In between is a buried valley of uncertain depth and trend.

Stop 8. Mount Pleasant

The accordant summits of the Appalachians are a subject long familiar to geomorphologists. In the gentle foreland fold belt of the Ithaca region, structure is not an important factor in summit height. Mount Pleasant is a cuestaform ridge held up by thicker-bedded sandstone and siltstone units in the Upper Devonian section. The plateau to the north, including the subsequent valley of Fall Creek at the base of the north-facing escarpment face, ranges in altitude from 1000-1400 feet and is underlain by more erodible shale formations. As the regional topographic slope is to the north and the strata dip south, there is considerable truncation of strata by the upland surfaces (Fridley, 1929, p. 116).

Fridley correlated the upland around Ithaca with the Schooley Peneplain of Pennsylvania, generally found in central New York at an altitude of 1600-1700 feet. He supposed that the somewhat higher hills trending northeast-southwest just south of Ithaca were not controlled by more resistant bedrock, but were along the preglacial divide between the St. Lawrence and Susquehanna drainages. Of course, the concept would imply that headward erosion and capture by a north-flowing stream has progressed as far south as the Ithaca region in preglacial time.

Mount Pleasant and its westward extension known locally as Turkey Hill have been segmented into north-south elongate elliptical ridges by glacial erosion and perhaps meltwater overflow. An especially clear example is a small gully that originates on the north face of Mount Pleasant one mile west of the observatory and drains south into Ellis Hollow across the trend of the ridge crest. No catchment area is available for this gully

to erode such a channel, but if it collected meltwater from an ice margin that briefly rested against the north face of Mount Pleasant in late-glacial time, an adequate flow would have been available to cut such a channel in the few centuries that were available for such events.

Stop 9. Belle School Road, Caroline.

The road, here just below 980 feet, forms the divide between the St. Lawrence and Susquehanna drainage. This is the floor of the meltwater channel that was the outlet of glacial Lake Ithaca. To the south, the gradient averages about 10 ft/mi in the 18 miles to the Susquehanna River at Owego. To the north, however, Sixmile Creek drops 600 feet in about 8 miles to Cayuga lake, a gradient of 75 ft/mi. The regional asymmetry of stream gradients is not obvious at Belle School Road, however; swamps drain in both directions from the flat valley floor.

This is probably the valley that was described by R.S. Tarr at the 1905 meeting of the Geological Society of America, when in complimenting Tarr for his perceptive analysis, W.M. Davis proposed that such valleys should be called "through valleys" (Tarr, 1905). Meltwater overflow from proglacial lakes on the north sides of the divides undoubtedly helped erode these valleys, since lakes formed both during the advance and retreat of each ice margin that crossed the region.

Most of the highest hills of the region lie within a few miles north or south of the present divide. In the valleys, the acute junction angles of tributaries point north on the north side of the divide and south on the south side. Even though the summits have been rounded and somewhat lowered by ice erosion, the valleys have been widened and lowered, and the former valley side spurs have been trimmed back, both summit heights and tributary junction angles suggest that the present divide is not far from its preglacial position. It is a little surprising that the nearly 10-fold gradient advantage of the north-flowing drainage, that must have been established by mid-Pleistocene time, has not yet begun to shift the divide toward the south from this point. The two tributaries that drain down the west valley wall from Durfee Hill cross the western kame terrace only a few hundred meters apart, and their alluvial plains are separated only by Belle School Road. In any local storm either one could divert and be captured by the other.

Stop 10. (optional). King Road Overlook (1300 feet)

The view north over the Cayuga trough and Ithaca provides an excellent review of the regional geomorphology.

Stop 11. Upper Robert H. Treman State park (Enfield Glen)

The final stop of the trip is in Upper Treman Park, where Enfield Creek has reexcavated an interglacial valley to a point a short distance downhill from the parking lot. There, the postglacial stream became superposed on a bedrock spur and has entrenched along vertical joints to a depth of at least 120 feet above Lucifer Falls and more than 300 feet below the falls, before it submerges in its interglacial valley. The westward retreat of Lucifer Falls has left at least one former tributary

hanging on the high south wall of the gorge. The interglacial valley goes through glacial drift on the hillside just north of the gorge entrance, and is marked by a zone of active seepage, soil creep, and tree throw on both its upstream and downstream face.

The large size of the valley below Lucifer Falls suggests that this valley, like the valleys of Taughannock Creek, Fall Creek, Six Mile Creek, and perhaps others, may have been repeatedly excavated and back-filled with glacial drift during the numerous ice ages of the Quaternary Period.

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ROAD LOG

CUMULATIVE MILEAGE	MILES FROM LAST POINT	ROUTE DESCRIPTION
0.0	0.0	Start, Snee Hall, Cornell University. Exit right on College Ave.
0.2	0.2	Turn right on Campus Road, continue straight ahead past traffic booth on Garden Ave.
0.6	0.4	Turn right on Tower Road. Ascending a series of hanging deltas through the campus.
1.1	0.7	Turn left on Judd Falls Road.
1.5	0.4	Turn right on Forest Home Drive, over bridge and bear right.
1.7	0.2	Turn left up Warren Road.
2.8	1.1	Turn right on Hanshaw Road.
3.2	0.4	Turn right on unpaved Bluegrass Lane.
3.4	0.2	STOP 1 at vegetable gardens.
3.6	0.2	Return to Warren Road, turn right. Note subdued ground moraine topography.
4.1	0.5	Turn right on Freese Road.
4.3	0.2	Turn left into driveway of Dyce Laboratory (Honey Bee Studies); continue straight on Farm Road.
4.5	0.2	STOP 2 in woods above Varna high bank.
4.6	0.1	Return to Freese Road, turn left.

- 5.5 0.9 Cross Fall Creek. Bridge and bank repairs postdate 1972 and 1981 floods.
- 5.6 0.1 Turn left on State Route 366 in Varna.
- 5.7 0.1 Turn left at fire hydrant down farm road in Cornell experimental fields. Follow best gravel road to north across the fields.
- 5.9 0.2 Down scarp of upper terrace.
- 6.2 0.3 Turn right on gravel road.
- 6.3 0.1 Turn left along forest edge.
- 6.4 0.1 STOP 3 at base of the "Varna Moraine" (actually an alluvial terrace scarp). Walk north through the woods to the bank of Fall Creek.
- 7.1 0.7 Return to State Route 366, turn right.
- 7.7 0.6 Turn right on Forest Home Drive. Flood of 1981 exposed bedrock in channel, here.
- 8.5 0.8 Turn left into Cornell Plantations, follow one-way signs to right.
- 8.7 0.2 Turn right, twice, up Plantations Road.
- 8.9 0.2 Turn left on Arboretum Drive.
- 9.1 0.2 STOP 4 at Cornell water tanks.
- 9.3 0.2 View east of Mt. Pleasant cuesta.
- 9.7 0.4 View southwest from Newman overlook: windgap of ancestral Fall Creek?
- 10.3 0.6 STOP 5. "Flat Rock"; Fall Creek at Plantations entrance.
- 10.8 0.5 Continue downstream on Forest Home Drive to stop sign at Caldwell Road. Go straight ahead into parking lot of Mundy Wildflower Garden. STOP 6 in Wildflower Garden.
- 10.9 0.1 Exit right from parking lot onto Caldwell Road. Turn right on Plantations Road through the underpass around an abandoned meander to Beebe Lake.
- 11.5 0.6 Turn right on Forest Home Drive.

11.6	0.1	STOP 7: intersection of McIntyre Place and Forest Home Drive. Walk to stone bridge at head of Beebe Lake.
11.7	0.1	Up McIntyre Place to Judd Falls Road. Turn right.
12.0	0.3	Turn right on Tower Road.
12.7	0.7	Turn left on Garden Ave, continue downhill past traffic booth to College Ave.
13.1	0.4	Turn left on College Avenue.
13.3	0.2	Turn left into Snee Hall parking lot. LUNCH STOP. (End of Sunday excursion.)
0.0	0.0	Start, Snee Hall, Cornell University. Exit right on College Avenue.
0.2	0.2	Turn right on Campus Road.
0.4	0.2	Turn right at traffic booth on Campus Road.
0.7	0.3	Traffic light. Continue straight ahead (east) on State Route 366.
2.0	1.3	View ahead of Mt. Pleasant, Ellis Hollow to the southeast.
4.6	2.6	Turn right up Baker Hill Road. Good views to north during ascent.
5.2	0.6	Small bench on hillside. Drainage on west side of road is to the south, through the ridge crest.
5.7	0.5	Turn left on Mt. Pleasant Road (unpaved).
6.6	0.9	STOP 8. Observatory, radio towers.
7.5	0.9	Return west on Mt. Pleasant Road onto paving straight ahead past Baker Hill Road.
7.7	0.2	Small gully noted at mileage 5.2 passes under road. Probable meltwater valley.
8.5	0.8	Views north and west from near Deer Haven Drive.
9.0	0.5	Turn left on Turkey Hill Road. Views southwest of Ithaca College. Upper row of college buildings are on an ice-marginal overflow channel at 1040 ft. above sea level.
9.7	0.7	Cross Cascadilla Creek. Old records described a well 100 feet deep that did not reach bedrock in the valley floor near here (Tarr, 1909a, p. 20).

- 10.6 0.9 Turn left on Ellis Hollow Road. Across the road is the main quarry of the Finger Lakes Stone Co., from which many Cornell University buildings have been faced with "Llenroc".
- 12.6 2.0 Bear right onto Thomas Road.
- 13.6 1.0 Good view to left of the 1060 ft. meltwater channel that controlled Freeville-Dryden Lake.
- 14.4 0.8 Beaver pond in south-draining floor of meltwater channel.
- 15.0 0.6 Cross State Route 79 at Caroline. Continue straight ahead on Lounsberry Road (County Route 113). Road follows radial slope of the large Brooktondale delta/fan.
- 16.1 1.1 Cross State Route 330. Continue ahead up White Church Road onto delta/fan top.
- 16.6 0.5 Turn left and continue south on White Church Road. Cross major distributary channel in church camp. Continue south on kame terrace with colluvial cover from steep hillside.
- 18.6 0.3 STOP 9 on St. Lawrence - Susquehanna divide.
- 19.1 0.5 Continue west on Belle School Road to Coddington Road. Turn right on Coddington Road. Drive north along kame terrace.
- 20.4 1.3 Turn left on Miller Road, up Steventown Hill.
- 22.3 0.5 Turn right on Nelson Road.
- 24.1 1.8 Turn right on Troy Road.
- 25.3 1.2 Turn left on King Road.
- 25.6 0.6 STOP 10. Cayuga Valley overlook.
- 26.6 1.0 Turn right on State Route 96B (Danby Road).
- 27.1 0.5 View of Cayuga trough and Cornell Campus. Note convex valley walls.
- 28.7 1.6 Turn left on Clinton Street (96B).
- 29.5 0.8 Turn left on Meadow Street (Rt. 34-13).
- 30.1 0.6 Stop light. Continue south on Elmira Road (Rt. 34-13).

- 31.2 1.1 Entrance to Buttermilk State Park. View of Buttermilk Falls.
- 31.7 0.5 Railroad overpass and junction with Floral Avenue. Note hanging deltas of Coy Glen to right rear.
- 32.7 1.0 Turn right on State Route 327 to Treman State Park.
- 33.1 0.4 Continue to right on State Route 327 past lower park entrance.
- 33.6 0.5 Abandoned gravel pit in hanging delta. Excellent hanging-delta morphology uphill for next half mile.
- 35.7 2.1 Turn left into upper entrance of Robert H. Treman State Park.
- 36.6 0.9 STOP 11. Upper Enfield Glen. Walk downstream into gorge to base of Lucifer Falls. Cross footbridge, return to parking lot via South Rim trail.

