



**Department of Geology and  
Environmental Science  
Hartwick College  
and  
Department of Earth Sciences  
State University of New York College  
at Oneonta  
Oneonta, New York**

**New York State  
GEOLOGICAL ASSOCIATION  
*Field Trip Guidebook***

**75<sup>th</sup> ANNUAL MEETING  
SEPTEMBER 12-14, 2003**

# **NEW YORK STATE GEOLOGICAL ASSOCIATION**

**75<sup>th</sup> Annual Meeting**

**September 12-14, 2003**

**Eric Lee Johnson, Editor**

## **FIELD TRIP GUIDEBOOK**

**HOSTED BY:**

**Geology and Environmental Science Department  
Hartwick College  
Oneonta, New York  
13820  
and**

**Department of Earth Sciences  
State University of New York  
College at Oneonta  
Oneonta, New York  
13820**

**PUBLISHED BY THE NEW YORK STATE GEOLOGICAL ASSOCIATION.**

Additional copies may be obtained from the Association by contacting  
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## DEDICATION

This guidebook is dedicated to all of those who helped make this meeting possible. Special thanks are given to Dr. Jim Ebert at SUNY Oneonta for his tireless efforts in making the arrangements for the meeting, and to Dr. Harrison Schmitt for coming and helping us celebrate NYSGA's 75<sup>th</sup> anniversary. Finally, this meeting would not have been possible without the dedication and hard work of all of the field trip leaders who have spent countless hours both in the field and at the keyboard constructing the fabulous array of trips presented on the following pages.

*The tops of mountains are among the unfinished parts of the globe, whither it is a slight insult to the gods to climb and pry into their secrets, and try their effect on our humanity. Only daring and insolent ones, perchance, go there.*

*Henry David Thoreau*

*Cover photo: Overlooking the Hudson Valley at the Mountain House site: North Lake, Catskill Park.*

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## ***Preface and Acknowledgements***

It is a great pleasure to welcome you to the 75<sup>th</sup> annual meeting of the NYSGA. Hartwick College and the State University of New York College at Oneonta are this year's host institutions and we hope that you take some time to visit both schools a tour our respective Geology Departments.

This year's field trips cover a broad array of topics and regions and will take you from the depths of caves to the tops of Catskill and Adirondack peaks. The range of topics is certainly varied including the fields of metamorphic and igneous petrology, structural geology, sedimentology, hydrology and paleontology.

In honor of the 75<sup>th</sup> anniversary of NYSGA, we are fortunate to have Dr. Harrison Schmitt as our keynote speaker. Having walked on the moon, Dr. Schmitt brings a very unique perspective on geology and we look forward to hearing his address. It is fitting that a lunar geologist present at NYSGA, since both the moon and the State of New York have one thing in common; terrific exposures of anorthosite.

NYSGA meetings are unique and refreshing in that they focus on learning about geology by going out into the field and seeing it first hand. Because of this focus, NYSGA depends on the many individuals who have dedicated themselves to planning and running field excursions. Their tireless efforts are what make NYSGA meetings so enjoyable and valuable. I thank all of the authors and trip/workshop leaders, for their hard work.

I would also like to thank Dr. Jim Ebert for arranging the meeting (as well as being a trip leader and author of multiple field guides) and for his help in making this field guide a reality. Jim served as the field guide editor for the 1991 NYSGA meeting held in Oneonta, and his expertise, advice and calming influence helped me immensely. Thanks Jim.

In addition, I thank the staff at SUCO and Hartwick College for all of their help and technical support, and finally, I thank my students who make what I do so enjoyable.

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Field Guide Editor  
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**Dr. James Ebert**  
President 2003 Meeting  
New York State Geological Association

*Trip A-1***LATE MIDDLE DEVONIAN BIOTIC AND SEDIMENTOLOGIC EVENTS IN  
EAST CENTRAL NEW YORK – TULLY FORMATION CLASTIC  
CORRELATIVE SUCCESSION IN THE SHERBURNE – ONEONTA AREA.**

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Terrigenous clastic deposits equivalent to the upper most Moscow Formation and Tully Formation record a number of key sedimentological-paleontological events that occurred during the latest part of the Middle Devonian Givetian Stage. These deposits, last mapped in detail by G. Arthur Cooper and H.S. Williams in the 1930s, have been the subject of extensive stratigraphic mapping work by the present authors over the past five years. Goals of this mapping project are: 1) to identify divisions and boundaries correlative to units identified in the Tully Limestone further west through the classic work of Phillip Heckel in the sixties and seventies; 2) to identify key facies that can be interpreted in the context of onshore-offshore paleoenvironmental models; 3) to reconstruct the faunal succession in this thick interval of strata so as to elucidate in greater detail two key levels of faunal change associated with the global Taghanic Bioevent; and 4) to use stratigraphy and sedimentology to identify structural features that were active or present during deposition of these units.

Preliminary results of our work show that most of the lower and medial parts of the Tully Formation clastic correlative succession (here in referred to as "TFCCS") between Sherburne, NY and New Lisbon, NY consist of relatively unfossiliferous strata that record the development of a structural basin (New Berlin Trough) probably centered in what is now the Unadilla Valley. Condensed units overlying regional discontinuities developed at several stratigraphic levels both in the basin and on its western flank; sediment-starvation associated with condensation led to the formation of "oolitic" chamosite under conditions not fully understood. Basin center deposits recorded dysoxic conditions and development of low diversity faunas dominated by auloporid corals and small brachiopods; eastward passage of sparsely fossiliferous deposits near New Berlin to richly fossiliferous equivalent units near Laurens reflects a shoreward basin-to-shelf transition. This contradicts the standard view that the lower and medial Tully record primarily shallow and even lagoonal conditions. Our model does vigorously support (and significantly enhances) Heckel's (1973) view of a "Tully clastic trap" controlled by structure in east-central New York.

Herein, we show that typical oxic Hamilton facies ends with onset of dysoxic, highstand conditions associated with deposition of the New Lisbon Member. We exclude the upper, Tully-equivalent part of Cooper and Williams' (1935) "New Lisbon Member", which is separated from our restricted New Lisbon Member by a basal Tully erosion surface. Currently, we believe that the restricted New Lisbon succession is equivalent to the Gage Gully submember of the Windom Member, although this is still uncertain. The TFCCS contains six internal mappable discontinuities; these are, in upward-succession: a, a base-Tully (DeRuyter Bed-equivalent?) contact; b, a sub-"Laurens" contact, possibly

equating to the base of the Fabius Bed to the west; c, a sub-Smyrna Bed disconformity traceable westward into central New York; d, a top-Smyrna flooding surface contact; e, an erosion surface flooring the West Brook Bed; and f, a diastem (base-Moravia Bed? discontinuity) above the West Brook Bed. The upper Tully, base-West Brook contact marks the greatest lowstand event with development of a prolific Hamilton faunal association characterized by diverse brachiopods, and corals, above the contact. East of Otselic, the Tully Formation contact with the overlying Genesee Black Shale is conformable.

Initial Tully Fauna incursion occurs at the base-Tully (DeRuyter Bed-equivalent?) contact at New Lisbon and Laurens. The Tully Fauna persists above the Smyrna Bed into the middle part of a sparsely fossiliferous post-Smyrna Bed highstand-succession probably correlative with the Taughannock Falls Bed within the Tully Limestone. The uppermost part of the highstand succession is characterized by a return of the pre-Tully Hamilton fauna. The West Brook Bed records a major, anomalous acme (“last hurrah”) of the Hamilton Fauna prior to onset of dysoxic and, finally, anoxic conditions associated with uppermost Tully and Genesee deposition. Preliminary work on eastern sections of the TFCCS suggests that nearshore, Tully equivalent “Chemung-type” facies may yield several taxa of the Ithaca fauna absent from outer shelf, Tully platform carbonates.

## INTRODUCTION

The Tully Formation in New York State is an anomalous carbonate unit within the thick terrigenous Devonian foreland basin succession. It was known for the occurrence of *Hypothyridina* “*cuboides*” (now *Tullythyridina venustula*; Sartenaer, 2003) and other distinctive taxa as far back as the mid-nineteenth century. Detailed study of the Tully Formation was initially undertaken by Cooper and Williams (1935) in New York State and by Willard (1937) in Pennsylvania. Cooper and Williams (1935) recognized that clean carbonates comprising the Tully Formation south of Syracuse and in the Finger Lakes region pass eastward into thin, partly oolitic deposits in the Sherburne area in Madison County before grading into a thick correlative clastic succession east of there. We have worked from the initial stratigraphic framework established by these workers (see Figures 1 - 3) in building the correlation schemes presented herein (see Figs. 4 - 11).

The most detailed and important study of the Tully Formation is that of Phillip Heckel in the 1960s and early 1970s (Heckel, 1966, 1969, 1973). Rather than defining large facies tracts within units as was the fashion of the day, Heckel came to recognize specific beds in the Tully Limestone as mappable events.

---

Figure 1. Named subdivisions of the Tully Limestone in New York. Tully Limestone comprises two members which include all units termed “bed” or “mound.” Unnamed sandstone, Unadilla Formation, and West Brook bed east of vertical dashed line are parts of eastern detrital equivalent of Upper Member. Wavy lines are major disconformities; other lines are bed, member, and formation boundaries; zigzag lines represent known facies relations. Datum levels are Bellona Bed and West Brook Bed. BMB – back-mound beds; MB – middle beds; UB = upper beds. From Heckel (1973).

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Figure 2. Correlation of key Tully divisions relative to major structural features based on work of Heckel (1973). From Heckel (1973).

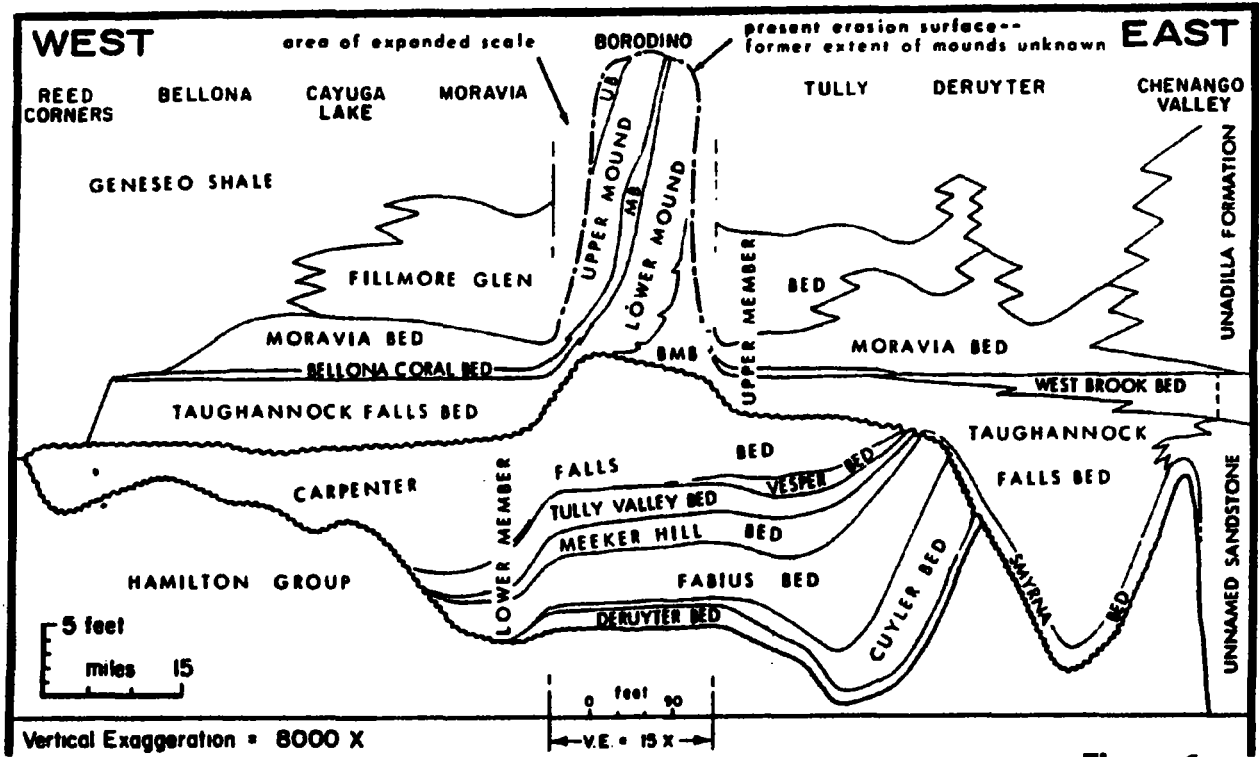


Figure 1

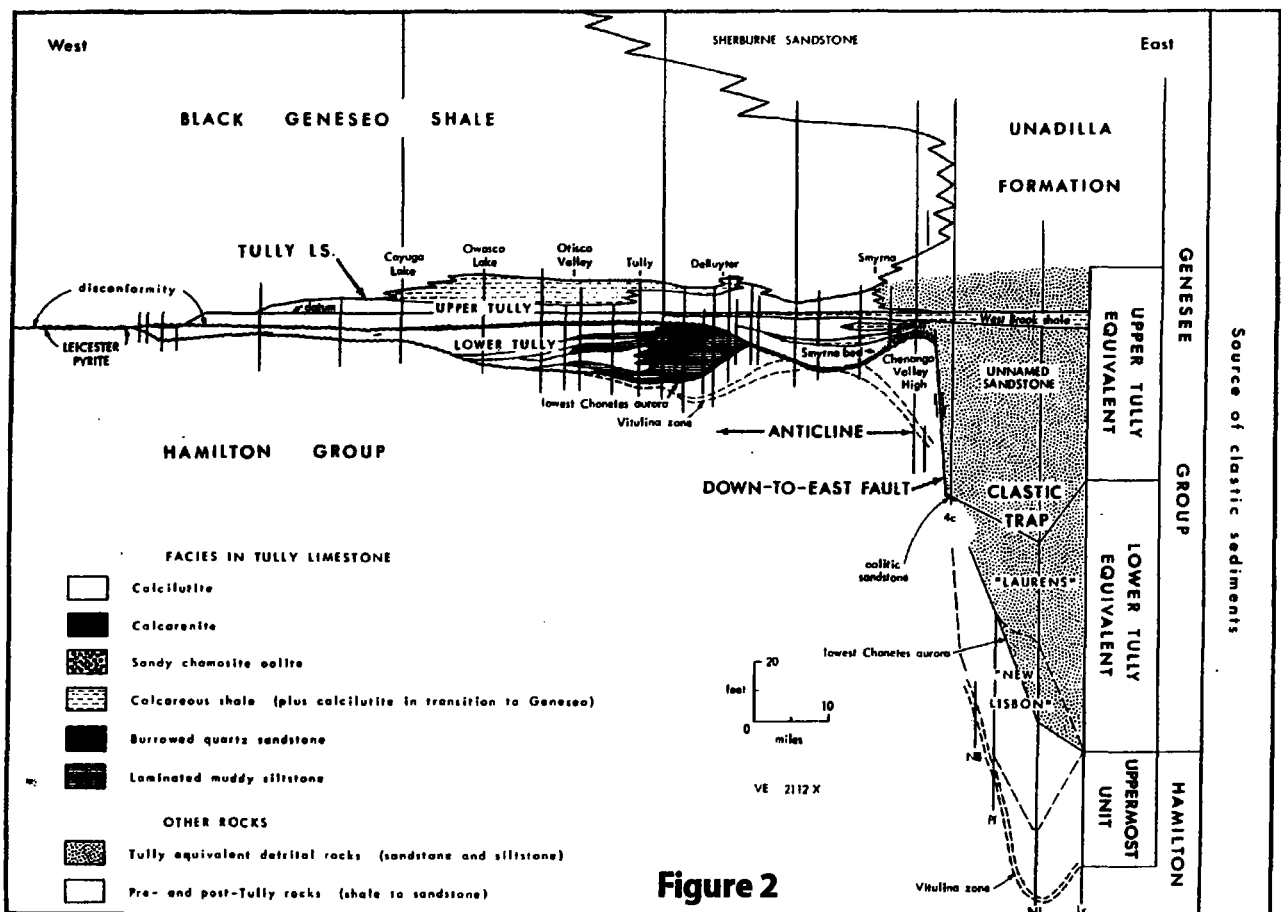


Figure 2

| GENESEE GP.<br>GENESEO SH.   |              | Stratigraphic subdivisions of Heckel, 1973 |                   | C & W "MEMBER" | FAUNAL ZONE  | C & W "MEMBER"  | Stratigraphic subdivisions of Heckel, 1973 |  | (Rickard, 1964)             |
|------------------------------|--------------|--|-------------------|----------------|--|-----------------|--|--|-----------------------------|
|                              |              | (Tully Limestone)                          | ↓                 |                |  |                 | (Tully equivalent)                         | ↓  |                             |
| TULLY LIMESTONE              | Upper Member | 5 beds                                     | West Brook        | West Brook     | <u>Metriophyllum tullium</u><br>(formerly <u>Lopholasma</u> )<br><br><u>Elytha fimbriata</u> | West Brook      | 4  | Upper Tully equivalent<br>(4 subdivisions) | EASTERN DETRITAL EQUIVALENT |
|                              |              | 1 bed                                      |                   |                |  |                 | 3  |  |                             |
|                              | Lower Member | 5 beds                                     | Apulia            | Apulia         | <u>Hypothyridina venustula</u>   | Laurens         | 1  | Lower Tully equivalent<br>(2 subdivisions) |                             |
|                              |              | 2 beds                                     |                   |                |  |                 | 2  |  |                             |
| HAMILTON GROUP<br>MOSCOW FM. | MOSCOW FM.   |  |                   | Tinker Falls   | <u>Chonetes aurora</u><br>( <u>C. aurora</u> only in top)                                    | New Lisbon      | 1  |  | GENESEE GROUP               |
|                              |              |  |                   |                | <u>Leiorhynchus mesacostale</u>  |                 | 2  |  |                             |
|                              |              | (W. CENTRAL NEW YORK)                      | (CHENANGO VALLEY) |                |  | (OTSEGO COUNTY) |  |  | HAMILTON GROUP              |

Figure 3. Comparison of Heckel's (1973) internal classification scheme for the Tully Limestone to that of Cooper and Williams (1935). Modified from Heckel (1973).

Numerous such beds were used to establish a detailed correlation network for the Tully (Figures 1 - 3). Using this approach, Phillip Heckel was able to reconstruct a pattern of eastward erosional overstep of lower Tully divisions along a dynamic contact within the Tully (Figures 1, 2). He, further, showed that medial Tully carbonate-rich strata (Taughannock Falls Bed) passed eastward into a thin sandstone and chamositic oolitic interval in the Chenango Valley before expanding into a thick silty-sandy facies succession in the Columbus-New Berlin area (Figure 2). In particular, Heckel (1973) recognized that during Tully time, a structural "clastic trap" represented by thick clastic sediment accumulation had developed east of a Sherburne area structural "high" (Figure 2). Heckel (1973) recognized a rapid and conspicuous eastward facies change within the lower Tully succession between Tully and DeRuyter from limestone to siltstone and fine sandstone towards the area of lower Tully overstep noted above (Figure 2). From this pattern, he concluded that uplift and subaerial(?) erosion of the lower Tully succession had led to redeposition of these clastics in areas within the Tully proximal to the uplift. Moreover, he argued that the clastic trap to the east, and the lower Tully uplift event locally, had blocked the influx of clastics from the east allowing for deposition of clean limestone across central New York. Heckel (1973) further characterized the upper Tully "*Elytha fimbriata*" interval of Cooper and Williams (1935) by defining the West Brook Shale in east-central New York and the Bellona Coral Bed from Tully westward to Gorham near Canandaigua Lake. Detailed faunal lists from each of the numerous Tully Formation divisions greatly improved the resolution of the known Tully faunal succession. Heckel (1973) argued that the Tully generally recorded shallow, even lagoonal conditions. Most important to this argument was the discovery of mud cracks on the top of the Carpenters Falls Bed at many localities.

This contact, marking his mid-Tully erosional unconformity (Figures 2, 3) was found encrusted with stromatolites at Bellona, NY, further supporting his shallow water interpretation of the Tully.

Johnson and Friedman (1969) examined siliciclastic deposits correlative to the Tully Formation from the Susquehanna Valley eastward to the red bed margin east of the Schoharie Valley. This sedimentological study illustrated a variety of nearshore facies proximal to the inferred paleoshore of the "Catskill Delta" within the Gilboa Formation. Similarly, Bartholomew, et al. (2003) are currently examining divisions within the Gilboa Formation and adjacent units in the Susquehanna Valley-Windham region. This work is still in progress.

Baird and Brett (1986) examined the relationship of the Leicester ("Tully Pyrite") Member to the Tully Limestone, the underlying Windom Shale Member and the overlying Genesee black shale. In contrast to various earlier workers who correlated the Leicester directly to the Tully, they found the Leicester to be a detrital pyrite-bone bed lag deposit that was essentially post-Tully in age and chronostratigraphically correlative with the Genesee Member (see also, Huddle, 1981). Macrofossils within the Leicester were found to be correlative to underlying Windom and (locally) Tully beds and the conodonts are post-Tully in age. We envision Leicester pyritic clasts and bone debris to be the product of combined processes of submarine erosion and submarine carbonate dissolution acting in a predominantly dysoxic to near-anoxic basin setting (Baird and Brett, 1986). In our reconstruction, Leicester pyrite deposition was associated with a major deepening event (Taghanic Onlap *sensu* Johnson, 1970) responsible for widespread

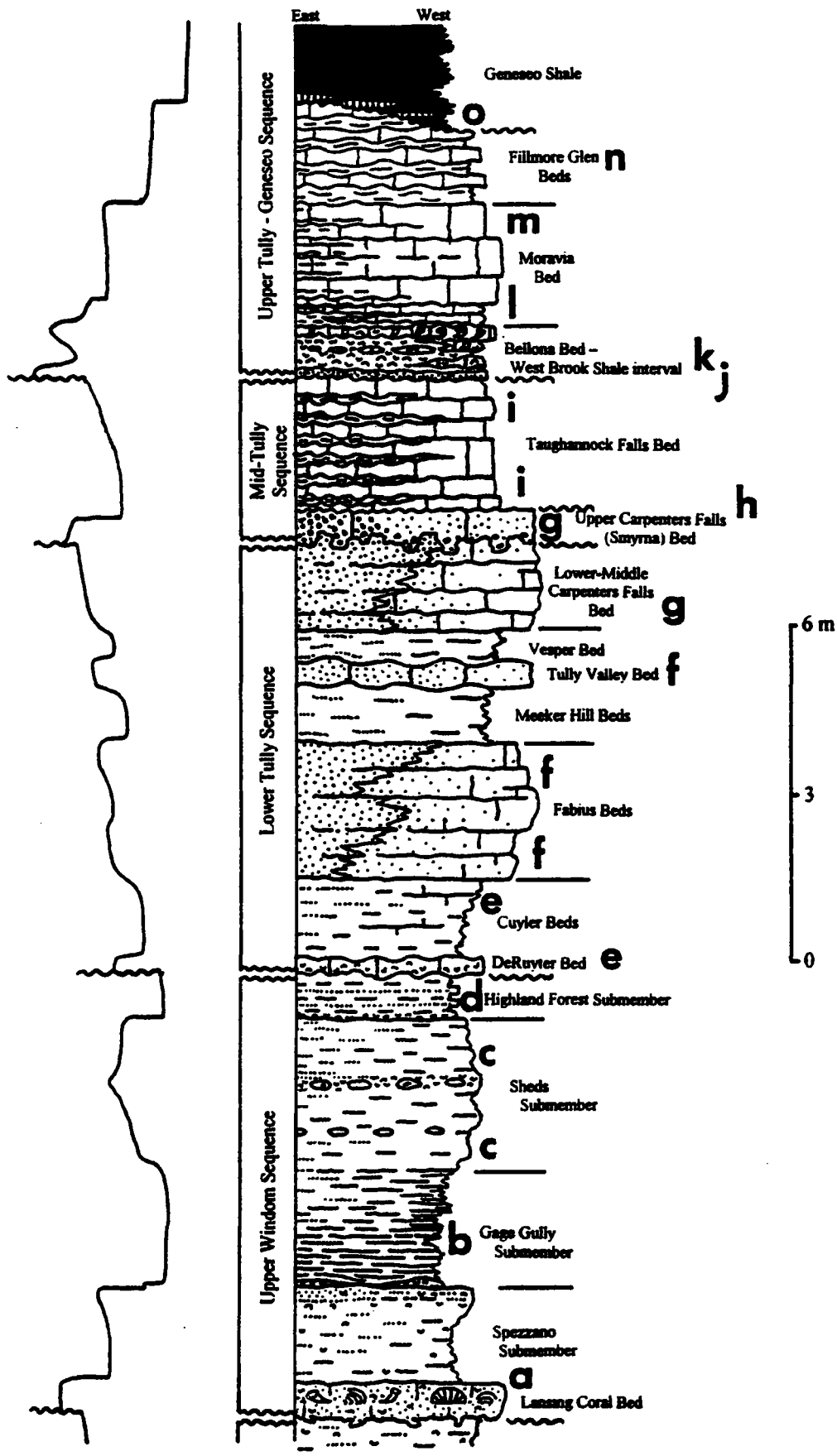


deposition of the Genesee black shale in the foreland basin. This transgression was apparently the local expression of a global eustatic highstand event (see Johnson, et al., 1985). However, the overspread of Genesee black mud lithofacies is also believed to be due to a concurrent flexural response of the craton to an inferred thrust-loading event (“Third Tectophase”) within the foreland basin (see Eddensohn, 1998). In this model, Tully deposition represented a time of stable tectonic equilibrium prior to thrust loading and consequent shelf collapse.

The “Taghanic Bioevent”, an interval of global faunal incursions and paleocommunity reorganization has been the subject of increased interest in recent years (Aboussalam and Becker, 2001; Aboussalam, et al., 2001). It followed a much longer (6 –7 M.Y.) interval of relative ecological-evolutionary stability that spanned the lower and middle Givetian (Brett and Baird, 1995). This longer period of community-level stability (coordinated stasis) is believed to be responsible for the temporal persistence of the observed long-standing Hamilton Fauna, characteristic of strata below the Tully Formation and its equivalents (Brett, et al., 1994; Brett and Baird, 1995). Onset of Tully Formation (DeRuyter Bed) deposition coincides with the demise of many typical Hamilton Fauna elements and the incursion of several Old World Realm taxa including: *Rhysochonetes*, *Tullypothyridina*, and *Schizophoria* (Cooper and Williams, 1935; Heckel, 1973; Baird and Brett, 2003). This “first wave” incursion establishes the Tully Fauna typical of the lower Tully succession up through the Carpenters Falls Bed and Smyrna Bed interval (Heckel, 1973). Beginning within the higher Taughannock Fall Bed interval and continuing through the remainder of the Tully is a “second wave” return of the Hamilton Fauna (Cooper and Williams, 1935; Heckel, 1973; Baird and Brett, 2003). The disappearance of many Hamilton taxa followed by the dramatic return of these organisms

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Figure 4. Generalized stratigraphic column for the upper Windom through basal Genesee Formation-succession in central New York region (Tully, New York-DeRuyter, New York area). Left column shows inferred sea level variations (shallower to left). Stratigraphic column uses standard symbols except that coarse stipple denotes chamositic ooids. Lettered features and events include: a, diverse Hamilton large coral biofacies; b, dysoxic small brachiopod biofacies; c, moderate diversity, mid-shelf Hamilton fauna; d, dysoxic, low diversity, outer-shelf Hamilton fauna; e, lower Tully fauna incursion event expressed in minimally oxic, mid-outer shelf, small-brachiopod biofacies; f, moderate diversity, mid-shelf expression of lower Tully fauna; g, peak development of lower Tully fauna; h, peak development of lower Tully fauna; h, “widespread middle Tully unconformity” of Heckel (1973) = maximum flooding surface boundary of present report; i, reappearance of many Hamilton fauna taxa, demise of lower Tully fauna; j, regional disconformity flooring West Brook Shale; k, high diversity middle to inner shelf Hamilton fauna biota of Bellona Bed – West Brook Shale interval; l, approximate level of first appearance of key goniatite *Pharciceras amplexum*; m, low diversity outer shelf biota of Moravia Bed (small rugose corals, auloporids, and phacopid trilobites predominate); n, dysoxic low diversity biota (auloporids and rare small rugosans); o, detrital pyrite lenses (Leicester Pyrite) recording submarine erosion and corrosion of uppermost Tully deposits. From Baird and Brett (2003).



indicates that the "Taghanic Bioevent" is complex and only partly understood. The present authors believe that the thick succession of the TFCCS will reveal much more concerning the dynamics of these faunal changes.

### **THE TULLY CHALLENGE**

Because of unusual lithologies and major patterns of faunal change, the Tully Formation and the TFCCS to its east present a major opportunity to examine a variety of coincident sedimentological, paleontological, eustatic, and structural changes during the latest Middle Devonian. Research during the past five years, involves study of Tully deposits in both New York State and Pennsylvania. In this report the emphasis will be the relationship of the TFCCS to the Tully limestone further west, with some mention of the Pennsylvania Tully sections included. We discuss a variety of units and boundaries in the TFCCS in the context of the reports of Cooper and Williams (1935) and Heckel (1973). New correlations will be presented and correlation issues discussed. The relative depth significance of ecological-evolutionary change will be discussed. Finally, some discussion of chamosite distribution and origin will be included.

### **STRATIGRAPHIC RELATIONSHIPS: GENERAL OVERVIEW**

Strata discussed include the topmost part of the Cooperstown Member, the New Lisbon Member, and the Tully Formation Clastic Correlative Succession (TFCCS). The upper part of the Cooperstown Member (=Windom Member in west-central New York) and the TFCCS correlates westward into the Tully Formation at the Sherburne Meridian and further west (Figures 1 -3, 5 - 8, 11, 12). One unit, the New Lisbon Member, described by Cooper and Williams, 1935, is redefined (restricted) to include only pre-Tully strata and to represent the highest division of the Moscow Formation (Figures 8, 11). Stratigraphic transects included are: 1) condensed easternmost Tully Formation in the Smyrna-Sherburne area in the Chenango Valley (Figure 5); 2) condensed easternmost Tully-into-expanded TFCCS succession transition between the Chenango and Unadilla Valleys (Figures 6, 7); and 3) lower-medial part of TFCCS interval from the Unadilla Valley eastward nearly to the Oneonta meridian (Figures 8, 11, 12).

---

Figure 5. Tully Formation stratigraphic transect across the western part of the "Sherburne High" structural axis showing the thinning of the Tully interval and development of chamosite at several levels. Datum for both A and B set at base of West Brook interval (unit h). A, transect from Smyrna area to Sherburne area. Asterisk between columns for localities 3 and 4, denotes uncertainty as to whether Smyrna Bed at locality 3 is correlative with basal Tully chamositic unit (designated Smyrna Bed on diagram) at localities 4 - 12, or whether is correlates to the higher "Taughannock Falls Oolite Bed at localities 4 - 12 (see discussion in text); B, Tully Formation sections in Sherburne-Harrisville area. Lettered units in A and B include: a, Cooperstown Member (Spezzano submember-equivalent?) strata; b, base-Smyrna Bed disconformity; c, Smyrna Bed succession; d, "unnamed sandstone" of Heckel, 1973; e, "Taughannock Falls Oolite Bed" of Heckel, 1973; f, basal bed of West Brook Shale interval; g, condensed, variably chamositic facies equivalent to Taughannock Falls Bed succession; h, basal bed of West Brook Shale interval; i, fossiliferous West Brook Shale interval; j, coral/crinoid-rich limestone bed above West Brook Shale; k, silty, small-coral and phosphate pebble-bearing limestone unit probably marking base of Moravia Bed-equivalent strata; l, Moravia Bed-equivalent strata. Numbered localities are listed in APPENDIX; C, map of area showing distribution of localities.

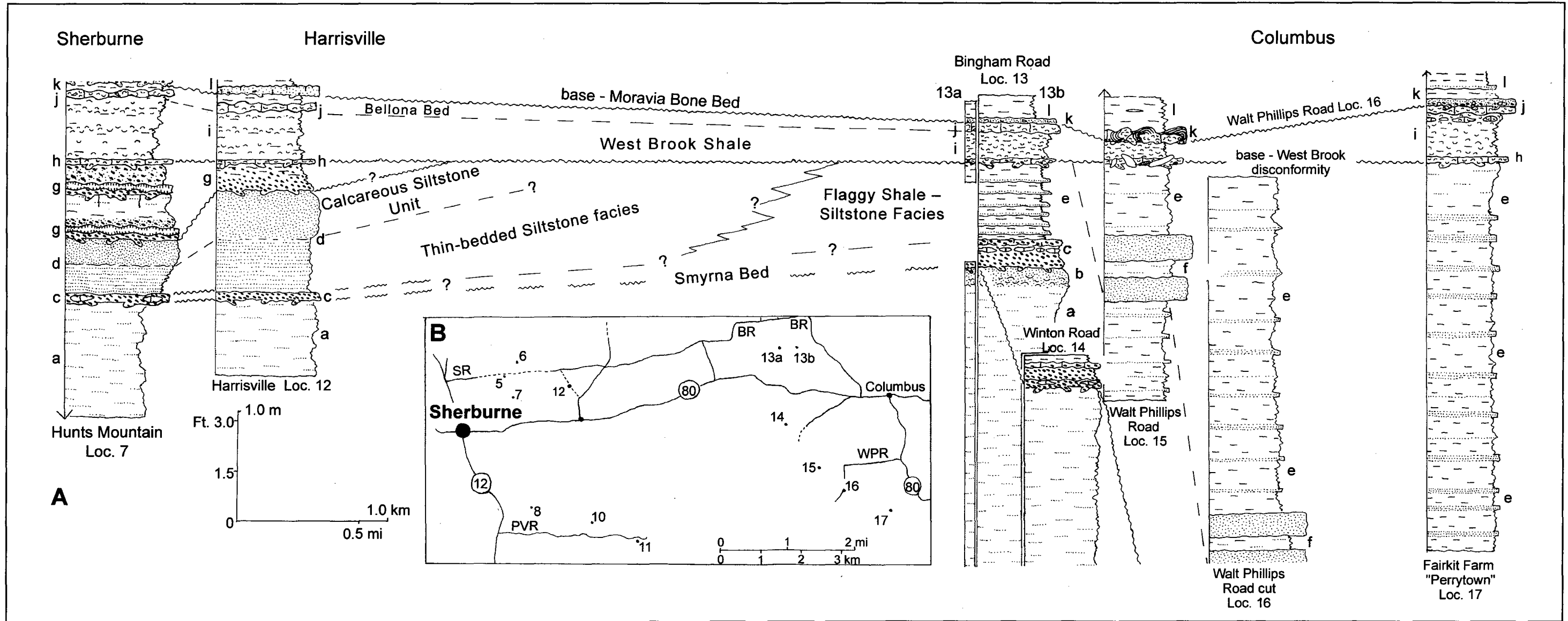


The Tully Formation is essentially a clean, typically micritic limestone forming ledges, quarry walls, and waterfalls from its western terminus near Canandaigua Lake eastward to Tully near the Syracuse meridian. The lower part of the Tully in this region (DeRuyter Bed-topmost Carpenters Falls Bed interval) yields various taxa of the Tully Fauna and is capped by a regional unconformity identified by Heckel (1966, 1973) (Figures 1 – 3). This aforementioned contact, marked by mudcracks at several localities and laminated stromatolites at one locality, was correlated eastward to the Smyrna which is characterized by lag debris and chamosite in the DeRuyter-Sherburne region (Heckel, 1973). Another key marker in the upper Tully Formation from Tully westward is the Bellona Bed which yields abundant rugose and tabulate corals and brachiopods typical of the Hamilton fauna; this unit was roughly correlated to the West Brook Shale Bed in localities east of Tully, though Heckel (1973) speculated that the Bellona Bed actually overlies the West Brook Bed where it first becomes recognizable (Figures 2, 3).

East of the Tully Valley, the lower part of the Tully Formation rapidly becomes silty and marginally sandy before it is erosionally overstepped along a contact at the base of the Smyrna Bed. The lower Tully unit succession (DeRuyter Bed-through-Carpenters Falls Bed) is beveled to the east over an interval of fifteen miles at most (Heckel, 1973; Figures 2, 3). Our work suggests that the beveling takes place over an even shorter distance. East of the lower Tully pinch-out, topmost units of the Windom-Cooperstown transitional succession are overstepped as well (Heckel, 1973).

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Figure 6. Stratigraphy of Tully Formation across eastern part of "Sherburne High" axis and westernmost part of Columbus area structural "sag." A, Stratigraphic reconstruction. Datum is base of West Brook Bed (unit h). Note abrupt thickening of Tully Formation southeast of Winton Road (Loc. 14), and also note conspicuous overstep of Walt Phillips Bed (unit f) and overlying sub-West Brook Bed clastic succession. Uncertainties in correlation between Harrisville (Loc. 12) and Bingham Road (Loc. 13a, b) center on whether key chamositic beds shown actually correlate as shown (see text). Note development of stromatolitic overgrowths on reworked limestone clasts of top-West Brook limestone bed (within unit k) at locality 15; these mark position of regional base-Moravia Bed diastem. Lettered units include: a, upper Cooperstown Member (Spezzano submember-equivalent?) succession; b, sub-Smyrna Bed regional disconformity; c, Smyrna Bed; d, "unnamed sandstone" division of Heckel, 1973; e, medial Tully Formation clastic succession probably equivalent to part of Taughannock Falls succession to west; f, Walt Phillips siltstone bed within unit e – succession; g, "Taughannock Falls Oolite Bed" succession of Heckel, 1973; h, basal erosional contact layer of West Brook Bed = datum; i, West Brook Shale succession; j, limestone bed capping West Brook Shale yielding corals, crinoid material and *Spinatrypa*; k, silty, calcareous, (base-Moravia Bed?) lag layer containing phosphatic pebbles and small corals near Sherburne and locally reworked, stromatolite-coated limestone clasts at locality 15; l, gray shale equivalent to lower part of Moravia Bed succession to west; Number sections are listed in APPENDIX; B, Map showing locality distribution. SR = School Road; PVR = Pleasant Valley Road; BR = Bingham Road; WR = Winton Road, WPR = Walt Phillips Road. (fold-out page).



The work also suggests that three recently defined submember units of the Windom/Cooperstown members (Gage Gully, Sheds, Highland Forest submembers in ascending order) are progressively beveled out between Sheds and Sherburne. As understood from existing literature (see Heckel, 1973) and on our recent study of Hamilton sections, the Smyrna Bed rests disconformably on Spezzano submember-equivalent strata within the Cooperstown Member in the Sherburne area (Figures 5, 6).

East of Upperville (Locality 1\*<sup>1</sup>) Tully sections thin significantly towards Sherburne with notable appearance of oolitic chamosite, a deposit characterized by sand-sized discoidal to sub-spherical black grains which often occur in a mud-silt supported context with bioturbated, calcareous debris (Figures 5, 6, 9, 10). The Smyrna Bed changes eastward from skeletal limestone near DeRuyter to sandy chamosite in the Sherburne area (Heckel, 1973). However, a second major chamosite bed (the "Taughannock Falls Oolite Bed"), believed by Heckel (1973) to be equivalent to part of the Taughannock Falls Beds of the medial Tully limestone further west, appears in Sherburne area localities (Figures 2, 3). An intervening terrigenous unit, the "unnamed sandstone" of Heckel, 1973, was recognized and mapped both by Heckel (1973) and by us (see Figures 2, 3, 5, 6).

Although we formally follow Heckel's correlation scheme, there is a real possibility that the Smyrna Bed of the Smyrna area actually connects to the Taughannock Falls oolite bed as one proceeds from locality 3 to locality 4 (Figures 5, 6). If this is true, the "unnamed sandstone" and the oolite bed beneath it represents a portion of the lower Tully "peeking through" beneath the sub-Smyrna unconformity.

East of Sherburne, Heckel (1973) noted the dramatic thickening of the sub-West Brook Bed Tully succession into a thick clastic wedge (Figure 3). However, the actual nature of the thickening was sketchy owing to availability of few sections. Work by the present authors led to discovery of many new outcrops near the hamlets of Columbus, Shawler Brook, Five Points and New Berlin (Figures 6, 7, 10). Heckel (1973) referred to the area of very thin Tully sections in the Chenango Valley as the "Sherburne high" and the occurrence of the chamosite as representing agitated oolitic shoals. The region of greatly thickened TFCCS deposits south of Columbus suggested the presence of a "down-to-the-east" fault resulting in thick sediment accumulation in a clastic trap (Heckel, 1973). This interpretation is supported by the present authors; our work further constrains the position of the apparent structure (Figures 6, 7).

Cooper and Williams (1935) and Heckel (1973) both believed that western New York Tully divisions should theoretically be present in the TFCCS succession given that the lower part of the TFCCS yields typical lower Tully guide fossils (Figures 1, 3). Work by the present authors shows that the Smyrna Bed can be traced eastward to the Pittsfield area and, more problematically to the Otego Valley (Figures 8, 10). We believe that lower Tully divisions appear beneath the Smyrna Bed in the vicinity of Walt Phillips Road and thicken eastward to locality 17 (Cooper and Williams, 1935, "Perrytown section"). Strata probably equivalent to the Fabius and Meeker Hill beds as well as even lower beds of the New Lisbon Member are present at this section (Figures 7, 10). However, information at locality 18 (Stop 6) and localities near New Berlin suggest that the lower Tully component of the TFCCS is again overstepped in the vicinity of that

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<sup>1</sup>\*Localities numbered in text are listed in APPENDIX.

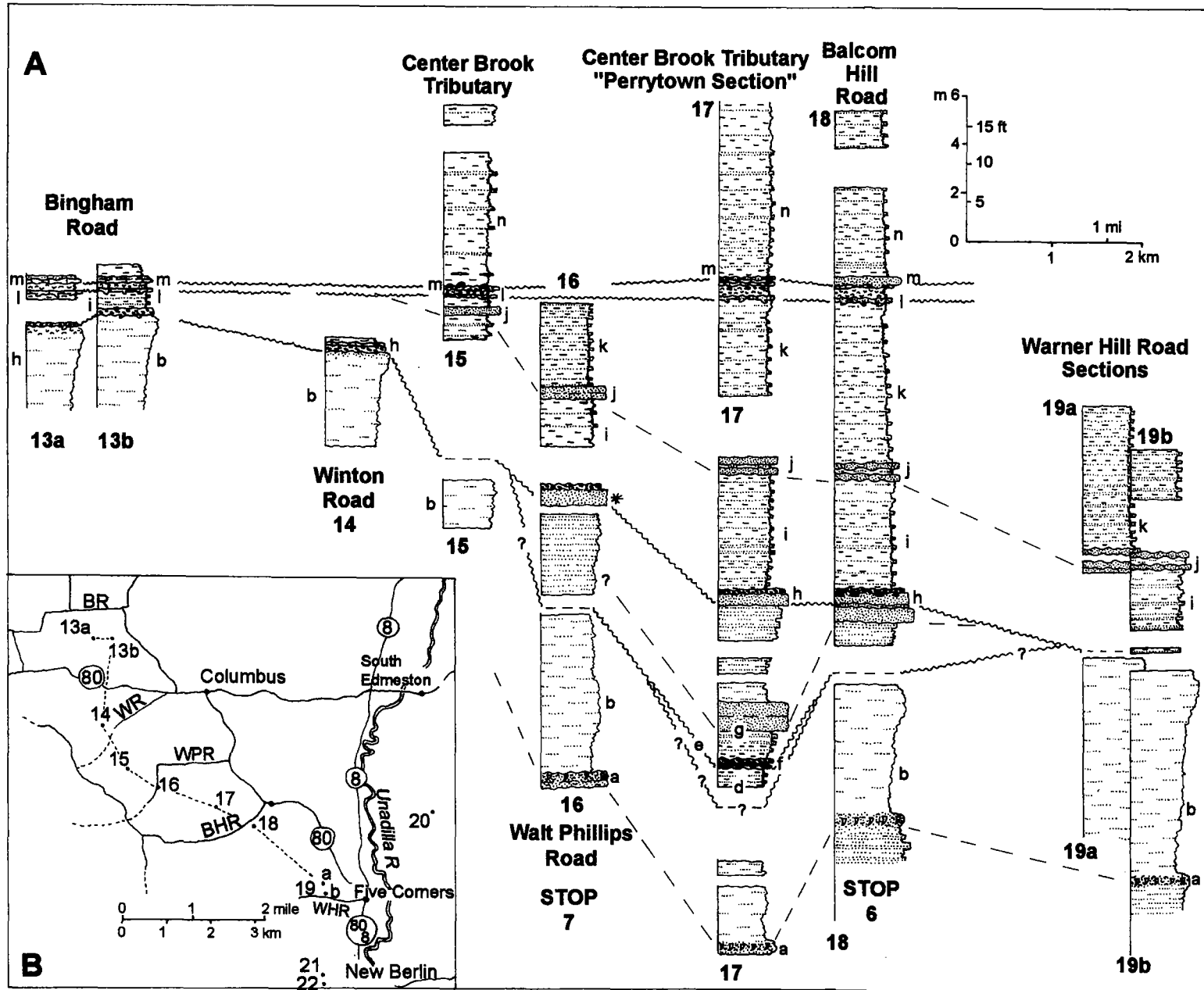
town (Figures 7, 8, 9). Hence, we refer to the small area where these units are visible below the Smyrna Bed as the "Columbus Sag".

Strata of the medial TFCCS between the Smyrna Bed and the base of the West Brook Bed thicken greatly to the southeast of locality 13 (Figures 6, 7). These sparsely fossiliferous beds, probably laterally equivalent to the Taughannock Falls bed-interval of the Tully Limestone, are abruptly cut out by an erosion surface at the base of the West Brook Bed west of locality 17; at locality 17, 37 feet of this interval is present; at locality 13, only 2.2 feet is observed with demonstrable northwestward overstep of an internal, *Rhysochonetes* and *Emanuella*-bearing marker unit herein designated the Walt Phillips Bed (Figures 6, 7). To the south and east of locality 17, the fossil-poor, medial TFCCS strata thicken and coarsen from the Unadilla Valley to the Otego Valley (Figures 7, 8). East of "Greens Gulf" (Locality 24) south of New Berlin the sub-Smyrna Bed TFCCS interval reappears and thickens eastward to the Otego Valley (Figure 8). The Smyrna Bed yields dilute chamosite, bioclastic nodular limestone nodules and minor phosphorite that is characteristically concentrated into burrow prods at localities 24, 27a and 28 (Figures 8, 10); this lag is associated with westward cut-out of lower Tully beds and most of-, or all of, the New Lisbon Member (Figures 8, 9).

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Figure 7. The Tully Formation stratigraphic transect across inferred Columbus area structural sag; A, Uppermost Moscow Formation-into-upper Tully succession showing unit matches between sections. Datum is base of West Brook Bed (unit L). Note overall southeastward thickening of Tully Formation and localized presence of probable New Lisbon Member (unit d) and lower Tully divisions up to disconformity at base of Smyrna Bed (unit h) between localities 15 and 19a. Asterisk at section 16 denotes loose (glacially dislodged) block believed to be Smyrna Bed; a chamositic interval in the block contains *Rhysochonetes* and a stromatolitic limestone layer. Note thin, persistent character of West Brook Bed within thick synjacent, basinal facies. Lettered units include: a, Lansing Coral Bed; b, Spezzano submember-equivalent part of upper Cooperstown Member; c, inferred base-New Lisbon Member discontinuity (concealed); d, New Lisbon Member; e, base-"Laurens" disconformity; f, nodular, chamositic and sideritic, condensed lag bed. The marks first appearance of the distinctive Tully Fauna. The level correspond to base of Fabius Bed to west; g, massive, hard, bioturbated siltstone layer. This may correspond to upper part of Fabius Bed; h, Smyrna Bed and associated base-Smyrna Bed disconformity; i, lower part of barren, flaggy shale-siltstone succession. This is believed equivalent to lower part of the Taughannock Falls Bed-succession in west; j, Walt Phillips Bed (bioturbated siltstone) yielding highest occurrence of *Rhysochonetes* in area; k, upper part of barren, flaggy shale-siltstone succession. This is believed equivalent to upper part of Taughannock Falls Bed succession in west; l, West Brook Bed. Regional discontinuity at base of West Brook is datum; m, discontinuity and silty lag bed capping West Brook Bed. This probably correlates to base of Moravia Bed succession in west. Numbered localities are listed in APPENDIX; B, Map of area showing distribution of localities. BR = Bingham Road; WR = Winton Road; WPR = Walt Phillips Road; BHR = Balcom Hill Road; WHR = Warner Hill Road





From the Gross Hill School section (Locality 29) westward to New Berlin and also to lower Tully sections near Columbus, the base of the observed TFCCS does not correspond to the lowest divisions (DeRuyter and Cuyler beds) of the classic Tully, but rather to higher layers corresponding, most probably, to the Fabius Bed and Meeker Hill Bed divisions of the Tully Limestone (Figures 8, 11). DeRuyter Bed- and Cuyler Bed-equivalent strata, corresponding to the "Tinkers Falls Member" of the basal Tully Limestone succession of Cooper and Williams (1935) make their eastward appearance at New Lisbon (Locality 34) and in the Otego Valley (Localities 36, 38) near Laurens (Figures 8, 11). Cooper and Williams (1935) originally included these basal TFCCS strata in their upper, *Rhysochonetes-aurora* and *Tullypothyridina*-bearing part of their "New Lisbon Member" at New Lisbon (Locality 34). We exclude these beds from our revised (restricted) New Lisbon Member succession and assert, along with Heckel, 1973, that they correspond roughly to Cooper and Williams' (1935) Tinkers Falls Member, and to Heckel's, 1973, DeRuyter and Cuyler beds (Figures 1, 3, 8, 11).

Thus, from Gross Hill School (Locality 29) westward to New Berlin and Columbus, observed sub-Smyrna Bed strata corresponds to a portion of Cooper and Williams, 1935, "Laurens Member". Cooper and Williams (1935) defined the "Laurens Member" to include TFCCS strata often yielding more diverse Tully fauna elements (*Tullypothyridina*, *Schizophoria*, *Spinatrypa*, *Pseudoatrypa*), that occurred above "*Leiorhynchus*"-dominated deposits ("New Lisbon Member" of original understanding) and below strata ("*Elytha fimbriata* zone") yielding Hamilton fauna taxa and lacking the distinctive Tully Fauna brachiopods (Figure 1). Thus, this "member" was largely based on zonal fossil content and is not a real lithologic entity as members are understood to be today. Because we have not yet really named member-scale units in the TFCCS succession, we use the term "Laurens Member" in quotation marks only, and herein restrict the interval to include only the lower-to-medial TFCCS interval generally characterized by bioturbated muddy siltstone to fine sandstone deposits yielding a variety of Tully Fauna taxa (Figures 8, 11). As presently understood, this interval has a sharp, often erosional, base upon "*Leiorhynchus*"-rich deposits of the lowermost TFCCS interval at New Lisbon (Locality 34) and near Laurens (Locality 38) and the New Lisbon Member from locality 29 westward (Figures 8, 11). This regional contact, yielding dilute chamosite in some sections, is herein referred to as the "base-Laurens disconformity". This contact marks the base of the TFCCS at localities 17, 27a, 28 and 29 (Figures 8, 11). The top of the Laurens is placed at the top of the Smyrna Bed or its eastward equivalents. Excluded from the revised Laurens is a thick unnamed (medial TFCCS) interval of sparsely fossiliferous strata between the Smyrna Bed and the West Brook Bed that are believed to be the eastern equivalents of the Taughannock Falls Bed.

Cooper and Williams (1935) placed strata in the upper TFCCS interval in the "*Elytha fimbriata* zone". Generally, this succession correlates to the upper part of Heckel's (1973) Taughannock Falls Bed, West Brook Shale, Moravia Bed, and, possibly, the Fillmore Glen Bed-division succession of the Tully Formation. Generally, TFCCS strata equivalent to the Taughannock Falls and Moravia Bed are expressed as thick, sparsely fossiliferous highstand facies somewhat resembling the progradational Penn Yan and Sherburne members of the overlying Genesee Formation. However, the West Brook Bed stands out in dramatic contrast to all TFCCS facies in being a richly fossiliferous

unit bounded by comparatively barren strata (Figures 5 – 8). Moreover, with exception of locality 26, the West Brook Bed is usually only a few feet thick and is bounded by sharp lower and upper contacts (Figures 5, 8). This unit, when encountered, is unmistakable in sections and constitutes a major datum in mapping. At localities 34 and 37, the base of the West Brook is an encrinitic limestone yielding rugose and tabulate corals as well as reworked concretions, and the overlying shale at all observed localities yields a diverse Hamilton Fauna biota dominated by filter feeding taxa including: brachiopods, bryozoans, camerate crinoids and blastoids. We envision the base of the West Brook Bed to be a major lowstand unconformity. The relative independence of this bed from the pattern of thickening and thinning of surrounding units is striking, both, within the TFCCS and across the central Pennsylvania region where the West Brook is also thin and well developed (Baird and Brett, 2003).

Extending eastward from the Tully Valley to locality 18 (Stop 6) near Columbus a persistent 2 – 8 inch-thick limestone bed yielding crinoidal debris, *Spinatrypa*, corals and both camerate crinoids and blastoids is observed to directly overly the West Brook Shale. Though we are not certain of the match at present, we believe that this unit correlates westward to a position between the Bellona Bed and the massive white limestone facies of the Moravia Bed. This unit is partly to nearly completely overstepped by a still higher thin layer at several localities (Figures 5 – 7). This higher unit, informally designated the “base-Moravia discontinuity bed”, can be traced from localities 3 and 4 southwest of Sherburne to Pittsfield (Locality 27b). This bed is silty to sandy in character, yields phosphatic pebbles and small corals, and appears to be weakly chamositic from Columbus southeast to Pittsfield. At locality 15 southwest of Columbus, reworked clasts of the post-West Brook Limestone bed within this unit are encrusted by stromatolitic overgrowths (Figure 6, 7). At Stop 6 (Locality 18) the post-West Brook limestone bed is partially overstepped and both carbonate clasts and possible chamosite can be found within the lag interval. This lag horizon is believed to link westward to the base of (or higher position within-) the Moravia Bed.

Between the Tully Valley and South Lebanon the Geneseo Black Shale rests unconformably on the Tully Formation in most sections and, locally, lenticular lag concentrations of detrital pyrite occur along the contact. At South Lebanon, a 2.5 foot-thick Moravia Bed-succession is directly succeeded by black, laminated Geneseo. However, to the southeast of South Lebanon the Moravia Bed interval thickens and grades to shale with the top contact becoming conformable from Upperville (Locality 1) eastward. The Geneseo Member thins eastward to only 11 – 12 feet in thickness at localities 5, 7 and 11 near Sherburne but is still tentatively recognized at locality 15 and, possibly, even at locality 26 southeast of New Berlin. East of locality 26, no dark, Geneseo-type facies was encountered. In the New Lisbon area (Localities 34, 35) strata of the uppermost TFCCS interval above the West Brook Bed begin to yield numerous fossils offering an opportunity to examine shelly faunas from levels higher than those examined in previous studies.

## **KEY UNITS AND ISSUES**

### **Introduction**

This section of the report focuses on particular aspects of several key stratigraphic (and chronostratigraphic?) divisions in order from oldest to youngest. Definitions of some units are clarified or revised and a few named beds are discussed.

***Topmost Cooperstown Member Divisions (see Figures 4 – 8)***

***Lansing Coral Bed(s)*** – Several localities in the study area display one to two sandy coral, bryozoan and large brachiopod-rich beds marking the top of a slabby, siltstone interval rich in large *Alanella*. The *Alanella*-rich interval and the overlying coral beds are herein believed to correlate westward respectively to the Taunton submember and Lansing Coral Bed within the Windom Member (Figures 7, 8, 11). Why there are two discrete, closely spaced coral-rich beds above the *Alanella*-rich interval in the Columbus-New Berlin area instead of one further west is still being investigated. In this region, the coral beds are expressed as silty-sandy ledges rich in pelmatozoan debris, occasional large rugosans and *Favosites*, brachiopods, including: *Spinocyrtia*, *Tropidoleptus*, *Rhipidomella*, *Spinatrypa* and *Mediospirifer*, numerous bivalves and occasional bryozoans. The Lansing Bed is understood to mark a lowstand event with a possible sequence boundary at its base (Brett, et al., 1994, Baird and Brett, 2003; see Figure 4).

***Post-Lansing Hamilton Succession (unnamed)*** – From Sherburne eastward into the Otego Valley, the post-Lansing-Pre-New Lisbon Member succession is characteristically a brown, olive-gray weathering, chippy to massive, variably hard silty mudstone to muddy siltstone succession (Figures 4 – 10). Cooper and Williams (1935) reported *Pustulatia* (*Vitulina*) and a few other Hamilton taxa from this interval at several localities in the New Berlin-Laurens area. The present authors found *Pustulatia* in this interval at New Berlin and Laurens, but not at New Lisbon as reported by Cooper and Williams (1935); at the type New Lisbon section (Localities 32, 34) this interval is entirely covered. The post-Lansing Cooperstown succession, ranging from 15 to 46 feet in thickness, yields a rather depauperate, nondescript mixture of Hamilton brachiopods and bivalves. The most common taxa encountered are small to medium size *Alanella* and protobranch bivalves including *Nuculites* and *Paleoneilo*. Other forms encountered include small *Tropidoleptus*, *Mucrospirifer*, *Pustulatia* and “*Eumetabolatoechia multicostatum* (formerly “*Leiorhynchus*” *multicosta*). The last entity, although rare and spotty in distribution, was encountered in this unit south of Columbus (Locality 17), south of South Edmeston (Locality 20) and near Laurens (Locality 38). *E. multicostatum* suggests the possibility that beds containing it are the eastward equivalents of the Gage Gully submember of the Windom Member. At the very least, the occurrence of small brachiopods and nuculoid bivalves indicates that this interval is transgressive relative to the underlying Lansing Bed. Near Columbus (Localities 13, 17) and south of Pittsfield (Locality 27a), the top of this interval is marked by a thin, silty, falls-capping unit yielding larger *Alanella* and *Spinocyrtia*. Depending on how the Cooperstown Member stratigraphy is constructed, this unit could be a regressive cap division of the Spezzano submember (below the Gage Gully interval if the New Lisbon Member is Gage Gully Member-equivalent; see Figures 8, 11A) or the lower part of the Sheds submember (if the New Lisbon Member is younger than Sheds; see Figure 11B).

***New Lisbon Member***

The New Lisbon Member, as originally described by Cooper and Williams (1935) from its type section (Localities 32, 34) northeast of New Lisbon, consists of approximately 60 feet of thin-bedded, flaggy siltstone-shale facies yielding numerous *Camarotoechia* (“*Leiorhynchus*”) *mesacostale* in its lower half and abundant *C. mesacostale* with rare *Rhyssochonetes* and *Tullypothyridina* in the upper half. As originally constructed, this unit is “Tully age” in the upper part and problematically “pre-Tully” in the lower half.

Recently, however, the present authors have identified a thin, contact bed yielding dilute chamosite and diagenetic siderite in association with abundant *Emanuella* and rare *Tullypothyridina* 28.5 feet below the top of Cooper and Williams' (1935) type New Lisbon section along Otto Stahl Road (Locality 34; Stop 8). This bed marks a physical contact (condensed bed/discontinuity) that allows for redefinition of the "New Lisbon Member" to include only strata in the lower half of the original unit. Hence, we herein restrict New Lisbon to include the flaggy leiorhynchid-dominated beds between the underlying typical Cooperstown facies and the base-Tully Formation contact (Figures 8, 11). To date, this "restricted" New Lisbon Member interval has yielded no bona fide Tully Fauna elements such as *Rhyssochonetes* and *Tullypothyridina*, although *Emanuella* of uncertain species assignment has been found in the lower part (Locality 32) and at the base (Locality 31) of the unit.

In the vicinity of the New Lisbon type section (Localities 21, 32, 34) the restricted New Lisbon Member interval is believed to be about 30 feet-thick although no one section exposes the entire interval (Figure 8). As noted above, we observe a significant discrepancy between the description of Cooper and Williams (1935) and what we actually observed at the New Lisbon type section (Localities 32, 34). They described *Pustulatia* ("*Vitulina*")-bearing Hamilton beds "23 feet" below exposed portions of New Lisbon in implied continuity with the New Lisbon succession. What we actually saw included only the topmost 10 feet of restricted New Lisbon Member along Otto Stahl Road (Locality 34; Stop 8) and a 20 foot-thick "hanging" New Lisbon Member section (Locality 32) along Stony Creek (Figure 8). No strata between the partial New Lisbon Member sections and the Lansing Bed have been observed by the present authors. Hence, the base of the New Lisbon Member succession is concealed at the two type locality sections. However, our recent mapping has led to discovery, not only of the top contact at locality 34 (Stop 8), as noted above, but also the basal contact with the underlying Cooperstown division at locality 31 as well (Figure 8). This past April, a third section (Locality 31) was found less than a mile from the Otto Stahl Road section (Locality 34); this shows the New Lisbon Member in sharp contact with Cooperstown Member strata yielding *Alanella*, *Mediospirifer* and numerous small bivalves. This contact, characterized by dilute chamosite in association with nodular, diagenetic siderite (Figures 8, 9), marks a regional discontinuity that can be traced westward to the Unadilla Valley (Figures 8, 11). East of the Butternut Valley north of Laurens, the base of the New Lisbon Member appears to be conformable and apparently non-chamositic at locality 36 (Figure 8). In the Unadilla Valley (Locality 20, 24) and west of there (Locality 17), the New Lisbon Member is believed to be very thin. South of Columbus (Locality 17) no more than three feet of green, gray fissile shale and thin siltstone beds yielding leiorhynchids (*C. mesacostale*?) are exposed with the base concealed (Figure 7). At "Greens Gulf" (Locality 24) only three inches of chamositic green shale with leiorhynchids is present between the Cooperstown Member and the Smyrna Bed (Figures 8, 9). Although the New Lisbon Member is missing on the east side of the Otego Valley (Locality 38), a fact originally noted by Cooper and Williams (1935) flaggy siltstone facies resembling New Lisbon Member is observed above typical Cooperstown facies in the Susquehanna Valley east of Milford and, still further east near Schenevus. These outcrops are still the object of reconnaissance study by the present authors.

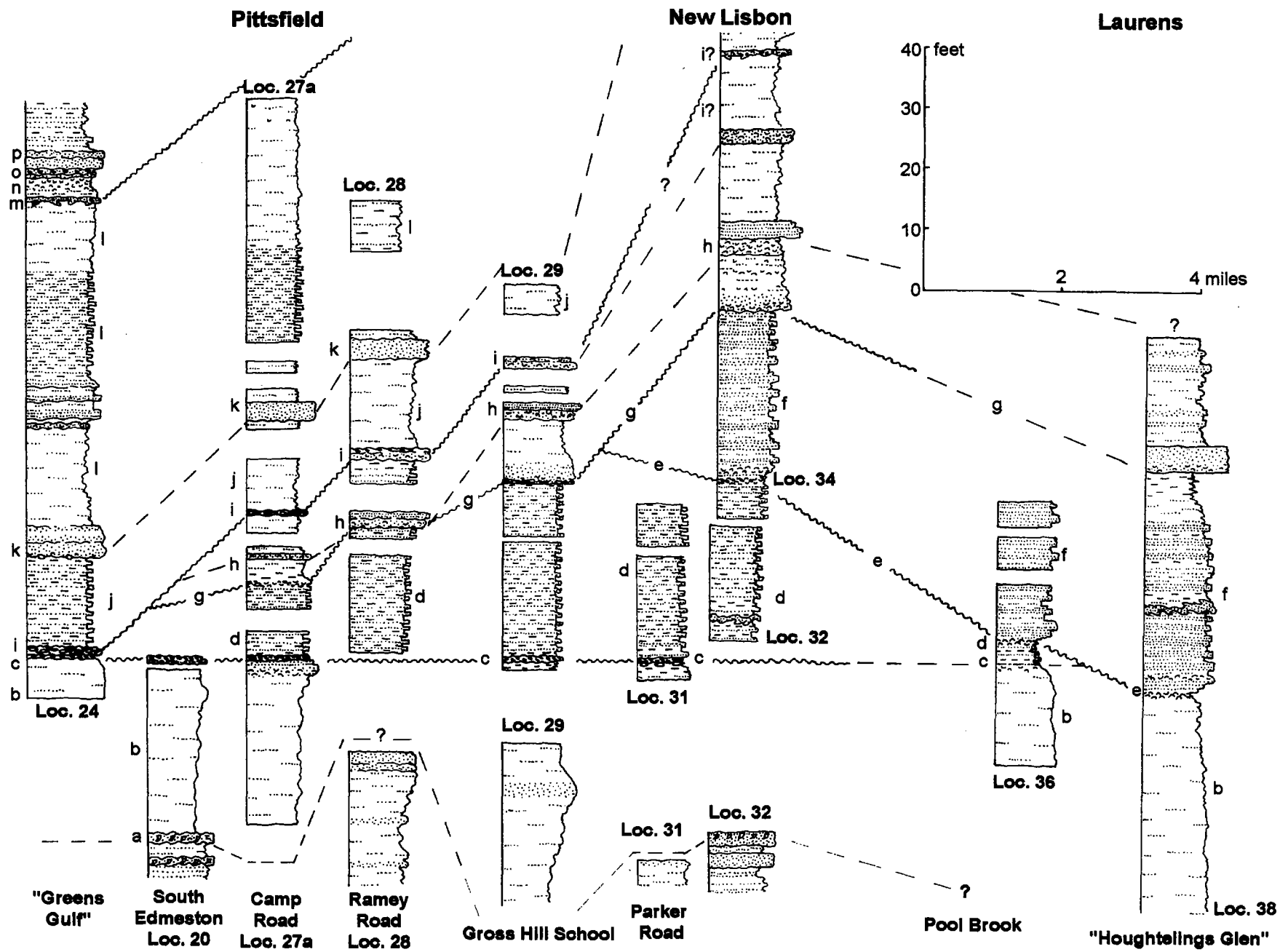
The basal New Lisbon contact bed is chamositic at several sections (Localities 20, 24, 27a, 29, 31) and it is spectacularly so at localities 27a and 29 (Figure 9). The fauna of this bed is extremely sparse and poorly preserved. Above the chamosite layer, macrofossils are almost exclusively leiorhynchids (a handful of small *Emanuella* were observed along Stony Creek [Locality 32] at one level). The flaggy siltstone-shale alternation that characterizes this unit resembles intervals within several other dysoxic aggradational deposits such as the Butternut Member in the medial Hamilton Group and the Sherburne Member within the Genesee Formation (Thayer, 1974; Baird et al., 1988, 2000). The New Lisbon Member marks a significant sea level highstand event. It correlates westward either to the Gage Gully submember or to the younger Highland Forest submember of the Windom Member depending on how correlations are reconstructed (Figure 11).

#### TFCCS Divisions

**Basal-Tully Division (DeRuyter-Cuyler beds-equivalent part of TFCCS)** – This interval of strata is solely represented by strata at localities 34, 36 and 38 (Figures 8, 11). It is represented by the “quarry sandstone” succession discussed by Cooper and Williams (1935) at locality 34, and by a similar interval of bluff-forming, slabby, cross-laminated siltstone or fine sandstone at “Houghtelings Glen” (Locality 38). The basal 2 to 3 feet of this interval yields abundant *Emanuella* as well as less common *Rhyssochonetes* and

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Figure 8. Lower and medial Tully Formation stratigraphy from vicinity of New Berlin in the Unadilla Valley to vicinity of Laurens in the Otego Valley northwest of Oneonta. Datum is basal contact of New Lisbon Member (horizon c) at top of typical upper Cooperstown Member (Spezzano submember-equivalent?) succession (unit b). Note conspicuous westward and eastward erosional overstep of New Lisbon Member (unit d) by Tully Formation units and overall westward thinning and overstep of lower Tully units east of New Lisbon such that probable Smyrna Bed (unit I) is nearly juxtaposed on Cooperstown Member south of New Berlin (Loc. 24), see text discussion. Lettered units include: a, Lansing Coral Bed; b, uppermost typical Cooperstown Member (probable Spezzano submember equivalent) succession; c, basal discontinuity (and continuity) flooring New Lisbon Member with associated chamosite and diagenetic siderite; d, New Lisbon Member (restricted) succession; e, base-Tully discontinuity with associated shell-hash and dilute chamosite flooring lowest Tully Formation division; f, lowest Tully Formation (probable “Tinkers Falls beds“-equivalent succession *sensu* Cooper and Williams, 1935) division; g, discontinuity and eastward continuity flooring “Lauren Member” succession *sensu* Cooper and Williams (1935). Bed above contact yields dilute chamosite in western sections and abundant shells in eastern sections; h, shell-bearing, bioturbated, muddy siltstone bed probably equivalent to upper part of Fabius Bed to the west. Interval g through h is probably equivalent to entire Fabius Bed succession; i, nodular, calcareous, siltstone bed which yields dilute chamosite at localities 24, 27b, and 28 and abundant brachiopods at localities 34 and 37. This is believed to be expression of Smyrna Bed in this area; units j, k, l, divisional components of medial Tully (probable Taughannock Falls beds-equivalent) succession; units m, n, West Brook Bed divisions; units o, p, q, Moravia Bed equivalent unit divisions. Numbered localities are listed in APPENDIX.



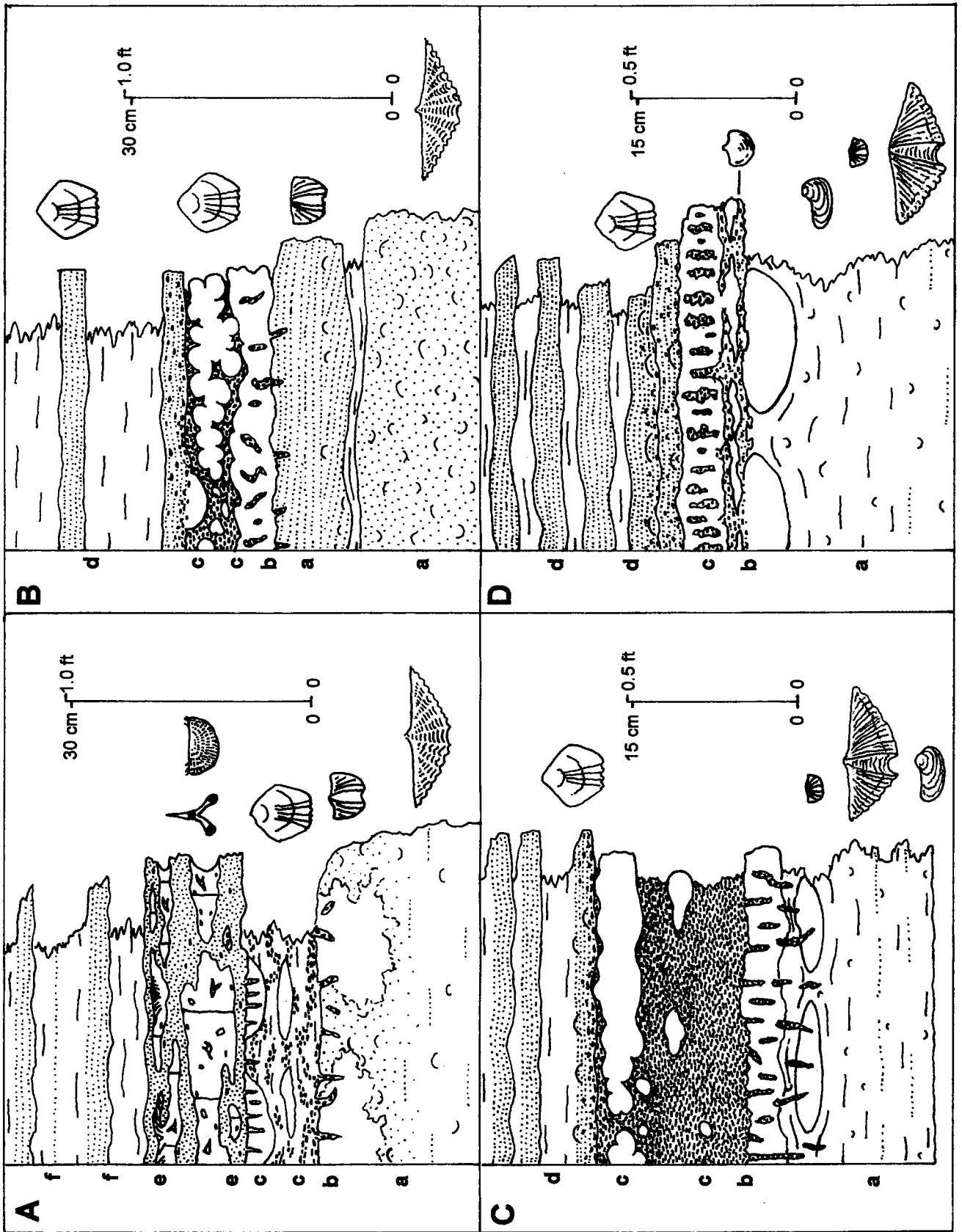
*Tullypothyridina* in association with abundant *Camarotoechia mesacostale*. The remaining 20 – 35 feet of the interval is overwhelmingly dominated by *C. mesacostale* to the near exclusion of other taxa. The facies is problematically coarse but is essentially a leiorhynchid biofacies. We concur with Heckel (1973) who argued that it corresponded to the DeRuyter-Cuyler interval in the Tully Limestone. We suspect that the basal lag and overlying *Emanuella*-rich interval correspond to the DeRuyter Bed and that the overlying *C. mesacostale*-dominated interval links to the Cuyler Bed (Figure 11). The biggest problem with this reconstruction is that no *C. mesacostale* are found in the type DeRuyter-Cuyler interval (Sessa, 2003). However, in central Pennsylvania, there is a basal-Tully black limestone interval abounding in *C. mesacostale* that appears to link to the DeRuyter-Cuyler succession. The basinal character of the Pennsylvania unit suggests that the DeRuyter-Cuyler beds-interval in the Tully Limestone displays upslope shelfward facies relative to equivalent deposits in east-central New York and central Pennsylvania (Baird and Brett, 2003).

**“Laurens Member”** - As asserted by Cooper and Williams (1935), and reviewed by Heckel, 1973, the “Laurens Member” records some improvement in bottom conditions with an increase in bioturbation and the appearance of shell beds in the easternmost sections. The base of the “Laurens” is sharp and probably erosional in all localities examined (Figure 7, 8). At localities 17 and 29 it is characterized by chamosite and diagenetic siderite; at localities 27a and 28, it is marked by reworked clasts and phosphorite; at localities 34 and 38 it marks an abrupt change from flaggy leiorhynchid-rich facies below to massive siltstone-sandstone deposits above yielding a variety of brachiopod and bivalve taxa (Figure 8). The base-“Laurens” contact at

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Figure 9. Base-New Lisbon Member unconformity at four key localities (see also Figure 8); A, “Greens Gulf” (Loc. 24) showing three inches of base-New Lisbon chamosite-rich shale overlain by Smyrna Bed (interval now concealed under shallow alluvium); B, Tributary of Wharton Creek (Loc. 27a) showing spectacular nodular diagenetic siderite “marbled” throughout the black, chamositic basal layer (interval now concealed under shallow alluvium); C, Unnamed gully northwest of former Gross Hill School (Loc. 29) showing chamosite with associated nodular diagenetic siderite; D, Tributary of Stony Creek (Loc. 31) near New Lisbon, NY showing easternmost development of chamositic and sideritic sub-New Lisbon Member contact. Note occurrence of thin *Emanuella*-rich layer coincident with base of chamositic bed (see text). Lettered units include: a, uppermost typical Cooperstown Member (probable Spezzano submember equivalent) succession yielding typical upper Hamilton Group taxa such as *Mucrospirifer*, *Mediospirifer*, *Alanella*, and nuculoid bivalves; b, erosional contact showing chamosite disc-infilled open burrows (or borings) as well as underbed diagenetic siderite nodules; c, black, shiny concentration of discoidal (flattened?) chamositic “oids” and sideritic nodules; d, interbedded, flaggy shale-siltstone facies of basal New Lisbon Member yielding Leiorhynchids, cf. *Camarotoechia mesacostale*; e, Smyrna Bed containing dilute chamosite, phosphatic nodules, nodular limestone and possible stromatolite development at top. Unit yields abundant auloporids and rare *Rhysochonetes*; f, barren flaggy shale-siltstone interval of post-Smyrna strata. Numbered localities are listed in APPENDIX.





Houghtelings Glen (Locality 38) is at the base of the massive falls-capping bed near the upper end of that section, yielding a variety of taxa (*Tullypothyridina*, *Orthospirifer mesastrialis*, *Stropeodonta*, large pteroid bivalves) that preview the little known proximal Tully biofacies to the east.

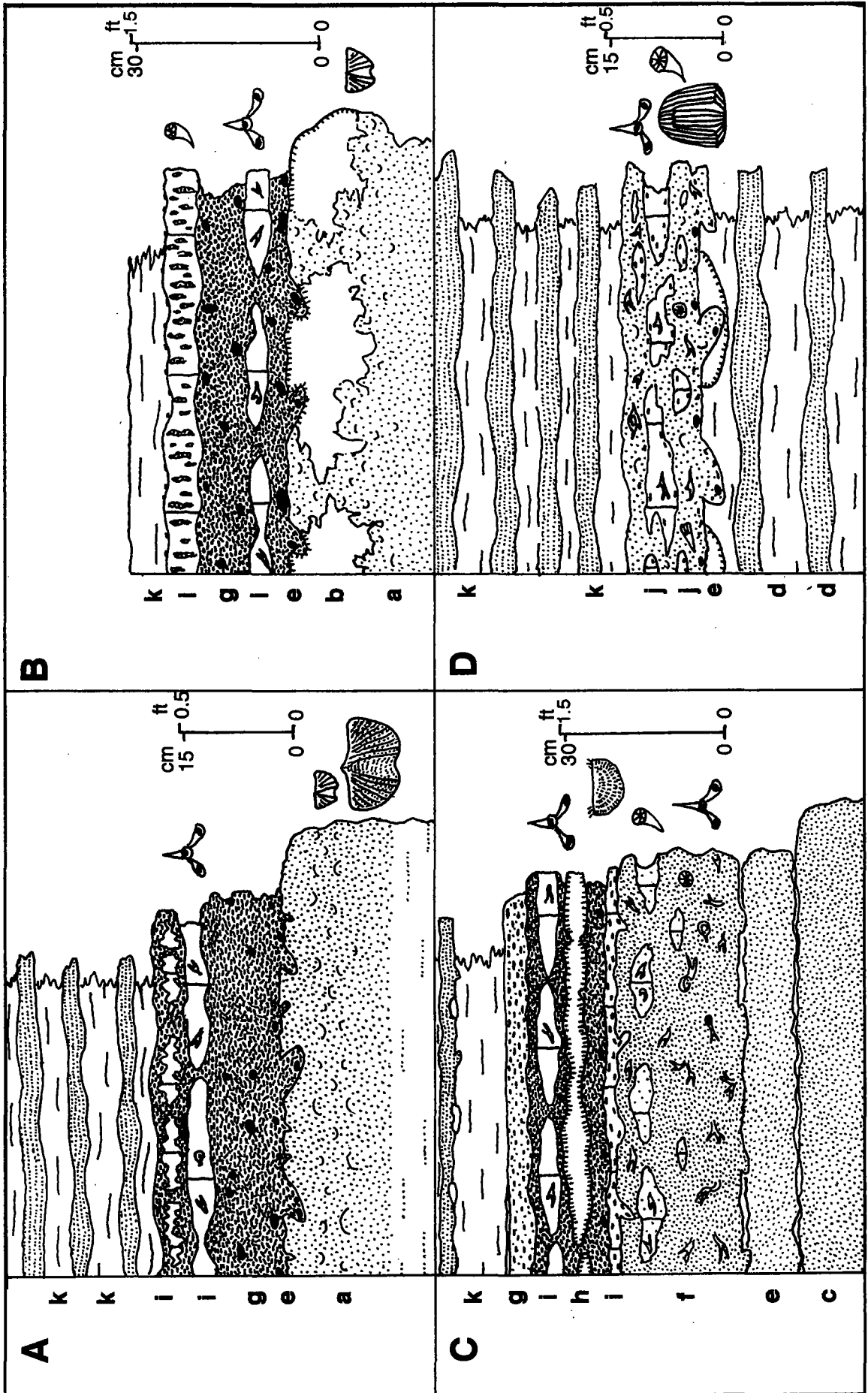
At Otto Stahl Road (Locality 34, Stop 8) Cooper and Williams (1935) described two fossil-rich units (respectively a “second” and “third” shell bed) that we have also located in that section (see Stop 8 description) along with still higher shell bed as well. Although correlations are tentative, we link the “base-Laurens” disconformity-through-“second shell bed” interval to the Fabius Bed and the “third shell bed”-to-the highest shell bed to the Tully Valley Bed-Smyrna Bed interval (Figures 8, 11), supported both by the lithofacies succession and by the mix of lower Tully Fauna taxa.

**Smyrna Bed** – The top of the “Laurens Member” is herein placed at the top of the Smyrna Bed and its possible eastern equivalents (Figure 8). Although, this violates implied principles of sequence boundary usage, the assignment is informal due to uncertainties in correlation. Few mappable boundary strata are recognized in the thick “barren” medial Tully interval between the Smyrna and the West Brook Bed. Hence, the top-Smyrna surface is as good a lithologic change as any so far observed. Moreover, at localities 17, 18, 24, 27a, 28 and 29, the Smyrna Bed is thin and the effect of the arbitrary placement minimal.

As noted earlier, the Smyrna Bed in the Sherburne-Pittsfield region is a nodular, calcareous, chamositic sandstone bed characterized by hypichnial burrow prods, phosphatic pebbles, and occasional fish debris (Figures 5 – 8, 10). Chamositic grains are sand-sized, mud-supported black, discoidal grains that are typically associated with abraded bioclastic debris in a bioturbated fabric (Figure 10). Nodular lentils of bioclastic limestone characterize the bed east to the Pittsfield area (Localities 27a, 28). Although chamosite is not observed at localities 29 and 35, abundant shells make their eastward appearance in these same sections. We believe that this bed is the Smyrna, based both on

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Figure 10. Smyrna Bed in Tully clastic correlative succession (note that Smyrna Bed at locality 24 is shown in Figure 9A). Locality sequence shown includes: Unnamed gully (Loc. 13b) south of Bingham Road (A); unnamed creek (Loc. 14) west of Winton Road (B); unnamed creek (Loc. 18) to be seen at Stop 6 southeast of Balcom Hill Road (C); and section (Loc. 27a) south of Pittsfield, NY (D). Lettered units include; a, uppermost (Spezzano submember equivalent?) division of Cooperstown Member yielding *Alanella* and *Spinocyrtia*; b, Cooperstown Member siltstone suffused with diagenetic, under-bed siderite related to proximity of sub-Tully contact; c, massive, bioturbated (Fabius Bed equivalent?) siltstone bed in lower Tully Formation succession; d, flaggy shale-siltstone facies in lower Tully “Laurens Member” succession; e, sub-Smyrna Bed erosion surface; f, lower Smyrna, non-chamositic, auloporid-pelmatozoan debris-rich calcareous siltstone bed; g, black chamosite concentration with associated phosphatix pebbles; h, diagenetic siderite within chamosite concentration; i, auloporid and pelmatozoan-rich limestone bands and nodules within chamosite; j, bioturbated siltstone with lean chamosite and bioclastic limestone nodules yielding auloporids, rugosans, *Tullypothyridina* and *Rhysochonetes*; k, barren, flaggy shale-siltstone interval in medial Tully Formation succession. Numbered localities are listed in APPENDIX.



its lithologic character and its regionally disjunct relationship to underlying units (Figures 7, 8). As with the Smyrna Bed further west, it rests on a variety of underlying units in the Columbus-Pittsfield area (Figures 7, 8). In particular, south of Pittsfield (Locality 27a) it rests upon strata above the Fabius Bed; at "Greens Gulf" (Locality 24) near New Berlin, it rests on the basal feather edge of the New Lisbon Member (Figure 9a).

The most exciting discovery associated with the new eastern Smyrna Bed sections are microbiolite stromatolites at the top of the bed at localities 16 and 24 (see Figure 9a; Stop 7 description). A glacial erratic block found in the bed of the creek at locality 16 (Stop 7) corresponds to a dislodged piece of the Smyrna Bed; this is marked by a limestone layer at its apparent top yielding chamosite, fracture networks ("mudcracks") and microbiolitic stromatolites. At "Greens' Gulf" (Locality 24), a 0 - 3 mm-thick crust of stromatolitic carbonate caps the Smyrna Bed at that locality as well (Figure 9A). The associated carbonate at both localities is a white-weathering chamosite-rich limestone which contains numerous auloporids and displays grains often aligned at an angle parallel to laminations or other internal structures. The stromatolite-bearing chamositic carbonate layer is very thin and its top is an undulatory knife-sharp contact with silty mudrock facies (Figure 9A).

Close examination of the microbiolite structures shows that there is a larger-scale lamination, often bordering vertical fractures; laminations are steeply oblique to near-vertical bordering fractures suggesting that much of the stromatolite had since been eroded away leaving only low, peripheral portions adjacent to the low cracks. This type of stromatolite displays an internal micropillar structure normal to the lamination. This type of stromatolite closely resembles that originally described along the intra-Tully unconformity at Bellona by Heckel (1973) (see Stop 11 description), and it is very similar to stromatolites along the same horizon seen by the present authors in the Tully Valley (see Stop 2 description). The second type of stromatolite is a non-chamositic, non-pillared, 1 - 3 mm-thick laminated crust that marks the bed top contact. This orange-red weathering crust is observed not only at localities 16 and 24 but it has recently been observed capping a correlative transgressive surface at Hughesville, Pennsylvania east of Williamsport.

We interpret the fractured, stromatolitic horizon to mark a major deepening event associated with a regional flooding surface marking the top of the Smyrna Bed. This surface corresponds most closely to the very thin, stromatolitic transgressive interval above Heckel's (1973) widespread unconformity that separates the two members in the Tully Limestone. Heckel (1973) correctly noted that the change from the unconformity horizon into the overlying Taughannock Falls Bed-interval marks a significant transgression. Both in the Columbus-New Berlin area and at Hughesville, the stromatolitic level is succeeded by dramatically, thick barren facies that signals the overspread of apparent dysoxic conditions and low energy early highstand facies.

## **DISCUSSION**

### **Regional Paleoenvironmental Overview**

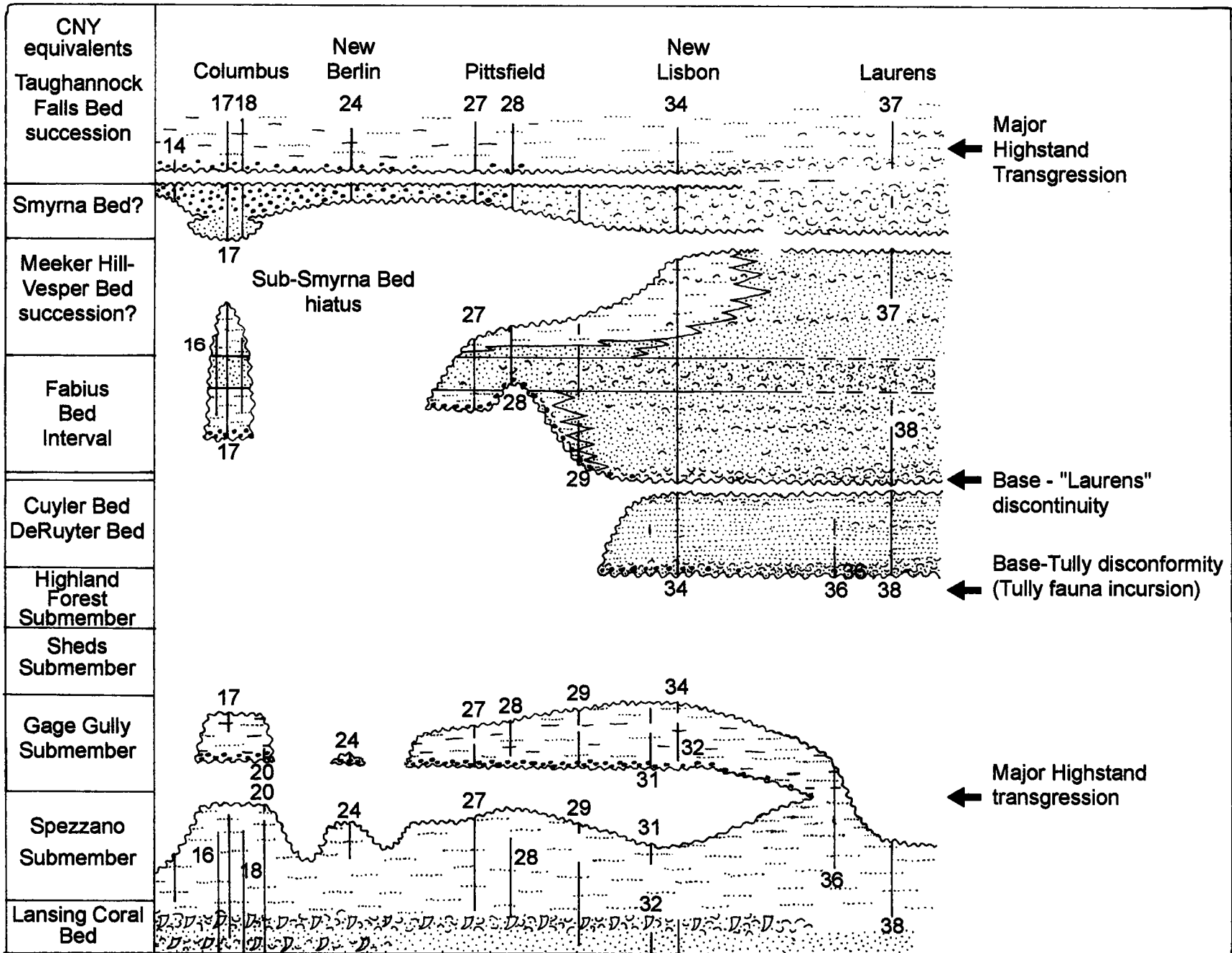
Heckel (1973) argued that several key Tully facies (chamositic beds, stromatolitic-mudcracked horizon in Carpenters Falls Bed) represented marine shoal and even lagoonal conditions. As such, part of the Tully Limestone would have modern

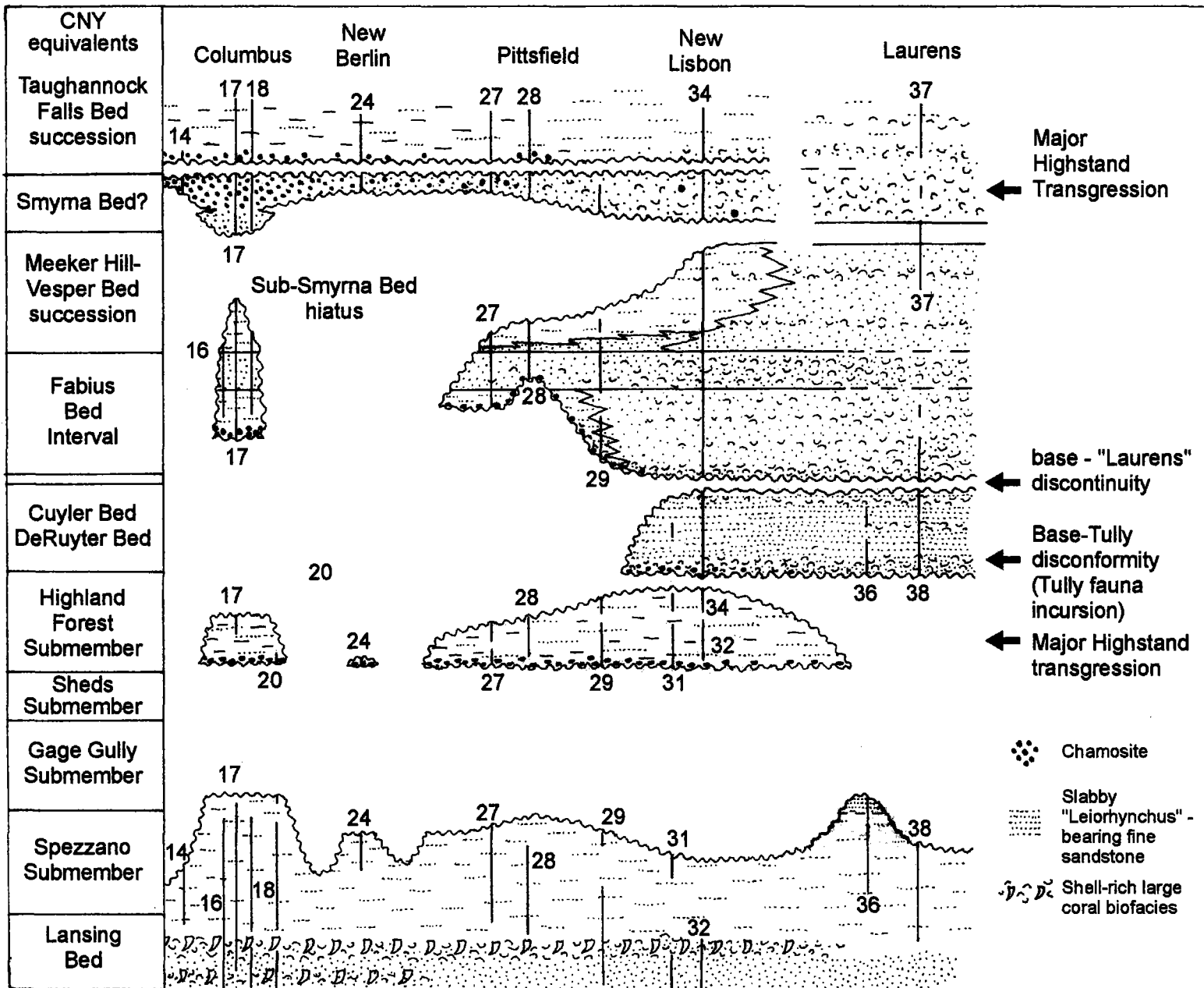
environmental counterparts represented by the Bahama Platform, etc. Comparable facies in New York State would be represented in the Black River Limestone (Upper Ordovician) and in the Lower Devonian Manlius Limestone. The present authors, however, view the Tully Formation as including a broad suite of settings ranging from shallow shelf conditions (Bellona Bed-West Brook succession) to low energy outer shelf settings and even dysoxic basin environments (Baird and Brett, 2003). We believe that the lower and medial Tully succession overall spans a mid-to outer shelf facies range with no preserved lagoonal or "Z" facies observed. We see no stromatoporoids and no pentamerids, often typical of proximal Devonian carbonates, and the stromatolites and "mudcracks" discussed herein are restricted to a single transgressive horizon in contrast to multiple levels typically seen in restricted nearshore deposits. It is very possible that the "mudcracks" are syneresis cracks of diagenetic origin. Stromatolites can occur in deeper subtidal slope environments as well as shallow settings (see Playford, 1980). We suspect that the chamosite, "mudcracks" and stromatolites are a combined depositional and diagenetic effect of carbonate condensation and sediment-starvation associated with widespread transgression. The occurrence of all of these features in close association with thick, unfossiliferous, low-energy, facies successions at New Berlin and Hughesville is more consistent with a comparatively lower energy setting platform margin or outer shelf setting.

As noted above, chamositic, oolitic ironstone facies is quite problematic in any depositional model because it is not convincingly represented by any precise modern analog. Aragonitic ooids, forming in modern carbonate shoals are spherical and often occur in cross-bedded, grain supported deposits (Sellwood, 1986). By contrast, Tully chamosite is characterized by discoidal to elliptical grains that occur in a mud-rich or mud-supported context. Typically the rock is intensely bioturbated, and, along the base of the New Lisbon Member, very dark and shaly (Figure 9). Texturally, the chamositic beds more resemble glauconitic greensand facies except that glauconite grains have a different color and shape. Still other mysteries exist, if chamosite formed under reducing

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Figure 11. Inferred chronostratigraphy of uppermost Moscow Formation-through-lower Tully Formation clastic correlative succession. Note westward expansion of hiatal intervals due to collective overstep by sub-New Lisbon Member-, sub-base-Tully, sub-base "Laurens," and sub-Smyrna discontinuities. Inferred Columbus area structural sag is indicated by localized "hanging bits" of lower Tully in that area; A, Preferred version showing New Lisbon Member as eastern equivalent of Gage Gully submember of Windom Member; B, Alternative scheme showing possibility that New Lisbon Member may be equivalent to the Highland Forest submember comprising the topmost Windom succession southeast of Syracuse. Lettered units include: a, uppermost part of typical Cooperstown Member succession; b, New Lisbon Member; c, sub-base-Tully discontinuity; d, lower part of Tully Formation clastic correlative succession ("Tinkers Falls" division of Cooper and Williams, 1935); e, base-"Laurens" disconformity; f, key fossil-bearing siltstone marker bed in "Laurens Member" succession; g, sub-Smyrna Bed disconformity; h, Smyrna Bed; i, inferred diastem at top of Smyrna Bed (regional flooding surface); j, sparsely fossiliferous highstand clastic facies comprising medial Tully Formation clastic correlative succession. Numbered localities are listed in APPENDIX. (on following two facing pages).





conditions as is argued by Curtis and Spears (1968) why does it occur in such abundance in condensed beds in association with skeletal debris? The concentration of chamosite on the “Sherburne High” is consistent with evidence of winnowing in oolitic ironstones elsewhere; the “Sherburne High” acme of chamosite distribution accords well with the “clastic trap” model of Huber and Garrels (1953) for enhanced chamosite concentration on elevated sediment-starved areas relative to lower areas of differential terrigenous accumulation. Preferential development of chamosite on the east flank of the “Sherburne High” at the western margin of the inferred New Berlin trough and the absence or near-absence, of chamosite on the east margin of that trough (Figure 12) supports the concept that sediment-starvations more than any other presently identifiable factor, allowed the chamosite to become concentrated (Baird and Brett, 2003). As for inferred water depth, chamosite occurs in shelfal settings but accumulates across a broad depth ranges of 10 – 170 meters (Porrenga, 1967) allowing only for a tentative paleoenvironmental “window” of “subtidal shelf” for this facies. (It is interesting to note that there is another deposit of chamositic ooids in the Appalachian Basin that may be coeval with portions of the Tully Formation. The Preston or Boyle Ore of central Kentucky is a chamositic ooid deposit that was the first iron ore mined west of the Appalachian Mountains, being used to make cannon balls of the Battle of New Orleans in the War of 1812.)

Another major pattern we observe within the lower and medial Tully succession is evidence for a region of greater inferred water depth centered on Columbus and New Berlin (Figure 12). This area of differential subsidence is herein referred to as the “New Berlin Trough” with a more local structural feature within it termed the “Columbus Sag” (Figure 7). Tully deposits in the Columbus-Pittsfield area (excluding the West Brook Bed) are sparingly fossiliferous at best. Condensed marker beds yield auloporids and pelmatozoan debris but very few shells. Strata between the marker beds are usually almost barren of macrofossils. Even in the Sherburne area, the sub-west Brook Tully yields very few shells and is notable mainly for auloporids and crinoid debris, a fact also noted by Heckel (1973). The observed rarity of brachiopods may be a taphonomic artifact of shell dissolution associated with condensation and, possibly, chamosite genesis. However the abundance of auloporids and crinoid ossicles, though corroded, indicates that robust shells, if once part of this biofacies, should be observed. Their absence hints at a general unhealthiness of biotas in the Sherburne condensed Tully beds. Examination of lower-medial Tully strata (DeRuyter Bed-Taughannock Fall Bed interval) west of DeRuyter shows that macrofossils are common through this succession (see Stops 1 – 4) in dramatic contrast to the Sherburne sections.

We suspect that the “Sherburne High” of Heckel (1973) may be an east-facing “Sherburne Slope” that recorded sediment-starved conditions on a deeper water sloped sea bed (Figure 12). This model, thus, represents an expansion of Heckel’s (1973) “Down-to-the-east” fault and “clastic trap” concepts (Figure 3). We herein argue that the clastic trap (New Berlin Trough) was bigger and deeper than in Heckel’s (1973) original reconstruction and that it may have connected southwestward to the large Tully basin in north-central Pennsylvania (Baird and Brett, 2003). Concurrent downwarping and filling of this structural ‘moat’ could explain coincident accumulation of Tully platform carbonates in a mud-free, craton-ward regime in west-central New York.



East of the New Berlin-Pittsfield area, the near-barren dysoxic facies of the trough axis is increasingly replaced by shell-bearing deposits. Both the "Laurens Member" and the overlying medial Tully succession yield fossils at many levels from the New Berlin

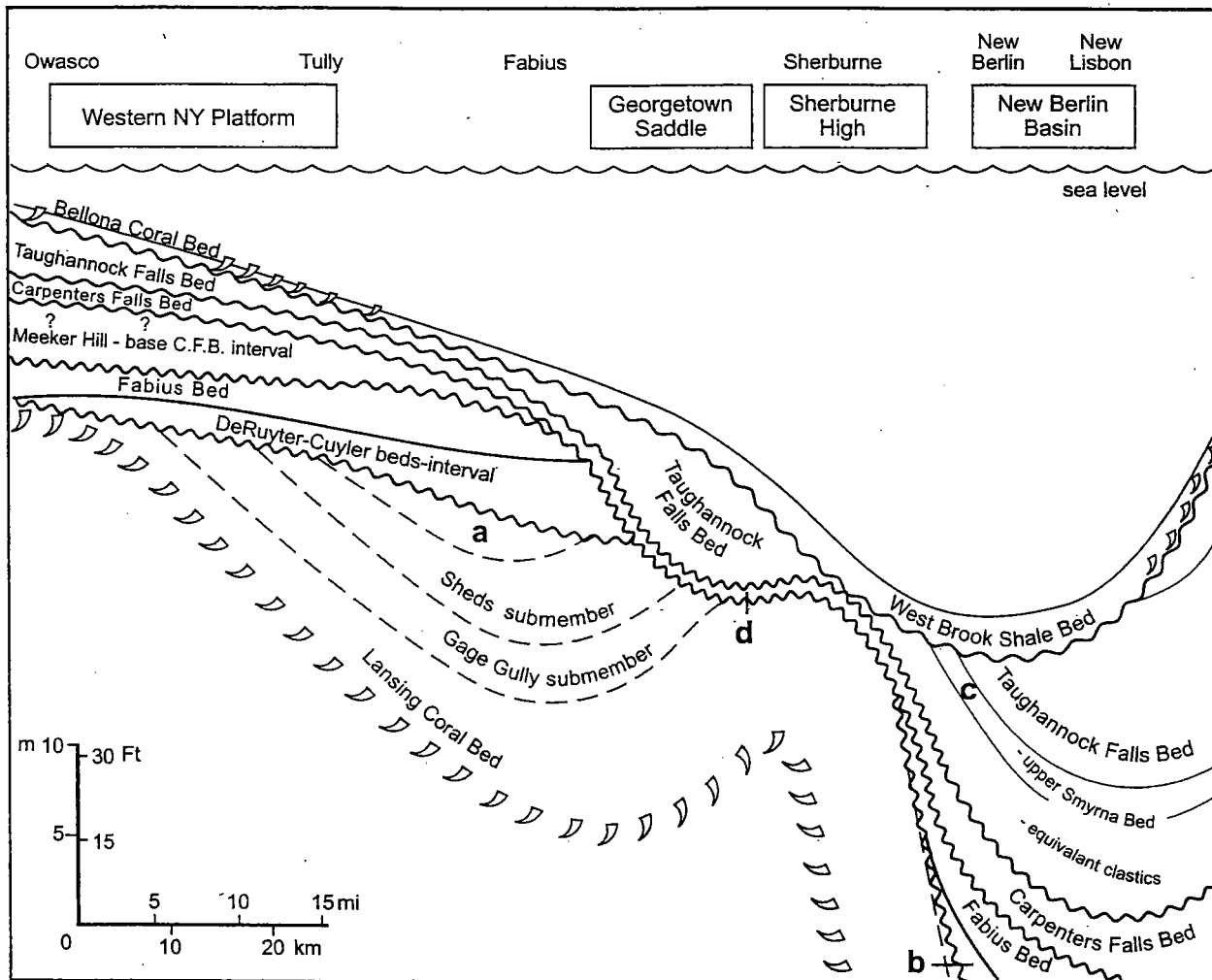


Figure 12. Schematic upper Windom Member-to-upper Tully stratigraphy from west-central New York to the Oneonta meridian. Note prominent erosive effects of the base-Tully, base-middle Tully, and base-West Brook disconformities (see text) and the relationship of component facies units and disconformities in the position of an inferred structural (flexural) basin (New Berlin Trough) centered between Sherburne and New Lisbon. This figure is drafted without a datum to emphasize the west- and east-facing paleoslopes of this basin during deposition of the West Brook Bed. The Taughannock Falls Beds - interval is shown as having the greatest differential paleodepth relief with the east - facing slope commencing well to the west of the "Sherburne High" of Heckel (1973). Lettered units include: a, Highland Forest submember; b, DeRuyter Bed-Cuyler Bed equivalents; c, prominent siltstone bed in medial Tully eastern clastic-succession; d, Smyrna Bed chamositic layer. N. L. = New Lisbon Member; C. B. = Cuyler beds; F. B. = Fabius Bed; C. F. = Carpenters Falls beds; T. F. B. = Taughannock Falls beds. From Baird and Brett (2003).

area (Locality 34) eastward, and all condensed marker beds become notably shell-rich (Figure 12). Although our mapping effort to date has completed work only to the Otego Valley, we recognize that lower and middle Tully facies in that area has a shelly “Chemung” aspect with the appearance of larger brachiopods and bivalves. This “up-to-the-east” faunal gradient serves to show the transition from the New Berlin Trough to an “eastern New York shelf setting” (Figure 12). The characterization of the eastern shelly fauna is work that is still ongoing (see below).

The last, and final, set of questions regarding the Tully clastic succession concerns the temporal chronology of biotic changes associated with the Taghanic Bioevent (see Baird and Brett, 2003). As noted earlier, onset of Tully deposition is closely timed with initial invasion of the Tully Fauna that dominates the lower Tully succession (Figure 13). During upper Tully deposition, the Hamilton Fauna stages a comeback and persists to the close of Tully deposition (Figure 13).

However, in western and central New York, the succeeding unit is the Genesee black shale and any possible connective relationship between Tully biofacies and the higher Ithaca Fauna of the Genesee Formation is obscured. Only where the Genesee Member passes eastward into dysoxic and oxic facies can the intervening faunal story be revealed. This is one goal of ongoing work.

Another goal is to see if the Tully Fauna in the lower Tully succession actually grades into inner shelf, sandy biofacies yielding Ithaca Fauna taxa. Cooper and Williams, (1935) indicated that some “Ithaca-type” taxa were found in Gilboa Formation deposits east of the Susquehanna Valley. It is, thus, possible that nearshore Tully biofacies may have an Ithaca appearance and that the Tully Fauna of the Tully Limestone is actually an outer shelf, carbonate sub-facies of the Ithaca Fauna. This problem will be addressed during the next two to three-year period.

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Figure 13. Faunal succession in the upper Windom Member and Tully Formation succession in east-central New York showing in the influx of the Tully fauna (“Lower Tully bioevent”) at the base of the Tully Formation and reestablishment of the diverse Hamilton fauna association (“Upper Tully bioevent”) in the upper Tully. Key lowstand unconformities include: the base-Tully (1), base-Middle Tully (2), and base-West Brook (4) sequence disconformities. Maximum flooding surface contacts include: the top-Smyrna Bed unconformity (3) and the top-Fillmore Glen corrosional discontinuity (5). Diagnostic and/or common taxa include: a, *Heliophyllum halli*; b, *Spinatrypa spinosa*; c, large bryozoans; d, *Mediospirifer audaculus*; e, *Allanella tullia*; f, *Devonochonetes scitulus*; g, *Camarotoechia* (“*Leiorhynchus*”) *mesacostale* [In the Tully, this taxon is restricted to the Tully eastern clastic correlatives in eastern New York and Pennsylvania]; h, *Emamuella praeumbona*; i, *Pustulatia* (*Vitulina*) *pustulosa*; j, *Tropidoleptus carinatus*; k, *Athyris spiriferoides*; l, *Mucrospirifer mucronotus*; m, *Rhyssochonetes aurora*; n, *Emamuella subumbona*; o, *Schizophoria tulliensis*; p, *Tullypothyridina venustula*; q, *Echinocoelia ambocoeloides*; r, *Spinatrypa sp.*; s, *Pseudoatrypa devoniana*; t, *Ambocoelia umbonata*; u, small rugosan; v, auloporids; w, *Leptaena rhomboidalis*; x, styliolines. From Baird and Brett (2003).

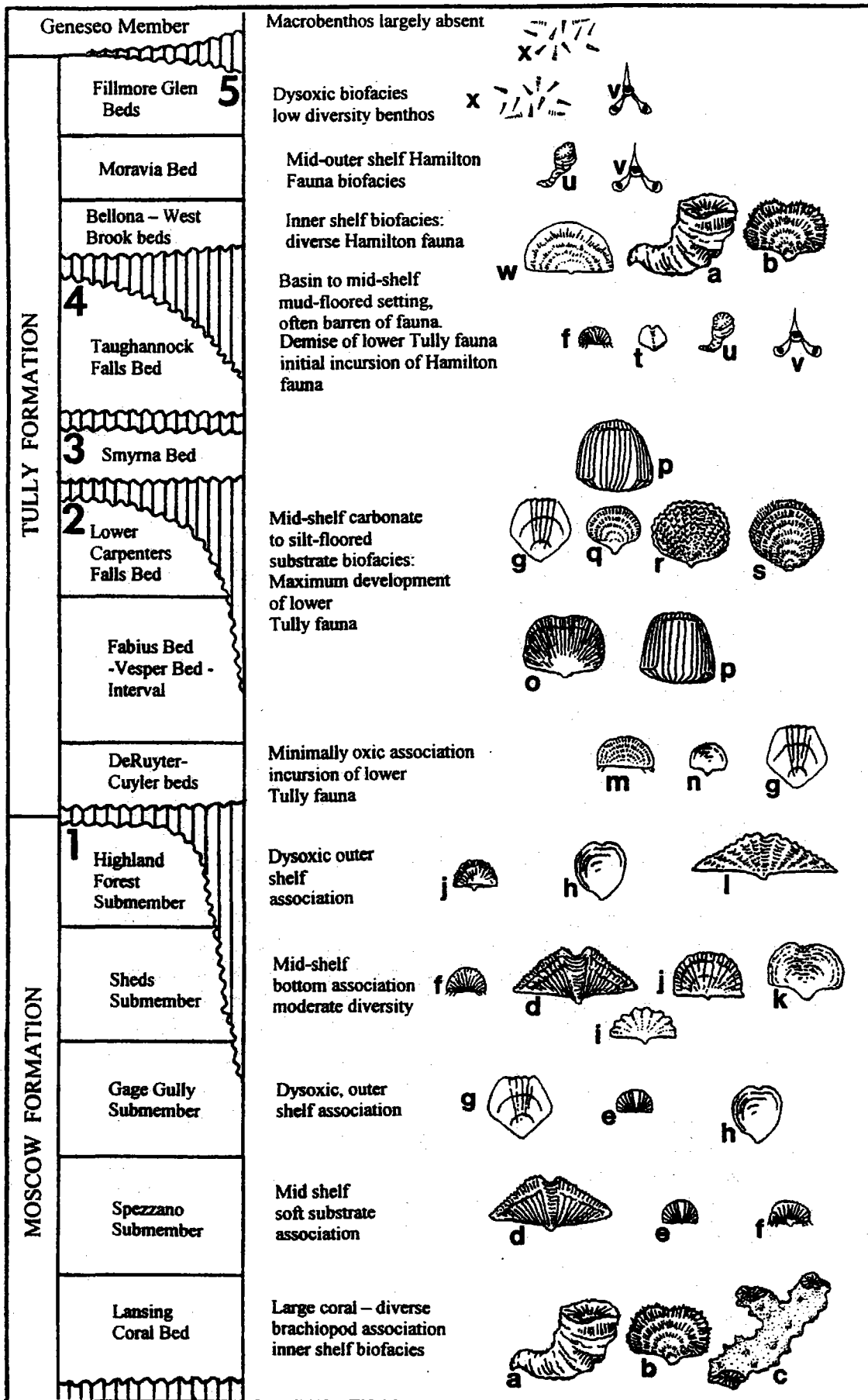


Figure 13

## ACKNOWLEDGEMENTS

We are grateful to Phillip Heckel for helpful discussion and criticism in the development of our Tully-related ideas. Carol Baird typed the bulk of this manuscript and Evelyn Pence helped with the drafting and processing the letter and line graphics. We are particularly grateful to the property owners who have kindly allowed us to visit their properties. The project was partially supported by grants from the USGS STATEMAP program and NSF EAR 9725441 (to C. E. Brett).

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**APPENDIX (LOCALITY REGISTER)**

New Lisbon Member – Tully Formation Localities in study area. (Localities discussed in text and/or shown in figures)

\* = Denotes sections listed in Heckel (1973).

Earlville 7.5' Quadrangle:

- \*1 Unnamed, northeast-flowing tributary of Pleasant Brook, 0.8 mile southeast of Upperville (complete Tully section).
- 2 Unnamed, northeast-flowing tributary of Pleasant Brook, 0.1 – 0.2 mile northeast of Coye Hill Road and 0.9 mile west of Smyrna (nearly complete Tully section).
- \*3 Unnamed south-flowing creek, 0.5 – 0.7 miles north-northeast of Sherburne Four Corners (nearly complete Tully section).
- 4 Very small, unnamed, northeast-flowing gully 0.15 mile north of Cush Hill Road (not shown as actual stream on map). Creek is 1.7 mile due west of Route 80/Route 12 Junction in Sherburne on Rogers Nature Center property (much of Tully section under thin, but removable, alluvium).

Sherburne 7.5' Quadrangle:

- 5 North-facing cut along School Road 1.3 miles northeast of Route 80/Route 12 Junction in Sherburne (Moravia beds – lower Sherburne siltstone succession).
- 6 Unnamed northwest-flowing creek 0.2 mile north of School Road and 1.5 mile northeast of Route 80/Route 12 Junction in Sherburne (base and top of Tully not exposed).
- 7 Unnamed south-southwest flowing tributary of Mad Brook, 1.1 miles east-northeast of Route 80/Route 12 Junction in Sherburne (Tully section mostly complete).
- \*8 Unnamed, southwest-flowing creek, 0.3 mile north of Pleasant Valley Road which parallels Negro Hollow. Locality is 1.6 miles southeast of Route 80/Route 12 junction in Sherburne (good Tully section; Tully- Geneseo contact not exposed).
- \*9 West Brook, 0.7 mile upstream from (east of) Route 12 overpass (complete Tully section, West Brook Shale type section). Field trip Stop 5.
- 10 Unnamed, southwest-flowing creek 0.25 mile upstream from Pleasant Valley Road which parallels Negro Hollow. Locality is 2.4 miles southeast of Route 80/Route 12 junction in Sherburne (some of Tully section under thin, removable alluvium).
- 11 Negro Hollow Creek (south fork) adjacent to, and upstream from, Negro Hollow Church (top-Tully contact with Geneseo Member and complete Geneseo section upwards into Sherburne succession).
- \*12 Unnamed, southeast-flowing tributary of Mad Brook parallel to unused road. Section is 0.4 miles north-northwest of hamlet of Harrisville (lower and medial Tully well exposed).
- 13a Very small, north-flowing gully, 0.45 miles south of Bingham Road and 1.8 mile west-northwest of Columbus and 0.5 miles north of Route 80. Section is nearly due south of the Campbell Cemetery but is not shown as a running stream on map (part of Tully accessible under alluvium).
- 13b Very small, north-flowing gully, 0.5 miles south of Bingham Road and 1.6 mile west-northwest of Columbus and 0.65 miles north of Route 80. Locality 0.25 miles east of locality 13a but is not shown as a running stream on map (lower and

medial Tully accessible under alluvium).

- 14 East-flowing south fork of Center Brook 0.25 miles west of (upstream from) Winton Road overpass. Section is 1.5 mile west-southwest of Columbus (Smyrna Bed chamosite well exposed above top-Moscow contact).
- 15 Unnamed, northeast-flowing tributary of Center Brook, 0.5 miles west of (upstream from) southward, right angle turn on Walt Phillips Road. Section is 1.5 miles southwest of Columbus (top-Tully-Geneseo succession is exposed on north [main] fork of gully; Walt Phillips Bed, West Brook Shale, and base-Moravia bone bed are exposed on smaller south fork).
- 16 Roadcut along Walt Phillips Road 0.4 miles south of the southward right angle turn on Walt Phillips Road and 1.5 miles southwest of Columbus as well as unnamed, northeast-flowing creek parallel to the road (part of lower Tully clastic correlative succession exposed on creek, part of medial Tully and highest Tully –succession exposed along road; Walt Phillips Bed type section). Field trip Stop 7.

New Berlin North 7.5' Quadrangle:

- \*17 Northeast-flowing tributary of Center Brook 0.6 miles south of Walt Phillips Road and 1.5 miles south of Columbus. Locality is 0.6 miles west of Balcom Hill Road (partial New Lisbon Member section, good Tully Formation clastic equivalent-succession. "Perrytown" section of Cooper and Williams, 1935).
- 18 Northeast-flowing tributary of Center Brook 0.3 miles southwest of junction of Balcom Hill Road and Route 80 at hamlet of Shawler Brook. Creek runs southeast of, and sub-parallel to, Balcom Hill Road (long section of Tully Formation clastic correlative succession, although top and base of succession is covered). Field Trip Stop 6.
- 19a Small, east-flowing tributary of Center Brook and adjacent abandoned quarry immediately north of gully, 0.5 miles west-northwest of Five Corners. Section is 0.2 miles north of section 19b (parts of medial Tully Formation clastic equivalent succession exposed).
- 19b Small, east-flowing tributary of Center Brook, 0.45 mile west of Five Corners. Portion of creek exposing Tully Formation clastic – equivalent beds is parallel to, and immediately north of, Warner Hill Road (part of medial Tully Formation clastic equivalent succession exposed).
- 20 West-flowing, unnamed tributary of Unadilla Creek, 1.7 miles south of South Edmeston and 1.6 miles northeast of Five Corners (chamositic base of New Lisbon Member exposed in glacially rucked section at approximately 1550 foot elevation on creek north fork).
21. North-flowing banks of Mill Creek, 0.5 – 0.7 miles west of Route 8 – 80/Route 80 intersection in New Berlin (hanging sections of probable medial Tully Formation [Taughannock Falls beds – equivalent?] clastic equivalent succession).
- 22 Small, north-facing cut along Terrace Heights Road, 200 feet south of Mill Creek and 0.45 miles west of Route 8 – 80/Route 80 intersection in New Berlin near south edge of quadrangle (uppermost Taughannock Falls beds - equivalent strata and basal part of West Brook Shale).

New Berlin South 7.5' Quadrangle:

- 23 Small, east-flowing tributary of Unadilla Creek, 0.05 miles south of Angell Road

medial Tully accessible under alluvium).

- 14 East-flowing south fork of Center Brook 0.25 miles west of (upstream from) Winton Road overpass. Section is 1.5 mile west-southwest of Columbus (Smyrna Bed chamosite well exposed above top-Moscow contact).
- 15 Unnamed, northeast-flowing tributary of Center Brook, 0.5 miles west of (upstream from) southward, right angle turn on Walt Phillips Road. Section is 1.5 miles southwest of Columbus (top-Tully-Geneseo succession is exposed on north [main] fork of gully; Walt Phillips Bed, West Brook Shale, and base-Moravia bone bed are exposed on smaller south fork).
- 16 Roadcut along Walt Phillips Road 0.4 miles south of the southward right angle turn on Walt Phillips Road and 1.5 miles southwest of Columbus as well as unnamed, northeast-flowing creek parallel to the road (part of lower Tully clastic correlative succession exposed on creek, part of medial Tully and highest Tully –succession exposed along road; Walt Phillips Bed type section). Field trip Stop 7.

New Berlin North 7.5' Quadrangle:

- \*17 Northeast-flowing tributary of Center Brook 0.6 miles south of Walt Phillips Road and 1.5 miles south of Columbus. Locality is 0.6 miles west of Balcom Hill Road (partial New Lisbon Member section, good Tully Formation clastic equivalent-succession. "Perrytown" section of Cooper and Williams, 1935).
- 18 Northeast-flowing tributary of Center Brook 0.3 miles southwest of junction of Balcom Hill Road and Route 80 at hamlet of Shawler Brook. Creek runs southeast of, and sub-parallel to, Balcom Hill Road (long section of Tully Formation clastic correlative succession, although top and base of succession is covered). Field Trip Stop 6.
- 19a Small, east-flowing tributary of Center Brook and adjacent abandoned quarry immediately north of gully, 0.5 miles west-northwest of Five Corners. Section is 0.2 miles north of section 19b (parts of medial Tully Formation clastic equivalent succession exposed).
- 19b Small, east-flowing tributary of Center Brook, 0.45 mile west of Five Corners. Portion of creek exposing Tully Formation clastic – equivalent beds is parallel to, and immediately north of, Warner Hill Road (part of medial Tully Formation clastic equivalent succession exposed).
- 20 West-flowing, unnamed tributary of Unadilla Creek, 1.7 miles south of South Edmeston and 1.6 miles northeast of Five Corners (chamositic base of New Lisbon Member exposed in glacially rucked section at approximately 1550 foot elevation on creek north fork).
21. North-flowing banks of Mill Creek, 0.5 – 0.7 miles west of Route 8 – 80/Route 80 intersection in New Berlin (hanging sections of probable medial Tully Formation [Taughannock Falls beds – equivalent?] clastic equivalent succession).
- 22 Small, north-facing cut along Terrace Heights Road, 200 feet south of Mill Creek and 0.45 miles west of Route 8 – 80/Route 80 intersection in New Berlin near south edge of quadrangle (uppermost Taughannock Falls beds - equivalent strata and basal part of West Brook Shale).

New Berlin South 7.5' Quadrangle:

- 23 Small, east-flowing tributary of Unadilla Creek, 0.05 miles south of Angell Road



- and 1.0 miles southwest of Route 8 – 80/Route 80 junction in New Berlin (uppermost Taughannock Falls beds equivalent strata, West Brook Shale, and part of Moravia beds equivalent succession).
- \*24 Large, steep, east-flowing tributary of Unadilla Creek, 1.5 miles south-southwest of Route 8 – 80/Route 80 Junction in New Berlin (complete Tully Formation clastic equivalent succession. New Lisbon Member feather edge, and Smyrna Bed uncovered through excavation, this is probable “Greens Gulf” section of Cooper and Williams, 1935).
  - 25 Small, west-flowing tributary of Unadilla Creek, 0.1 – 0.2 miles upstream from junction of Route 18 and Shacktown Road. Gully parallels Shacktown Road on south side (parts of medial Tully Formation clastic equivalent succession exposed).
  - 26 West-northwest flowing tributary of Unadilla Creek, 2.1 miles east-southeast of Route 8 – 80/Route 80 junction in New Berlin. Waterfall and bank section 0.15 – 0.2 miles north of closest approach of Route 13 (excellent exposure of upper part of Taughannock Falls beds-equivalent succession, West Brook Shale, Moravia-Fillmore Glen beds? – equivalent strata, Geneseo Member?, and Sherburne Member).
  - 27a Unnamed, north-flowing tributary of Wharton Creek, 0.3 – 0.6 miles south of Pittsfield. Section is west of Ouleout Road and straddles New Berlin North and New Berlin South Quadrangle boundary (long section exposing upper Moscow-through-upper part of Taughannock Falls beds equivalent succession. Base of New Lisbon Member, base of Tully Formation clastic correlative succession, and probable Symrna Bed exposed).
  - 27b West flowing east fork of Locality 27a creek, 0.15 miles east of (upstream from) Ouleout Road overpass and 1.3 miles southeast of Pittsfield (upper Taughannock Falls beds equivalent strata, West Brook Shale, and part of Moravia beds equivalent succession exposed in waterfall).

**New Berlin North 7.5' Quadrangle:**

- \*28 Small tributary of larger unnamed tributary of Wharton Creek, 1.5 miles east of Pittsfield. Section is 0.3 – 0.1 mile north of (downstream from) Ramey Road overpass. Creek is not shown as running stream on map (most of New Lisbon Member exposed, base-Tully contact and much of succeeding Tully Formation clastic correlative succession is exposed, this is probable “Pittsfield” section of Cooper and Williams, 1935).

**Edmeston 7.5 Quadrangle:**

- 29 Small, unnamed, northwest flowing gully, 0.3 miles northwest of Gross Hill School and 0.9 miles north, northwest of Crystal Lake. (Nearly complete section of New Lisbon Member with base-and-top-contacts exposed. Base-Tully contact and portions of the succeeding Tully Formation clastic correlative succession exposed).

**Morris 7.5' Quadrangle:**

- 30 Unnamed, southeast flowing tributary of Butternut Creek, 1.25 miles southwest of intersection in New Lisbon. Section is 0.2 – 0.4 miles northwest of (upstream from) the Route 51 overpass (upper part of Tully Formation clastic correlative succession, including West Brook Shale, exposed).
- 31 Unnamed, south flowing west tributary of Stony Creek, 0.9 – 1.4 miles north, northeast of New Lisbon and 0.15 – 0.25 miles west of Parker Road (most of New

- Lisbon Member, including basal chamositic contact with Cooperstown Member, exposed).
- 32 East facing banks of Stony Creek both downstream and upstream from Miller Road overpass. Section is 1.3 – 1.6 miles northeast of intersection in New Lisbon (hanging section of New Lisbon Member without base-or top-contacts, this is probably part of Cooper and Williams, 1935, type New Lisbon Member section).
- 33 Southeast flowing, unnamed tributary of Stony Creek, 1.0 miles southwest of Welcome and 0.15 miles southwest of terminus of Hudson Road (medial Tully Formation clastic equivalent exposed).
- \*34 Unnamed, west flowing tributary of Butternut Creek, 1.0 – 1.2 miles east, northeast of intersection in New Lisbon. Section extends for 0.4 miles upstream from the New Lisbon-Hartwick Road overpass and includes an abandoned quarry and cuts along Otto Stahl Road which parallels the creek (upper part of New Lisbon Member type section exposed as is the chamositic New Lisbon-base-Tully contact near the lower end of this creek, Tully Formation clastic correlative succession almost completely exposed, West Brook Shale and highest Tully exposed along upstream south fork of creek, “New Lisbon” section of Cooper and Williams, 1935). Field trip Stop 8.
- 35 Unnamed, northwest flowing tributary of Butternut Creek, 1.25 miles south of intersection in New Lisbon. Creek is immediately north of, and parallel to, Gulf Road (strata probably equivalent to topmost Tully Formation clastic correlatives, exposed at lowest end of section 0.15 miles southeast of Gulf Road/Pegg Road intersection).

Mt. Vision 7.5' Quadrangle:

- 36 East facing cut-bank of Pool Brook 2.4 miles north, northwest of Laurens, Section is west of Poolbrook Road and 1.85 miles northwest of Poolbrook Road/West Valley Road intersection (top of Cooperstown Member-into-basal Tully Formation succession).
- \*37 Unnamed, northwest flowing tributary of Otego Creek, 1.2 miles northeast of center of Laurens. Section commences 0.2 miles upstream from Route 205 overpass (fossiliferous upper part of lower Tully [“Laurens Member”] and most medial Tully Formation clastic correlative succession exposed; excellent, easternmost exposure of West Brook Shale; this is “Laurens” section of Cooper and Williams, 1935).
- \*38 Unnamed, west flowing tributary of Otego Creek, 2.0 miles northeast of center of Laurens. Section extends 0.1 – 0.35 miles upstream from Route 205 overpass (lower Tully clastic correlative succession exposed; Moscow Formation-base-Tully contact exposed; this is “Houghtelings Glen” section of Cooper and Williams, 1935).

## ROAD LOG

Leave SUNY Oneonta staging area and proceed to I-88 (westbound). We will proceed from Oneonta to Tully, NY via Binghamton on Interstates 88 and 81 giving us an initial drive of 111 miles. We will stop at a rest area on I-81 (McGraw area or Preble area depending on passenger needs). This part of the trip will involve a driving time of approximately 2 hours and 10 minutes. However, it allows us to proceed from the west towards Oneonta so that the Tully Limestone-to-Tully clastic correlative transition can be viewed fully. The road log starts from the I-81 northbound exit ramp junction with Route 80 west of Tully and ends at the Route 28/I-88 junction northeast of Oneonta. The Sunday trip agenda starts at the east end of the Saturday road log venue and proceeds westward. Additional Sunday stops west of the Tully Valley are included (along with brief routing information) in an appended section following the detailed road log section.

| Accumulated Miles | Incremental Miles | Saturday Road Log  |
|-------------------|-------------------|--|
| 0.0               | 0.0               | Junction of northbound I-81 exit ramp with State Route 11. Turn left (north) on Route 11.                        |
| 4.0               | 4.0               | Junction of Route 11 with State Route 80 in Tully. Turn left (west) onto Route 80.                               |
| 4.25              | 0.25              | Leave Tully. Continue west on Route 80.  |
| 4.45              | 0.2               | Turn right (north) into entrance for Best Western Motel and Tully Lakes Restaurant.                              |
| 4.5               | 0.05              | Proceed by car to the rear of the restaurant-motel complex and park by large shale borrow pit and exit vehicles. |

Stop 1: Tully Shale Borrow Pit at Tully Center (20 minutes). Windom Fauna and Basal Tully Contact.

This well known stop for professionals and amateurs alike is best known for diverse taxa of the Hamilton Fauna that occur in strata of the Windom Member. Windom divisions exposed here include; in ascending order: The Taunton submember (25 feet); the Lansing Coral Bed (1.5 feet); the Spezzano submember (5 feet); the Gage Gully submember (8 feet); and the Sheds submember (6 feet). The Taunton submember is a shoaling (RST) interval. The Lansing probably marks the base of a subsequence interval. All of it (or part of it) commences a transgressive (TST) succession that continues through the Spezzano submember and into dysoxic early highstand facies of the lower Gage Gully submember. The Sheds marks the return of shoaling condition followed by another transgression recorded by the newly defined Highland Forest submember (here absent due to erosion) that is represented by dysoxic *Emanuella*-bearing strata (Baird and Brett, 2003). Diverse Hamilton taxa (too numerous to discuss here) are characteristic of the shallower facies. Dysoxic taxa, represented by *Eumetabolatoechia* ("*Leiorhynchus*") *multicostatum*, *Emanuella praeumbona*, and diminutive *Allanella tullia* are characteristic of the dark gray Gage Gully submember. Not discussed in most earlier reports is the occurrence of the basal layer (DeRuyter Bed) of the Tully Formation at the top of this section. This bed is represented by a 0.3 foot thick layer of impure limestone abounding in the characteristic lower Tully strophomenid *Rhyssochonetes aurora* and scattered phosphatic pebbles. This layer rests disconformably on the Sheds submember; at Junes' Ravine, 1.3 miles east-northeast of this section, the topmost division of the Windom

Member (here eroded) is represented by the Highland Forest submember. At Junes' Ravine it is represented by 0.9 feet of silty, calcareous shale rich in *Emanuella*, small *Tropidoleptus*, small *Allanella*, and *Mucrospirifer*.

- |      |      |   |
|------|------|---|
| 4.55 | 0.05 | Return to motel/restaurant entrance and turn right (west) onto Route 80.  |
| 4.7  | 0.15 | Junction of Route 80 and Route 11/Route 281 at red light. Continue straight (west) on Route 80.   |
| 4.75 | 0.05 | Pass under Interstate 81. Continue straight (west) on Route 80.   |
| 4.95 | 0.2  | View of Tully Lakes (Kettle Lakes) to left (south). Song Mountain Ski complex in distance. For the next 1.8 miles we will cross the Tully Valley on the Valley Heads Moraine. Notice the elevated drift-floored plain to the south and the approaching bedrock valley wall to the west. |
| 5.75 | 1.8  | Long Road/Route 80 intersection. Notice sign for Song Mountain Ski area. This is a sharp hairpin turn. Turn left (south) with caution onto Long Road.   |
| 6.3  | 0.55 | Private driveway on right. However, we must turn left into "jug handle" turnaround in order to ascend driveway lane.  |
| 6.4  | 0.1  | Park by house at top of driveway incline. We will proceed on foot to outcrop along driveway.  |

Stop 2: Tully Limestone Divisions in Driveway Section Next to Carrs' Creek – Quarry Locality.

The Carrs' Creek-Quarry section serves to illustrate the typical, prominent carbonate succession of the Tully Formation in west-central New York. Fortuitous excavation of the new driveway and the kindness of the owners who built it, preclude our need to climb the slippery creek section. Moreover, the driveway cut reveals weathered joint surfaces that show excellent textural details and fossils. The driveway section exposes in upward-succession: the Fabius Bed (3 feet visible), the Meeker Hill Bed (3.75 feet), the Tully Valley Bed (1.5 feet), the Vesper Bed (1.5 Feet), the Carpenters Falls Bed lower part (3.4 feet); the upper part of the Carpenters Falls Bed separated from the lower part by a discontinuity (0.5 feet); the Taughannock Falls Bed (2.7 feet); the Bellona Coral Bed (0.3 feet); and the Moravia Bed (2.75 feet visible). Basal Tully units visible in the creek but not along the driveway include: the DeRuyter Bed, the Cuyler Bed, and the lower part of the Fabius Bed.

The present authors believe that the lower part of the Tully Formation succession (DeRuyter Bed-to-the-intra Carpenters Falls discontinuity) records an interval of overall shoaling with increased Tully Fauna diversity upward (see text; Baird and Brett, 2003). The topmost Carpenters Falls Bed and the Taughannock Falls Bed (medial Tully succession) records a time of relative deepening following the intra-Carpenters Falls lowstand event. This is roughly timed with the demise of the Tully fauna and redevelopment of the Hamilton Fauna (see text; Baird and Brett, 2003). A second regressive episode is recorded by the Bellona Bed. The Bellona "coral plantation" (*sensu* Heckel, 1973), yielding large rugosans and tabulates such as *Favosites* and *Alveolites* at this locality and others further west, records maximum return of the diverse Hamilton Fauna Biota. Actually, the basal 0.5 feet of the overlying Moravia Bed is composed of

dark limestone yielding corals and *Spinatrypa*; we believe that this interval correlates eastward to a limestone bed that caps the West Brook Shale (Figures 5, 6).

The five-six inch thick topmost part of the Carpenters Falls Bed is characterized by "oolitic" grains somewhat resembling chamositic "ooids" in the Smyrna Bed to the east. Moreover, the top of this interval displays fracture systems resembling the "mudcracks" reported by Heckel (1973). Most significant is the presence of laminated carbonate texture resembling algal (microbialite) stromatolites (see text discussion). We believe that these structures underlie Heckel's (1973) regional "intra-Tully unconformity" which we interpret as a flooding surface. We also argue that the topmost part of the Carpenters Falls Bed, as presently defined, overlies a lowstand disconformity and may be the western equivalent of the Smyrna Bed (see discussion).

|       |      |  |
|-------|------|--|
| 7.0   | 0.6  | Return to Long Road/Route 80 junction (REMEMBER to use available "jug handle" at base of driveway in order to turn left (north). The right turn onto Route 80 is too sharp to negotiate; we will turn left instead and proceed to one or more turnaround places where we can go eastbound on Route 80. We will then collect together on the shoulder opposite the entrance to the Best Western Motel-restaurant complex. |
| 7.6   | 0.6  | Long Road/Route 80 intersection again. Continue straight (east on Route 80).   |
| 9.8   | 2.2  | Temporary pull-off spot to collect vehicles opposite Best Western complex. Continue east on Route 80.  |
| 10.25 | 0.45 | Junction of Route 80 with Route 11 at intersection in Tully. Continue straight (east) on Route 80.   |
| 10.6  | 0.35 | East edge of Tully   |
| 11.25 | 0.65 | Type Tully Formation section (Junes' Ravine) to the north (left) in the uphill woods.  |
| 12.35 | 1.1  | Excellent view of glacial U-shaped profile of Labrador Valley to right (south).  |
| 12.75 | 0.4  | Middle of hamlet of Apulia. Continue straight (east) on Route 80.  |
| 13.55 | 0.8  | Middle of hamlet of Apulia Center. Continue straight (east) on Route 80.   |
| 19.2  | 5.65 | Enter town of Fabius. Continue straight (east) on Route 80.  |
| 22.7  | 2.80 | Entrance to Highland Forest Park (Onondaga County – Syracuse Metropark System). Turn right (south) and proceed uphill into park.   |
| 23.75 | 1.05 | Park headquarters and Visitor Center in Highland Forest Park. Turn right into parking area.  |
| 23.80 | 0.05 | Park vehicles at southwest of parking area and proceed on trail to footpath bridge over creek near parking area.   |

Stop 3: Tully Formation sections adjacent to Highland Forest Metropark Headquarters.

The fortunate location of this section allows for the use of lavatory facilities and time for box lunch consumption. Two Tully sections can be studied, time permitting. The first, below a trail bridge near the parking lot and bordered by paths, will be the main object of

attention. This exposes all of the Tully Formation except a portion of the Fillmore Glen Bed-interval near the top of the Tully. In this region, a prominent regional disconformity, first characterized in detail by Heckel (1973) is observed to divide the Tully succession. The lower Tully succession (DeRuyter Bed-into-Carpenters Falls Bed interval) is observed to be overstepped rapidly eastward from the meridian of Highland Forest Park such that no lower Tully is observed in sections east of the meridian of Sheds. This erosion was believed to commence eastward from the level of the post-Carpenters Falls "intraformational unconformity" southeast of Fabius (Heckel, 1973; Figures 2, 3). From DeRuyter Reservoir eastward, this disconformity surface was believed to coincide with the base of the Smyrna Bed (Heckel, 1973) that is traceable to the Sherburne area. The present authors connect the regional, sub-Smyrna disconformity westward, not to the post-Carpenters Falls contact, as did Heckel (1973), but to the base of what Heckel termed "Carpenters Falls Bed" in the Fabius-Labrador Valley area (Baird and Brett, 2003). The thin "Carpenters Falls Bed" interval of the Fabius-Labrador Valley region is a sandy, abraded grain encrinite characterized by a sharp irregular base marked by giant hypichnial burrow prods and minor phosphorite. Careful examination of sections between Fabius and DeRuyter shows that this unit is present in most sections across the region and that it connects laterally to the Smyrna Bed. At this locality, this bed is present, but is of variable thickness; under the walk bridge it is only a few inches thick, but, at the adjacent south fork section, it is almost a foot-thick. This may reflect local Smyrna Bed cutout beneath the sub-Taughannock Falls Bed flooding surface contact. The intra-Carpenters Falls discontinuity at Stop 2 may connect to the sub-Smyrna disconformity.

Another major aspect of the lower Tully succession in this region is that the Fabius Limestone Bed at Stop 2 has transformed into a calcareous siltstone-silty limestone unit and the underlying Cuyler Bed has thickened considerably. In contrast, much of the upper Tully succession consists of clean limestone and clay shale. This is a major facies change described in detail by Heckel (1973). He related it, in part, to terrigenous sediment-sourcing from the region of lower Tully erosional cutout between DeRuyter and Sherburne (Heckel, 1973).

- |      |     |   |
|------|-----|---|
| 24.9 | 1.1 | Return to park entrance road/Route 80 Junction. Turn right (east) onto Route 80.                        |
| 25.2 | 0.3 | Route 80/DeRuyter Road-Oran-Delphi Road intersection. Continue straight (east) on Route 80.             |
| 25.7 | 0.5 | Cross Limestone Creek. Morainal deposits exposed in creek banks.  |
| 26.2 | 0.5 | East valley rim of Limestone Creek. Valley head moraine plain for next 0.4 miles. Enter Madison County. |
| 28.1 | 1.9 | Enter New Woodstock. Continue straight (east) on Route 80.  |
| 28.7 | 0.6 | Leave New Woodstock. Continue straight (southeast) on Route 80.   |
| 31.6 | 2.9 | Route 80/Route 13 crossroad in hamlet of Sheds. Continue straight (east-southeast) on Route 80.         |
| 32.6 | 1.0 | Junction of Route 80 with Carpenter Road. Turn right (south) onto Carpenter Road.                       |
| 33.5 | 0.9 | Tully Formation roadcut to left. Pull off on shoulder and proceed                                       |

to outcrop.

**Stop 4: Tully Formation Roadcut and Road Ditch Section Showing Complete Overstep of Lower Tully Member Division by Sub-Smyrna Bed Disconformity.**

This section displays the two uppermost submember divisions of the Windom Member overlain by units of the upper Tully Formation succession. The upper part of the Sheds submember is visible in the lowest 2 – 3 feet of the bank and yields numerous Hamilton Fauna brachiopods and clams. This is abruptly overlain by 2.5 feet of Highland Forest submember that is characterized by siltstone beds and a low diversity association of *Emanuella praeumbona*, small *Tropidoleptus* and *Mucrospirifer*. This section exposes the highest strata of the Windom below the Tully in central New York, although considerably thicker Highland Forest sections are observed in central Pennsylvania (Baird and Brett, 2003).

The visible Tully succession rests disconformably on the Windom. At this locality the basal 16 inches of the Tully is represented by the Smyrna Bed, a hard overhanging ledge composed of abraded calcarenite and minor “oolitic” chamosite. This bed displays a sharp base with development of large hypichnial burrow prods filled with encrinite, minor sand, minor phosphate, and chamosite. Lucky persons will find clusters of 3-D *Tullypothyridina* in some of the basal protruding burrows. As noted by Heckel, 1973, the lower Tully succession is absent due to regional overstep by the sub-Smyrna disconformity. This overstep is characteristic of Tully sections eastward to the Sherburne area. Post-Smyrna strata are represented by an expanded shale-limestone phase of the Taughannock Falls Bed interval. In the road ditch, strata of the West Brook Shale and its capping limestone unit can be seen. These units yield many Hamilton Fauna taxa. Excellent trilobite fossils, camerate crinoids and blastoids are found in this interval.

|      |     |  |
|------|-----|--|
| 34.4 | 0.9 | Return to Carpenter Road/Route 80 Junction. Turn right (east) onto Route 80.   |
| 39.5 | 5.1 | Intersection of Route 26 and Route 80 in Georgetown. Turn right (south) onto Route 26/80.  |
| 39.7 | 0.2 | Pass Spirit House on left. Special ornamentation and rounded corners were devices designed by owner to fool the Devil.   |
| 43.1 | 3.4 | Routes 80 and 26 split. Turn right (east) on Route 80.   |
| 43.2 | 0.1 | Cross Otselic Creek and enter village of Otselic.  |
| 43.5 | 0.3 | Leave Otselic. Continue straight (east) on Route 80.   |
| 49.2 | 5.7 | Pass through Upperville. Continue straight (east) on Route 80.   |
| 49.6 | 0.4 | Notice major exposure of medial Windom Member/Cooperstown Member strata on floor of Pleasant Brook to the right.   |
| 50.1 | 0.5 | More visible Moscow Formation strata on floor of Pleasant Brook. Major Tully Formation exposure (Upperville section: Locality 1) is across brook on along a tributary. |
| 51.5 | 1.4 | Enter Village of Smyrna. Continue straight (east) on Route 80.   |
| 52.2 | 0.7 | Leave Village of Smyrna. Continue straight (east) on Route 80.   |
| 55.0 | 2.8 | Pass Rogers Nature Center. Geese, giant carp, and snapping turtles can be viewed from a bridge near the parking lot.   |
| 55.4 | 0.4 | Cross Chenango Creek.  |
| 56.0 | 0.6 | Intersections of Routes 12 and 80 in Sherburne. Turn left (south)  |

|      |     |   |
|------|-----|---|
|      |     | onto Route 12.  |
| 56.5 | 0.5 | Leave Sherburne. Continue straight (south) on Route 12.   |
| 59.3 | 2.8 | Turn left (east) onto Park Road from Route 12.  |
| 59.8 | 0.5 | Car pull-off to left near Stop 5 ( We will pass it and turn around in a driveway so as to be facing west when stopped), |
| 60.1 | 0.3 | Pull off on shoulder of Park Road facing west. Exit vehicles and proceed to stream cut on West Brook.                   |

Stop 5: Thin, condensed Tully Formation Section on "Sherburne High" Structural uplift.

This is an important section (Locality 9) described in Cooper and Williams (1935) and Heckel (1973) publications. This is the type section of the West Brook Shale, a nearly three foot-thick unit of calcareous shale abounding in diverse Hamilton Fauna taxa recorded in extensive faunal lists (Cooper and Williams, 1935, Heckel, 1973). It is the "last hurrah" for the Hamilton Fauna prior to the gradual onset of transgressive anoxia associated with Genesee black shale deposition.

Eastward erosive downcutting of Hamilton Group strata by the sub-Smyrna Bed disconformity places the Cooperstown Member-Tully Formation contact on the Spezzano submember interval of the Cooperstown (Figures 5, 6). The Smyrna Bed at this locality is a 4 - 5 inch-thick unit of "oolitic" chamosite and subsidiary sandy, bioclastic abraded calcarenite that rests disconformably on underlying Hamilton Group deposits (Figures 5, 6). Discoidal dark gray to black, sand-sized chamosite grains dominate over subsidiary phosphorite pebbles and rare shells within the bed. Though not encountered by the present authors, Heckel (1973) reports a cluster of *Tullypothyridina* within the Smyrna Bed. Above the Smyrna Bed is a 2.2 foot interval of hard, slabby siltstone or fine sandstone facies yielding few fossils. Heckel (1973) referred to this unit as the "unnamed sandstone." Above this unit is a complex interval of "oolitic" chamosite, tabular siderite and limestone comprising the interval between the "unnamed sandstone" and the basal contact of the West Brook Shale (Figures 5, 6). Heckel (1973) referred to this interval of the "Taughannock Falls Oolite Bed" and indicated that it was a condensed lateral equivalent of the Taughannock Falls Bed (Heckel, 1973).

Careful work by Baird and Brett (this paper) shows the probability that Heckel's (1973) correlation is correct. However, the disjunct character of section match between localities 3 and 4 (Figure 5A) and localities 12 and 13a, b (Figure 6) offer the possibility that the Smyrna Bed near Smyrna (Localities 1, 2) and Sherburne Four Corners (Locality 3) may actually connect to the "Taughannock Falls Oolite Bed" rather than the lower chamositic layer (Figure 5A). Thus, from Cush Hill Road (Locality 4) to Harrisville (Locality 12), it is possible that a part of the lower Tully Member (Fabius Bed?) may be represented by the lower chamositic layer and the "unnamed sandstone" (Figure 5, 6). Both chamositic beds are well developed at this locality. Chamosite, an iron-rich silicate, of somewhat problematic origin is characteristic of several Tully (and even pre-Tully) levels in east-central New York. The absence of chamosite in western New York and east-central New York sections shows the occurrence of this mineral to be restricted to a particular depositional belt (see text discussion). The absence of common robust fossils in chamositic beds, the mud-rich character of some chamosite-rich layers, and the occurrence of chamosite in dysoxic facies yielding "leiorhynchid"-type rhynchonellid



brachiopods suggest that the origin of this unusual deposit may be some physical (or diagenetic) process occurring across a range of outer shelf and dysoxic, sediment-starved environments (see discussion).

- |       |      |   |
|-------|------|---|
| 63.9  | 3.8  | Backtrack to intersection of Routes 80 and 12 in Sherburne. Turn right (east) onto Route 80.  |
| 64.4  | 0.5  | Leave Sherburne. Continue straight (east) on Route 80.  |
| 65.5  | 1.1  | Cross Mad Brook. Upper part of Cooperstown Member exposed below bridge.   |
| 65.6  | 0.1  | Intersection in hamlet of Harrisville. Continue straight (east) on Route 80.  |
| 70.6  | 5.0  | Southward turn of Route 80 in hamlet of Columbus. Turn right and continue south on Route 80.  |
| 71.7  | 1.1  | Junction of Route 80 and Walt Phillips Road. Turn right (west) onto Walt Phillips Road.   |
| 71.8  | 0.1  | Fairkit Farm to left. Major "Perrytown" section (Locality 17) of Tully clastic correlative succession (see Cooper and Williams, 1935; Heckel, 1973) high up on hill to the south. |
| 72.7  | 0.9  | Southward turn on Walt Phillips. Bear left.   |
| 73.0  | 0.3  | Roadcut along Walt Phillips Road. This will be a later stop.  |
| 73.7  | 0.7  | Junction of Walt Phillips Road with New Turnpike Road. Turn left (southeast) onto New Turnpike Road.  |
| 74.9  | 1.2  | Junction of New Turnpike Road with Balcom Hill Road. Turn left (northeast and east) onto Balcom Hill Road.  |
| 75.5  | 0.6  | Continue straight into dead end driveway where Balcom Hill Road bends left down a steep hill. Driveway splits off just before Balcom Hill Road descends over brow of hill.        |
| 75.55 | 0.05 | Park cars near house or adjacent available spaces. Proceed across field and down side of unnamed gully (Locality 18) below the end of the field.                                  |

**Stop 6: Thick Tully Clastic Correlative Succession Along Unnamed Creek (Locality 18) parallel to, and east of Balcom Hill Road.**

Cooper and Williams (1935) and Heckel (1973) devoted considerable discussion to the first thick clastic Tully outcrop ("Perrytown" section) that was encountered proceeding east from the area of thin Tully occurrence on the "Sherburne High". That outcrop (Locality 17) is too far from any road to be visited on this trip. However, another adjacent section (Locality 18; Figure 7) that is almost as complete is herein included as a stop. We will intersect this creek at the topmost exposed Cooperstown Member above a small falls. The upper coral layer of the Lansing Bed is 15 feet below the falls lip and the strata comprising the falls is probably equivalent to the Spezzano submember in the Windom (Figures 7, 11). A six foot covered interval follows the Cooperstown succession that is, in turn, succeeded by 5 feet of lower Tully strata and about 45 feet of continuous upper Tully section. At the adjacent "Perrytown" section (Locality 17) part of the New Lisbon Member is exposed, as is the base of the Tully clastic succession above the New Lisbon (see text; Figure 7). The basal Tully contact on that creek is characterized by

chamosite, nodular diagenetic siderite, and Tully Fauna taxa (*Hypothyridina*, *Schizophoria*, *Spinatrypa*), probably corresponds to what we call the "base-Laurens" disconformity (see text discussion). Heckel, 1973, (Figure 3) tentatively correlated this bed to (or near) the position of the Smyrna Bed in the Sherburne area.

However, on that same creek, and on this section as well, there is a second, more complex and prominent bed characterized by calcareous, auloporid-rich sandstone in the lower part and nodular chamositic limestone and siderite in the upper part (Figures 7, 10). The present authors correlate the upper part (and tentatively the lower part) of this unit to the Smyrna Bed instead (Figures 7, 11). The coarse, hashy mix of fossil debris and the nodularity of the carbonate component are texturally similar to the Smyrna Bed though the unit is sandier overall.

Although the "base-Laurens" disconformity is covered here and the underlying New Lisbon is concealed (or absent due to overstep: see text discussion and Figures 7, 11), we will see higher divisions of the lower Tully clastic succession. A massive two foot-thick siltstone bed exposed above the covered interval may correspond to part of the Fabius Bed. Above this, participants can view the 8-inch-thick nodular, calcareous auloporid-pelmatozoan-rich sandstone ledge of the lower Smyrna Bed. Calcareous chamosite and diagenetic siderite are characteristic of the upper Smyrna (Figures 7, 10). This is a condensed, presumably transgressive, phase probably corresponding to a major mid-Tully deepening event. In this interpretation the lowstand (sub-Smyrna) erosional contact is understood to be at the base of the chamositic bed or at the base of the 8-inch auloporid-rich sandstone (Figures 7, 10).

Above the Smyrna Bed is a 37 foot-thick interval of flaggy to slabby thin siltstone beds with shale interbeds. With exception of a thin medial layer, the Walt Phillips Bed, this interval yields very few macrofossils. Auloporid corals, sparse pelmatozoan debris, plant fragments and the occasional mollusk comprise the observed biota. We will view this interval at greater advantage at Stop 7. The present authors correlate this thick "barren" unit to the Taughannock Falls Bed in the Tully Limestone, although the upper part of it may be missing due to regional overstep to the west of this area (Figure 11).

One may properly question correlations based mainly on lithologic and textural parameters and minimally on distinctive guide fossils. We acknowledge this problem with numerous caveats, but have to point out that lower and medial Tully deposits in this area are conspicuously poor in fossil content. Although Tully Fauna taxa occur both in the lower and medial Tully, they are uncommon to rare. Even in the Sherburne area, the chamositic beds contain few shells and these are mostly small. The only unifying taxon that characterizes this facies is the auloporid coral that abounds in the condensed units. In fact much of the Tully clastic correlative succession in this area resembles portions of the Penn Yan and Sherburne members of the Genesee Formation. It is, thus, not surprising that Thayer (1974) described the biota of the dysoxic prodelta slope facies of the Genesee Group as the "*Cladochonus*" (auloporid) biofacies. As such, we interpret this region to be a structural subsiding basin relative to a platform setting for the Tully Limestone further west.

Another line of evidence for subsidence of this area is the dramatic thickening of Tully units and the reappearance of lower Tully beds east of the "Sherburne High" (Heckel, 1973). Heckel (1973) correctly envisioned a "down-to-the-east" fault and a clastic trap within the basin that prevented clastics from overspreading the Tully platform. We

confirm this model (Figures 5 - 8) but show that lower Tully beds are, again, overstepped to the southeast of this section before reemerging to the east of New Berlin (Figures 6, 8). At the upper end of this transect, we will see the West Brook Bed and a capping discontinuity probably corresponding to the base of the Moravia Bed. The West Brook is richly fossiliferous and astonishingly thin relative to surrounding units. Corals, diverse brachiopods, bryozoans and other stenotopic Hamilton Fauna taxa can be found here. The West Brook has an erosional base marked by burrow prods and encrinite; this contact marks a significant lowstand event and it is responsive for conspicuous overstep of the medial Tully section west of this locality (Figures 6, 7).

- 77.4      1.85      Retrace to junction of New Turnpike Road and Walt Phillips Road. Turn right (northeast) onto Walt Phillips Road.
- 78.1      0.7      Roadcut to left. Pull off on shoulder above adjacent creek and proceed to the roadcut section.

#### Stop 7: Walt Phillips Bed and Associated Medial Tully Strata on Roadcut Along Walt Phillips Road (Locality 16).

This short stop serves to show the sparsely fossiliferous medial Tully clastic interval in good light. In particular, this outcrop is the type section of the Walt Phillips Bed a 16 inch-thick double-ledge interval of bioturbated siltstone that can be regionally correlated across the Columbus area (Figures 6, 7). At this section, this unit is at least 10 feet below the sub-West Brook disconformity. At locality 15, 0.5 miles northwest of this section, the Walt Phillips Bed is 18 inches below that contact and at locality 17, 0.9 miles east of this section, it is 17 feet below the West Brook. This pattern shows that there is a major trend of westward erosional overstep of the medial (Taughannock Falls Bed equivalent?) Tully succession below the sub-West Brook disconformity (Figures 6, 7, 11).

The Walt Phillips Bed is notable for the highest occurrence of *Rhyssochonetes aurora* in this region. It is a rare component along with *Emanuella subumbona* and scattered auloporids within the bed. This distribution shows that a diminished Tully Fauna component did survive within the Tully well past deposition of the Smyrna Bed. In the creek below the road is an incomplete in-place succession of lower Tully beds probably corresponding to the Fabius Bed succession to the west. In addition to this, however, are large loose Tully blocks derived from the till upstream from the actual outcrop. One of these blocks, resembling the lower Smyrna auloporid-rich sandstone unit at Stop 6, is capped by a thin, chamositic limestone bed yielding abraded calcarenite with auloporids, and rare *Rhyssochonetes*. Moreover, this bed displays "mudcrack"-like fractures and well-developed microbialite stromatolites similar to those originally described by Heckel (1973) along the intra-Tully discontinuity at Bellona and those observed by the present authors in the Tully Valley (see Stop 2). We believe the loose block to be a Smyrna Bed erratic and that the stromatolitic bed is associated with the top-Smyrna transgressive flooding surface (see text for discussion).

- 79.4      1.3      Retrace to junction of Route 80 and Walt Phillips Road. Turn right (south) onto Route 80.
- 80.0      0.6      Strata of lower part of Cooperstown Member in stepped ramp falls

|      |     |  |
|------|-----|--|
|      |     | in floor of Center Brook to the right.   |
| 82.3 | 2.3 | Junction of Routes 8 and 80 at Five Corners. Turn right (south) on Route 8/80.   |
| 83.0 | 0.7 | Enter New Berlin.  |
| 83.6 | 0.6 | Route 80 splits to east (left) from Route 8 at center of New Berlin. Continue straight (south) on Route 8.   |
| 83.9 | 0.3 | Junction (fork) of County Route 13 and Route 8. Continue straight (south) on County Route 13.  |
| 84.1 | 0.2 | Cross Unadilla Creek, leave New Berlin and enter Otsego County.  |
| 85.0 | 0.9 | Creek to left displaying Cooperstown Member strata including two separated ledges of the Lansing Coral Bed.  |
| 85.7 | 0.7 | Borrow pit on right displaying strata of Sherburne Member.   |
| 85.8 | 0.1 | Upper Tully Formation clastic correlative succession in creek (Locality 26) below road on left.  |
| 89.4 | 3.6 | Borrow pit on right displaying Sherburne Member strata.  |
| 91.7 | 2.3 | Enter Town of Morris.  |
| 91.9 | 0.2 | Junction of Routes 51 and 13 in Morris. Turn left (east) onto Route 51.  |
| 92.2 | 0.3 | Leave Morris on Route 51. Butternut Valley to right.   |
| 95.8 | 3.6 | Junction of Route 51 and Route 12. Turn right (east) onto Route 12 and proceed towards New Lisbon and Gilbert Lake.  |
| 96.0 | 0.2 | Cross Butternut Creek.   |
| 96.2 | 0.2 | Enter New Lisbon.  |
| 96.5 | 0.3 | Junction of Route 12 with Route 14. Turn left (north-northeast) onto Route 14 and leave New Lisbon.  |
| 97.5 | 1.0 | Junction of Route 14 with Otto Stahl (dirt) Road. Turn right (southeast) onto Otto Stahl Road.   |
| 97.6 | 0.1 | Pull off on shoulder bordering tributary of Stony Creek. Exit vehicles and examine adjacent creek and abandoned quarry sections as well as outcrops along Otto Stahl Road. |

**Stop 8: New Lisbon Member Type Section and Higher Tully Clastic Correlative Units in the Vicinity of Otto Stahl Road.**

This section (Locality 34 of present paper) was the focus of detailed study by Cooper and Williams (1935) particularly for definition of the New Lisbon Member. However, a long and nearly complete section of the Tully clastic correlative succession is also present along and near this creek. Cooper and Williams (1935) established the New Lisbon Member as a 60-foot-thick succession of flaggy siltstone beds and shale interbeds rich in the brachiopod *Camarotoechia* (“*Leiorhynchus*”) *mesacostale*. The lower-middle part of the type section of this unit is exposed on adjacent Stony Creek (Locality 32) and the upper 38 feet of Cooper and Williams (1935) original “New Lisbon Member” is present at Stop 8. Cooper and Williams (1935) and Heckel (1973) both recognized that this unit was pre-Tully in its lower half and Tully equivalent (yielding *Rhyssochonetes* and *Tullypothyridina*) in its upper half. Thus, this unit filled the interval between the underlying Cooperstown Member, yielding *Pustulatia* (“*Vitulina*”), and fossil-rich strata

of the overlying lower-middle Tully succession ("Laurens Member" of Cooper and Williams, 1935; Figures 1, 3).

As noted in the text, the present authors have found a chamosite-bearing bed on this creek yielding *Rhyssochonetes*, rare *Hypothyridina*, conularids, and abundant *Emanuella* 28.8 feet below the base of the "Laurens Member" and 10 feet above the lower end of this outcrop (Figure 7). This unit, located nearly at road level in the creek, is herein taken to be the base-Tully contact (Figures 7, 11). Moreover, a chamosite-rich contact bed is now observed to floor the New Lisbon interval at nearby locality 31 allowing us to formally bracket the unit with physical contacts (Figures 7, 11). With this new information, we herein restrict the New Lisbon Member to include only pre-Tully highstand deposits between the two chamositic beds (Figures 7, 11). The "upper New Lisbon" of Cooper and Williams (1935) is herein included in the basal Tully clastic succession. We agree with Heckel (1973) that the 28.8 foot-thick succession of *C. mesacostale*-rich slabby siltstones and sandstones in the abandoned quarry and creek walls near the car pull off probably corresponds respectively to the DeRuyter and Cuyler beds in the Tully Limestone. Also significant is the fact that this pre-"Laurens Member" Tully interval appears to be absent in all localities between this section and Sheds (Figures 7, 11).

A few hundred yards east (uphill) from the car pull off are a few small road ditch cuts in the "Laurens Member" succession. The first of these exposes a 20 inch-thick silty mudrock unit abounding in *Emanuella*, *Echinocoelia?*, *Hypothyridina*, *Schizophoria*, *Spinatrypa*, and *Mucrospirifer*. This corresponds to Cooper and Williams, 1935, "second shell bed" of section parlance. 14 feet further up the road is another bed yielding abundant, well preserved *Tullypothyridina* and *Pseudoatrypa*. This is Cooper and Williams' (1935) "third shell bed" of the section. Finally, 12 feet above that unit is a thin, shell hash bed rich in *Emanuella*, *Echinocoelia?*, large burrow prods and phosphatic pebbles. We tentatively link the first shell bed in the road ditch to part of the Fabius Bed and one (or both) of the two higher shell beds to the Carpenters Falls-Smyrna bed interval in the Tully Limestone (Figures 7, 11).

Higher strata of the clastic Tully succession can be viewed only from private land in the creek. Hence, we will not see this part of the section. It is worth noting that the West Brook Bed is spectacularly developed on the south fork of this creek and is floored by an eleven-inch-thick clean, crystalline calcarenitic limestone bed yielding rugose and tabulate corals. It appears that there is eastward overstep of medial Tully strata by the sub-West Brook disconformity beneath this limestone layer.

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| 98.0  | 0.4  | Junction of Otto Stahl Road with Buck Schoolhouse Road at acute Y-intersection. Continue straight on Buck Schoolhouse Road. Strata of medial Tully Formation clastic correlative succession visible in creek falls to the right. |
| 98.05 | 0.05 | Intersection of Buck Schoolhouse Road with Gross Road. Turn right (south) onto Gross Road.   |
| 98.1  | 0.05 | Cross north fork of Stony Creek tributary.   |
| 99.5  | 1.4  | Junction of Gross road with County Route 12. Turn left (east) on Route 12.   |
| 101.2 | 1.7  | Entrance to Gilbert Lake State Park on the left. Continue southeast on Route 12.   |
| 104.6 | 3.4  | Enter Town of Laurens.   |

- 104.8 0.2 Junction of Route 12 and Main Street in Laurens (T-junction). Turn left (east) on Main Street.
- 105.2 0.4 Turn right on Otsego County Route 7A.
- 105.25 0.05 Bridge over Otego Creek.
- 105.5 0.25 Junction of Route 7A with Route 205. Turn left (north) onto Route 205.
- 106.9 1.4 "Houghtelings Glen" section of Cooper and Williams, 1935, to right. This outcrop (Locality 38) exposes an excellent lower Tully Formation clastic correlative succession (see text) and was intended as a stop. However, a major windstorm has tumbled trees blocking the stream course for the immediate future.
- 107.4 0.5 Junction of Route 205 with Dutch Hill Road. Turn right (east) onto Dutch Hill Road.
- 107.9 0.5 Region of smashed and toppled trees. This was the result of the same tornado (or downburst) that flattened trees at "Houghtelings Glen."
- 108.9 1.0 Intersection of Dutch Hill Road and East Road. Continue straight (east) on Dutch Hill Road.
- 109.0 0.1 Fork junction of Dutch Hill Road with Concrete Road. Bear right (east) on Dutch Hill Road.
- 111.9 2.9 T-junction of Dutch Hill Road with County Route 44. Turn right (south) on Route 44.
- 112.6 0.7 Junction of Route 44 with State Road 28. Turn right (west) onto Route 28.
- 113.1 0.5 Hamlet of Milford Center. Susquehanna Valley to left. Continue south on Route 28.
- 114.7 1.6 View of Goodyear Lake to left.
- 114.85 0.15 Pull off on right shoulder by long roadcut along Route 28. Exit vehicles and examine section.

**Stop 9: Tully Clastic Correlative Succession (or Tully equivalent Gilboa Formation) on South Side of Route 28 by Goodyear Lake.**

This is a "floating section" probable corresponding to part of the middle-upper part of the "Laurens Member" (Cooper and Williams, 1935). About 35 feet of section is exposed; spectacular flow rolls (seismites?) are visible in the lower third of the section; about two thirds-up is a 20-inch bed rich in *Tullypothyridina*, *Echinocoelia*, *Mucrospirifer*, *Spinatrypa* and bivalves. In the topmost 5 feet, pelmatozoan debris is prominent at several levels. This section, and, to a lesser extent, Stop 8, stand in stark contrast to the sparsely fossiliferous sections (Stops 5, 6, 7) of lower-middle Tully facies in the Sherburne-New Berlin area (Figures 7, 8, 11). We hypothesize that a structural, bathymetric trough centered at New Berlin separated the Tully carbonate platform in west-central New York from a sandy shelf area in eastern New York (Figure 12). Hence, the coarse, fossil-rich, Tully clastic deposits of the Otego and Susquehanna Valleys corresponds to the "Chemung" facies belt for the Taghanic Stage (Baird and Brett, 2003). Current explorational work is directed to localities in this valley and further east to

characterized both the biofacies and stratochronology of this poorly known interval. We hope to determine whether inner shelf biotas at the Taghanic level have a unique taxonomic composition or correspond to the Ithaca Fauna of overlying units.

A great final mystery is centered on the relationship of various marine units (Cooperstown Member, clastic Tully succession, and Geneseo Member) to the Gilboa Formation. This problem is beyond the scope of the present report, but is the subject of ongoing investigation (see Bartholomew and Brett ???).

115.2      0.35      Junction of Route 28 with Route 7. Continue straight (south) on Route 28.

115.4      0.2        Bridge over Susquehanna River.

116.4      1.0        Junction of Route 28 with Interstate 88. Enter I-88 westbound to Oneonta.

End of Saturday Road Log

### ROAD LOG ADDENDUM

This is Sunday's Road Log venue following completion of Saturday stops (in reverse). The addendum has an abbreviated road route description for two stops west of the Tully Valley.

Leave from Carrs' locality driveway section (Stop 2 of Saturday venue) to junction of Long Road and Route 80. Turn left onto Route 80 (dangerous right turn) and turn around to be eastbound on Route 80.

Enter Interstate 81 (southbound) and continue south to Cortland. Exit Interstate 81 at exit 11 and take Route 13 southwest towards Ithaca.

Route 13 merges with Route 34. Continue south on Route 13/34 into Ithaca.

Junction in Ithaca with Route 89. Turn right onto Route 89 and proceed northwest out of Ithaca along west side of Cayuga Lake. When you get to Taughannock Falls State Park, turn left into parking area at foot of hill just south of bridge over Taughannock Creek. Waterfalls over Tully Formation visible from road. Park vehicles and proceed to falls overlook.

Stop 10: Tully Formation Section on Taughannock Creek. Type Section for the Taghanic Stage.

This outcrop provides an excellent look at the Tully Formation section and adjacent units. Several feet below the base of the Tully is the westernmost occurrence of the Lansing Coral Bed; *Favosites*, *Heliophyllum* and other corals can be found as well as large fistuliporoid bryozoans and diverse brachiopods. The Spezzano submember is recessive below the disconformable base of the Tully. Large hypichnial burrow prods characterize the basal Tully surface where the Hamilton shale has eroded away below it. The

remaining Tully succession to the base of the 5-foot-thick Fillmore Glen succession is massive, dove-white weathering fine limestone that is clean enough for quarrying as is done at Portland Point across the lake. Key interval units and contacts will be pointed out at this stop.

This is the key reference section for the Taghanic provincial stage, the Taghanic Unconformity and the Taghanic bioevent (see Johnson, 1970; Heckel, 1973; Aboussalam and Becker, 2001; Baird and Brett, 2003). However, it is significant that this Tully section is very incomplete due to westward erosive beveling of most lower Tully strata between Skaneateles Lake and this locality (Heckel, 1973). As such, the Carpenters Falls Bed, yielding *Tullypothyridina* and other lower Tully taxa, rests disconformably on the Hamilton. This is Johnson's (1970) "Taghanic Unconformity" which was believed to mark the start of a major craton-ward stratigraphic onlap event. Careful examination of higher Tully in western New York suggest, however, that the lowstand base-line for the transgressive Taghanic onlap event is at a higher level within the upper Tully Member or near the top of the Tully (Baird and Brett, 2003). The lowstand base-line for the upper Tully TST interval is the regressive Bellona Coral Bed and the actual transgressive onlap is understood to commence at one or more horizons bounding (or within) the Fillmore Glen Bed (Baird and Brett, 2003). The top-Moravia Bed omission horizon (knobbly surface) and five calcilutite-shale repetitions comprising the Fillmore Glen succession are well displayed at this section. Note the black, fissile strata of the Geneseo Member in the bank walls. This unit records major transgressive deepening in the foreland basin owing to combined effects of eustatic transgression and thrust loading to the east (Johnson et al., 1985; Ettensohn, 1998). Higher outcrops of this unit and overlying deposits, exposed in the great fall on this creek, are worth seeing, but are too far upstream to be accessed on this trip.

Continue northwest on Route 89. Turn left (west) at sign for Interlaken.

Turn right (north) onto Route 96 in Interlaken and continue northwest to Ovid.

Route 96 splits into Routes 96 and 96A. Continue straight (west) on Route 96A. Route 96A passes Willard, the now-defunct Seneca Army Depot, and Sampson State Park.

Junction of Routes 96A and US 5/20 and continue to west edge of Geneva.

Junction US 5/20 with Preemption Road near west edge of Geneva. Turn left (south) onto Preemption Road.

Enter village of Bellona. Just past bridge over Kashong Creek in middle of village, pull left (or right) into available parking. Exit vehicles and proceed to bridge near abandoned stone mill building.

Stop 11: Tully Limestone section on Kashong Creek; Stromatolites Above Mid-Tully Unconformity; Bellona Bed; Undulatory Top-Tully Erosion Surface.

We will descend into Kashong Creek upstream from the bridge and proceed to a breached up-fold of Tully in the creek floor. Along the contact between the underlying Carpenters Falls Bed and the overlying Taughannock Falls Bed is a layer of fracture networks, domal limestone knobs and stromatolitic carbonate deposits filling low areas peripheral



to the knobs. This is Heckels', 1973, stromatolitic horizon associated with his regional intra-Tully unconformity. Close examination of the stromatolite texture shows that the lamination is arrayed concentrically around the structureless limestone knobs and parallel to fracture networks. Very close examination of the laminite shows that it has a perpendicular "pillared" microstructure that is normal to the visible lamination, a characteristic that Heckel, 1973, also noted. The stromatolitic layer is usually only millimeters-thick in swales and it is erosionally breached over knobs. The base of the stromatolitic layer contains skeletal debris, pyrite, phosphate grains and some fish bones; this marks the horizon of the discontinuity surface. It is important to compare the stromatolitic surface at this locality with material from Stops 2 and 7 as these latter occurrences are believed to be essentially the same level.

Several feet above the stromatolitic horizon is the Bellona Coral Bed which is thin, muddier than surrounding Tully facies, and profusely fossiliferous. Where it is swept clean on bedding surfaces, hundreds of corals can be seen. Easily spotted taxa within the bed include: *Heliophyllum*, *Cystiphyllum*, *Favosites*, *Alveolites*, and the brachiopod *Spinatrypa*. This is the "coral plantation" of Heckel, 1973, that serves as a key datum in mapping over huge areas. Above the Bellona Bed are clean carbonate deposits of the Moravia Bed. However, Heckel, 1973, correctly noted that the top-Tully contact at this locality is undulatory. Up to a foot of Moravia is present above the bridge, but, below the bridge, the Geneseo black shale rest directly on coral-rich Bellona deposits.

The Geneseo Member rests on the Tully with knife-sharp definition. Tully carbonate is locally separated from the well-jointed Geneseo by a thin lag bed of pelmatozoan calcarenite, but not detrital pyrite or bone debris. However, at Gorham, the next section to the west, detrital pyrite and bone debris characteristic of the Leicester pyrite rests on the Bellona Bed. Still further to the west, the base-Geneseo corrosional discontinuity is observed to cut out the Tully Formation and truncate the topmost beds of the Windom Member (Baird and Brett, 1986). The occurrence of channel-like grainstone lentils within the Geneseo Shale 16 inches above the Tully contact at Bellona, suggest the possibility that the real Leicester horizon is actually within the basal Geneseo on this creek and that the observed Tully-Geneseo contact at Bellona is a pre-Leicester surface that may project eastward to some level within (or at the base of-) the Fillmore Glen Member. This issue is still being investigated.

End of Sunday Road Log.

***TRIP A-2***  
**Coastal Margin Interfluvial Paleosols and their Stratigraphic  
Relationships with Tidally-Influenced Deltaic Deposits  
in the Sonyea Group (Frasnian)  
of Northwestern Delaware County, New York**

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### **Introduction**

The Catskill Clastic Wedge has been broadly subdivided into three magnafacies: Chemung, Cattaraugus and Catskill. These magnafacies have been interpreted as representing offshore, coastal/nearshore and terrestrial (fluvial) paleoenvironments, respectively. Of these magnafacies, the Cattaraugus has been the most poorly characterized and the nature of the Catskill shoreline has long been the subject of debate. Tectonic, eustatic and climatic processes interacted in complex ways to produce nearshore paleoenvironments in the Catskill Wedge which had aspects of both marine and terrestrial settings.

In this study, we describe tidally-influenced sediments of the Cattaraugus Magnafacies in parts of the Sonyea Group (Late Devonian, Frasnian) in Delaware County, New York (Fig. 1). However, our main focus is on paleosols in these strata as key features in understanding the complex nearshore environments which bordered the Catskill Sea. Several types of paleosols are documented, which yield insights into depositional and climatic processes at the Sonyea shoreline. More significantly, we show that paleosols, a typical feature of the Catskill Magnafacies, developed directly on sediments deposited under brackish, deltaic/estuarine conditions and on definitively marine sediments. In addition, many of the paleosols that we describe are immediately overlain by sediments deposited under marine or brackish conditions.

Such abrupt paleoenvironmental changes provide insights into processes of tectonic subsidence, eustatic changes in sea level and autocyclic processes, such as avulsion, which controlled the nature and position of the Catskill shoreline during Sonyea deposition.

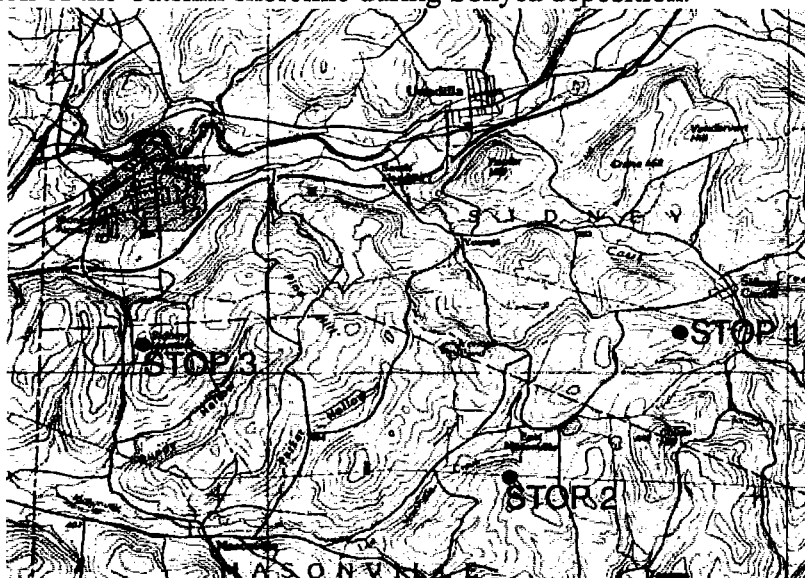


Fig. 1 Location map of field trip stops in the vicinity of Sidney and Sidney Center, New York. Sidney and Unadilla 7.5 minute quadrangles.

### Chemung Magnafacies

The Chemung Magnafacies is defined as the fossiliferous rocks that represent the open-marine inner shelf. It comprises complexly interbedded sandstone, siltstone, and shale. Chemung stratigraphy is dominated by thick beds of sandstone and siltstone, separated by thin dark shales. Several authors have identified repetitive coarsening-upward sequences of mudstone through sandstone 15-30 meters thick within the Chemung Magnafacies in several other groups in the Middle and Upper Devonian (Craft and Bridge, 1987; Kirchgasser and others, 1994). In this study, two facies have been assigned to the Chemung Magnafacies. These are: 1) the dark gray shale facies, and 2) the hummocky cross-stratified facies. Continuous exposures of these facies are more abundant than rocks of the overlying Cattaraugus Magnafacies, but outcrops are still difficult to trace laterally.

### Dark Gray Shale Facies (Facies $S_{dg}$ )

Facies  $S_{dg}$  occurs as uniform, dark gray shale and siltstone or as distinct interbeds within the hummocky cross-stratified facies with rare lenses of very fine sublitharenite. Planar bedding is most common in this facies. Discontinuous, ripple cross-lamination in sandy siltstone interbeds is cryptic, but occurs throughout the facies. These are interpreted as distal tempestites. Biogenic structures and invertebrates are absent in facies  $S_{dg}$ .

Discrete intervals of trough and wedge-shaped forms characterize this facies. Basal surfaces truncate underlying planar-laminated shale, and are overlain by a wedge of shale. Most of the truncation surfaces occur as solitary wedges, which overlie a sharp, concave-up discontinuity surface. The shale is inclined along the discontinuity surface and is in angular discordance with underlying planar beds. Inclination of shale decreases to sub-horizontal to horizontal planar laminations progressively upward within the wedges. Truncation surfaces also occur as facing pairs of intersecting, U-shaped troughs filled with shale that truncate each other. The shale that fills the U-shaped troughs is form concordant, and progressively flattens upward within troughs. Most wedges and troughs measure 5 to 20 meters in width, and truncate 1 to 2 meters of underlying shale.

Frasnian time experienced an overall rising sea level (Johnson, Klapper, and Sandberg, 1985), which is manifested in the Sonyea Group as numerous black shale tongues of facies  $S_{dg}$  that are probably coeval with drowned shoreface systems. Facies  $S_{dg}$  records three rapid transgressive pulses within an overall transgressive succession. Relative rapid sea-level rises associated with facies  $S_{dg}$  were apparently induced by a combination of epeirogenic lithospheric downflexure and differential subsidence (Quinlan and Beaumont, 1984), resulting in a rise in base level and trapping clastics in drowned estuaries, delta plains, and distal alluvial plain rivers (Marzo and Steel, 2000; Kirchgasser and others, 1994). The complete absence of invertebrates and biogenic structures, and the predominance of silt and clay in this facies indicate that deeper, anoxic conditions were introduced into formerly shallow areas.

The truncation surfaces that form shallow wedges and troughs are probably rotational gravity-slide failure scarps (Davies, 1977). Large quantities of unlithified sediment were disaggregated during liquefaction and removed by gravity sliding, owing to the fact that large-scale breccia, rotated blocks, or crumpled or other disturbed bedding structures are absent. The rotational slumps were probably triggered by storm events, overloading, or seismic activity.

Facies  $S_{dg}$  resembles descriptions of the Sawmill Creek Shale (Sutton, Bowen, and McAlester, 1970). However, we are reluctant to make this assignment because many shale intervals in the area meet the general description of Sutton and others (1970). Correlation of Sonyea black shale tongues (e.g., Montour, Sawmill Creek, Moreland) into the nearshore environments remains problematic.

#### **Hummocky and Swaley Cross-Stratified Facies ( $H_{cs}$ ) and Amalgamated Hummocky Cross-Stratified Subfacies ( $H_{cs-A}$ )**

The Hummocky Cross-Stratified Facies is commonly interbedded with the dark gray shale facies. Hummocky cross-stratification is the dominant sedimentary structure in facies  $H_{cs}$  and shows striking similarities to examples cited by Craft and Bridge (1987), Halperin and Bridge (1988), Hamblin and Walker (1979), and McCrory and Walker (1986). Hummocky cross-stratified beds with wave-rippled caps are primarily found interbedded with siltstone and shale, but also occurs as amalgamated beds of very fine sublitharenite. Siltstone and shale interbeds are planar-laminated, blocky, or exhibit small-scale, cryptic and discontinuous ripple cross-lamination, and are commonly heavily bioturbated. Hummocky cross-stratified beds also occur as sheet-like sandstone beds with planar-lamination with lateral transition to hummocky cross-stratification similar to beds described by Halperin and Bridge (1988).

Hummocky cross-stratified beds have sharp erosional bases accompanied by soft sediment deformation. Directional sole marks such as flute and tool marks, coquinite layers, coquinite-filled scours, coquinitic hummocks or ripple forms, graded bedding and conglomerate are commonly found at the base of hummocky cross-stratified sandstones. Channel-filling sandstone bedsets, large sandstone pillows with kneaded surfaces, sandstone pillows with relatively undeformed strata, and gutter casts also occur at the base of sandstones in this facies. Additional details of structures and the diverse fossil assemblage present in facies H<sub>cs</sub> are discussed in Bishuk, Ebert, and Applebaum (1991) and Bishuk (1989).

Hummocky cross-stratified intervals become increasingly amalgamated higher in the stratigraphic section (Subfacies H<sub>cs</sub>-A). A sharp, scoured and loaded contact with numerous ball and pillow structures and intraformational conglomerate occurs where amalgamated hummocky cross-stratification commences. Very fine and fine sublitharenite with rare conglomerate composed of shale and siltstone clasts are the dominant lithologies.

In amalgamated beds, concave-upward swaley surfaces pass laterally into convex-upward hummocky surfaces, thereby producing adjacent hummocks and troughs. Hummocks and swales are truncated laterally by hummocky cross-stratification or planar strata. Pebble lags commonly line these erosional surfaces. Most amalgamated beds are form-concordant, but discordant sets also occur. Structures, internal to amalgamated hummocky bedforms, include climbing ripples with steeply inclined axes of propagation (70 to 80 degrees) along the flanks of hummocks. The ripples generally show migration toward hummock crests. Hummocks and swales commonly contain several horizons of *Rhizocorallium* (?) burrows and sinuous epichnial trails of *Scalarituba*. Amalgamated hummocky beds rarely have distinct sole marks along erosional bases. Where present, sole marks include flute casts, load casts, and a variety of tool marks.

Fossils decrease markedly in abundance and diversity where amalgamated hummocky cross-stratification dominates the section. However, carbonized plant fragments are abundant. The most common fossil is the rhynchonellid, *Cupularostrum*. Other fossils, in order of decreasing abundance, include the bivalve, *Sphenotus*, the spiriferid, *Platyrachella*, other unidentifiable brachiopod fragments, fish fragments, crinoid ossicles, and the bivalve, *Cypricardella*. The fossils are most commonly found associated either with loaded scour and fill bases as coquinite shelly lags or within intraformational conglomerates.

Hummocky cross-stratification in the study area is interpreted as a storm-produced structure occurring below fair weather wave base and above storm wave base (Harms, 1975, and many others). The lower portion of the hummocky cross-stratified facies is therefore interpreted as representing deposition on a storm-dominated shelf. Similar interpretations have been made for other parts of the Chemung Magnafacies (Craft and Bridge, 1987; Halperin and Bridge, 1988).

The stratigraphic position of this facies, the predominance of fine sand, and a marked upward decrease in fauna suggest that the amalgamated hummocky cross-stratified portion of this facies occupies the lower shoreface. Siltstone and shale interbeds are rarely preserved, indicating that waves or currents (e.g., longshore-, tidal-, and storm-driven) had effectively winnowed the fine fraction (Swift, 1984). Paleoflow direction in this facies is consistently to the north, which may represent the longshore or geostrophic current direction. The marked decrease in faunal abundance indicates environmental stress associated with the nearshore zone (Thayer, 1974).

The hummocky cross-stratified facies is integrally involved in the overall progradation of the Catskill clastic wedge. Vertical upbuilding of hummocky cross-stratified beds at rates greater than the average rate of subsidence caused shallowing in the nearshore zone (Hamblin and Walker, 1979). Given the high volume of sediment delivered to the Catskill Sea, vertical upbuilding of hummocky beds was probably so rapid that accommodation space was consumed rapidly, thereby accelerating progradation rates of shelf and shoreface deposits. Progradation ceased when interrupted by rapid transgressive pulses and when tectonic conditions and/or the weight of nearshore deposits were sufficient to cause rapid subsidence. This is substantiated by hummocky cross-stratified beds overlying paleosols at outcrop 6. Since hummocky cross-stratification is chiefly deposited below fair weather wave base, relative sea level change must have risen by a minimum of 10-15 meters, which is attributable to a combination of eustasy and subsidence. Subsidence alone may not account for this disparity. A major river avulsion and/or a directional change in dispersal of sediment from the source area likely contributed to abandoned or diminished nearshore deposition, which would allow subsidence to outpace accumulation.

### **Cattaraugus Magnafacies**

The Cattaraugus Magnafacies comprises the transitional rocks between obvious non-marine river channel and floodplain deposits of the Catskill Magnafacies and the obvious fossiliferous, open-marine deposits of the Chemung Magnafacies. Hence, the Cattaraugus Magnafacies is somewhat ambiguous in that it may possess aspects of both marine and non-marine environments (Linsley, 1994). An example of the intriguing, yet confusing nature of the Cattaraugus Magnafacies is the presence of channelized, fluvial-like deposits with marine shells at the base, which have been documented in several Upper Devonian studies (Johnson and Friedman, 1969; Bridge and Droser, 1985; Halperin and Bridge, 1988; and Bridge and Willis, 1991). In these examples, depositional environments change abruptly both laterally and vertically. These authors also note that marine and non-marine environments are separated by erosion surfaces where beach and intertidal facies might be anticipated. Within the Sonyea Group, we have documented multiple facies which offer new insights into the marine-brackish transition zone of the Cattaraugus Magnafacies. Regrettably, the transition between brackish and fresh water settings has not received additional clarification because no rocks that may be definitively assigned to the Catskill Magnafacies have been identified in this study area.

Shoreface deposition in the Cattaraugus Magnafacies is recorded by three facies. These are: 1) the trough cross-bedded facies; 2) the planar bedded multistorey sandbody facies; and, 2) the heterolithic bedded facies. Limited exposure has made reconstruction of facies relationships difficult. However, it is readily apparent at outcrops that the three facies are stacked in repetitive sequences. A survey of location and elevation of the outcrops using GPS as well as quarry activity in the last decade has helped to alleviate some of the difficulty in facies reconstruction (See Appendix I: listing of outcrops).

### **Trough Cross-Bedded Facies ( $T_{xb}$ – with Subfacies $T_{xb-td}$ and $T_{xb-ti}$ )**

The trough cross-bedded facies ( $T_{xb}$ ) consists of two subfacies, tidally-dominated subfacies ( $T_{xb-td}$ ) and tidally-influenced subfacies ( $T_{xb-ti}$ ), which are both assignable to an unnamed formation

of the Cattaraugus Magnafacies. Subfacies  $T_{xb}$ -td occurs stratigraphically lower than subfacies  $T_{xb}$ -ti. Both subfacies are lithologically and structurally similar, but the sediment in subfacies  $T_{xb}$ -td is clearly differentiated into laminae of different well-sorted grain sizes and contains a greater abundance of tidally-produced sedimentary structures.

### **TIDALLY-DOMINATED SUBFACIES $T_{XB}$ -TD**

Subfacies  $T_{xb}$ -td consists of repetitive fining-upward sequences of moderately well sorted, fine sublitharenite. Sequences are separated by erosional scour surfaces and reactivation surfaces. Erosional surfaces are locally marked by a coarse sandstone-supported conglomerate and/or pebble lag. Clast lithologies are mainly siltstone and shale granules and pebbles, with rare calcrete clasts. Carbonized stems, branches and occasional logs are common at erosional bases and are locally abundant throughout the facies. Rare concentrations of invertebrates line the bottoms of troughs at and just above erosional bases at some outcrops. These concentrations are nearly monospecific, comprised of either the rhynchenellids *Cupularostrum* or *Camarotoechia* with rare crinoid ossicles. Trace fossils are restricted to rare occurrences of *Archanodon*-like, roughly cylindrical, nearly vertical burrows measuring 7-10 cm in diameter and up to 0.75 m long (Fig. 2). The caliber of the *Archanodon*-like burrows is consistent with *Archanodon* burrows identified by Bridge and others (1986), Gordon (1988), and Miller (1979). These burrows occur only at outcrops 7 and 21. No body fossils of *Archanodon* have been found.



Fig.2 Trough cross-bedded facies showing tidal drapes and coarse-fine bundling. An *Archanodon*-like burrow is present to the left of the scale.

Small- to medium-scale trough cross-bedding is the dominant sedimentary structure in the facies (Fig. 2). Many sets are sigmoidal in shape. The most striking and characteristic feature of subfacies  $T_{xb-td}$  is that the sediment on trough foresets is clearly differentiated into repetitive couplets of laminae of different well-sorted grain sizes. These couplets consist of fine sand alternating with very fine sand and silty mud drapes. Weathered outcrop surfaces commonly accentuate the alternating grain sizes, where finer grained drapes are weathered into recessed grooves adjacent to more resistant ridges of fine sand. Thick-thin alternations are observed in couplet bundle thicknesses on trough foresets. In addition, thickness variations in couplet bundles are also observed in some bottomset beds. Dip angles on trough foresets range from 10 to 25 degrees with an average inclination of approximately 15 degrees. Cross-beds commonly climb at low angles of propagation and are largely unidirectional, but locally are multidirectional. In general, the thickness of cross-bed set thins upward within the facies with reactivation surfaces become more common. In addition, dip angles on troughs decrease upward in the facies. Dune bedforms are uncommon. These sedimentary structures are strikingly similar to those described by de Mowbray and Visser (1984) and Nio and Yang (1991).

Some cross-beds contain normally graded, lenticular laminae with asymmetrical ripples and closely associated crinkled laminations, which oppose the dominant cross-bed paleoflow direction. The asymmetrical ripples migrate up the avalanche slope of dunes, occur in discrete units within the foresets, and are commonly bounded by very fine sand and silty mud drapes. Therefore, they clearly are not backflow ripples. In general, the asymmetrical ripples migrate from the base to the middle portion of foresets, and less commonly migrate to the crest of foresets. Paleocurrent directions obtained from trough cross-beds and asymmetrical ripples are dominantly northward, ranging from northwest to northeast. Asymmetrical ripples on dune foresets shows a subordinate southward paleoflow (southeast to southwest).

Some trough cross-beds are truncated by interbeds of fine sandstone that appear massive, bioturbated, and relatively structureless as compared to subjacent units. These interbeds are subhorizontal and of variable thickness. They pinch and bulge laterally and are bounded by recessively weathered grooves of very fine sandstone drapes. Some of these interbeds possess suspect, wavy crinkle laminations and crude bulges resembling starved ripples. The frequency of these massive interbeds increases upward in the facies.

Trough cross-beds are commonly intercalated with vertically-stacked, thinly-interlaminated couplet sequences of rhythmites consisting of fine sand with very fine sand drapes or very fine to fine sand with silty sand or silt-mud drapes (Fig. 3). These rhythmites are arranged as horizontal to subhorizontal planar lamination and gently inclined, long planar-tabular cross-beds (dip angles less than 5 degrees). Two types of lamina occur in couplet sequences. As defined by Williams (1991), simple laminae include only one sandy layer with one fine drape. Composite laminae, which consist of semilaminae embedded within simple laminae, are less common in the study area. Each simple lamina or semilamina is a graded layer 0.2 to 3 mm thick with a lower layer of fine sand passing upward into a thin band of silty sand or silt-mud. Double mud drapes also occur, but are uncommon.

Couplets appear to be unequal in thickness with a regular alternation of thick and thin couplets. These alternations are arranged in groups of thicker couplets with fine sand and very fine sand drapes and groups of thinner couplets with very fine to fine sand and silty mud drapes. In addition, the silt-sand rhythmites show a cyclical pattern where regular alternations of sand-poor packages and sand-rich packages can be identified at intervals of 17 to 22 cm. Rhythmites in the



study area are similar to examples described by Nio and Yang (1991), Williams (1991), and Fenies and Tastet (1998). In general, planar-laminated interbeds become more common upward in the facies.

Contacts between subfacies  $T_{xb-td}$  and other facies units are covered throughout the study area. However, based on the distribution of closely-spaced outcrops, subfacies  $T_{xb-td}$  appears to occur consistently between facies  $H_{cs-a}$  and  $P_{bms}$ . At outcrop 21, the upper contact of  $T_{xb-td}$  with the overlying facies  $P_{bms}$  appears to be gradational. Subfacies  $T_{xb-td}$  is laterally traceable for approximately 7 miles from the east at outcrop 20 in Sidney Center to the south-west at outcrop 7 on Sidney Mountain. Estimated maximum thickness of subfacies  $T_{xb-td}$  is approximately 10-15 meters, based on moderately-spaced outcrops in Sidney Center.



Fig. 3 Sub-horizontal gently inclined tidal rhythmites.

New observations from subfacies  $T_{xb-td}$  have been added to the data of Bishuk, Ebert, and Applebaum (1991), and now provide overwhelming evidence of tidal activity along the Frasnian paleoshoreline. The repetitive couplets of laminae of different well-sorted grain sizes are clear evidence of tides in subfacies  $T_{xb-td}$  (See also de Mowbray and Visser, 1984; Smith, 1988; and Nio and Yang 1991). The unidirectional nature of the trough foresets (i.e., paleoflow toward northwest) records a strongly asymmetric to asymmetric, ebb-directed tide. The asymmetrical ripples and closely associated crinkled laminations, which oppose the dominant paleoflow, represent flood-directed subordinate tidal currents.

It is unclear whether the large caliber vertical burrows with spreites found in subfacies  $T_{xb-td}$  were actually made by *Archanodon* or a bivalve with similar feeding and dwelling habits. The presence of *Archanodon*-like burrows in tidally-influenced deposits is intriguing, since it is widely accepted that *Archanodon* inhabited primarily fresh water fluvial environments (Bridge and others, 1986, Thoms and Berg, 1985). Thoms and Berg (1985) stated that the burrows are found only in nonmarine deposits near the Catskill paleoshoreline. Miller (1979) and Friedman

and Chamberlain (1995) have suggested alternative interpretations that similar burrows occur in tidal channel and shallow subtidal deposits. Gordon (1988) states that if all of the structures that are currently called "large vertical (meniscate) burrows" have a common origin, then an explanation for their broad environmental range (from fully marine to clearly nonmarine deposits) is warranted. As a result, the burrows should not be interpreted as structures resulting from a "freshwater" bivalve. It is clear that the *Archanodon*-like burrows in the study area inhabited tidal channel and tidal flat locales. Since no body fossils have been found associated with the burrows and in subfacies  $T_{xb-td}$ , we cannot positively identify them as having been made by *Archanodon*.

Sedimentary structures in the study area indicate that sedimentation patterns near the Frasnian shore are more consistent with estuarine than deltaic processes. Trough crossbeds record deposition by migrating dunes within a laterally accreting, tidal inlet and tidal creek complex at or near the mouth of a tidally-influenced delta. These deposits formed across an area from the tidal inlet at the river mouth into the estuarine funnel and into protected tidal creeks nearest to the river mouth. This interpretation is supported by the predominance of sigmoidal trough crossbeds, evidence of current reversals with differing flow magnitude, a sparse marine fauna, and the stratigraphic position of this facies, between the multi-storey planar-bedded facies (upper flow regime sand flat deposits of the medial portion of the estuarine-like river mouth of the delta) and the underlying amalgamated hummocky cross-stratified facies (storm-influenced lower shoreface/proximal offshore deposits). Paleoflow was dominantly toward the north, with minor variations toward the northeast and northwest, which suggests that tidal inlets migrated in this direction. This northerly paleoflow may record the dominant storm track, the prevailing wind direction and longshore current, or dominant tidal flow. More detailed interpretation of these and other structures within this facies may be found in Bishuk, Hairabedian and Ebert (in preparation).

### ***TIDALLY-INFLUENCED SUBFACIES $T_{XB-TI}$***

Subfacies  $T_{xb-ti}$  is nearly identical to subfacies  $T_{xb-td}$ , but with a few important differences. Like subfacies  $T_{xb-td}$ , subfacies  $T_{xb-ti}$  consists of repetitive fining-upward sequences of moderately well sorted, fine sublitharenite. Small- to medium-scale, commonly sigmoidal trough cross-bedding is the dominant sedimentary structure. Trough cross-bedding is intercalated with horizontal planar lamination and gently inclined, long planar-tabular cross-beds with dip angles less than 5 degrees, which become dominant upward in the facies.

The most notable difference is that subfacies  $T_{xb-ti}$  rarely displays the repetitive couplets of laminae and associated differential weathering described above. Instead, very fine sand and silty mud drapes occur sporadically on asymmetrical ripples which migrated up the avalanche face of trough cross-beds. The asymmetrical ripples occur near the base and toe of foresets similar to those described by de Mowbray and Visser (1984) and Nio and Yang (1991). Rare occurrences of reformed ripples with small-scale herringbone cross-lamination are found at outcrop 42. The herringbone cross-lamination is confined to scours at the crests of ripples marked by reactivation surfaces. These reformed ripples display asymmetry opposite the dip direction of the primary cross-lamination within the ripple.

The basal contact of  $T_{xb-ti}$  with underlying paleosols is visible at outcrops 43, 47, and 53 (Figs. 4, 5). The contact is sharp, planar, and non-erosive (e.g., the A horizon in underlying paleosols

is preserved), but when traced laterally across the outcrop, the contact becomes locally erosive with channel forms up to 0.5-0.75 m deep. Medium to coarse sand-supported conglomerate and/or pebble lags measuring up to 0.3 m thick overlies the contact. Clast lithologies are mainly siltstone, shale, and calcrete granules and pebbles. Locally, the basal contact and base of trough cross-set boundaries are spectacularly ornamented with flute casts and tool marks. Although paleocurrents from flutes show a dominant northeast to northwest azimuth, the basal contact at outcrop 43 has flutes which are oriented to the southeast. In addition, small dunes with trough cross-stratification immediately overlying the basal contact at outcrop 43 also bear consistent southeasterly azimuths. At outcrop 53, conglomerate-filled channel forms cut into trough and planar, tabular cross-strata. These concave-up channel forms measure up to 1.5 meters deep and 5-10 meters wide.

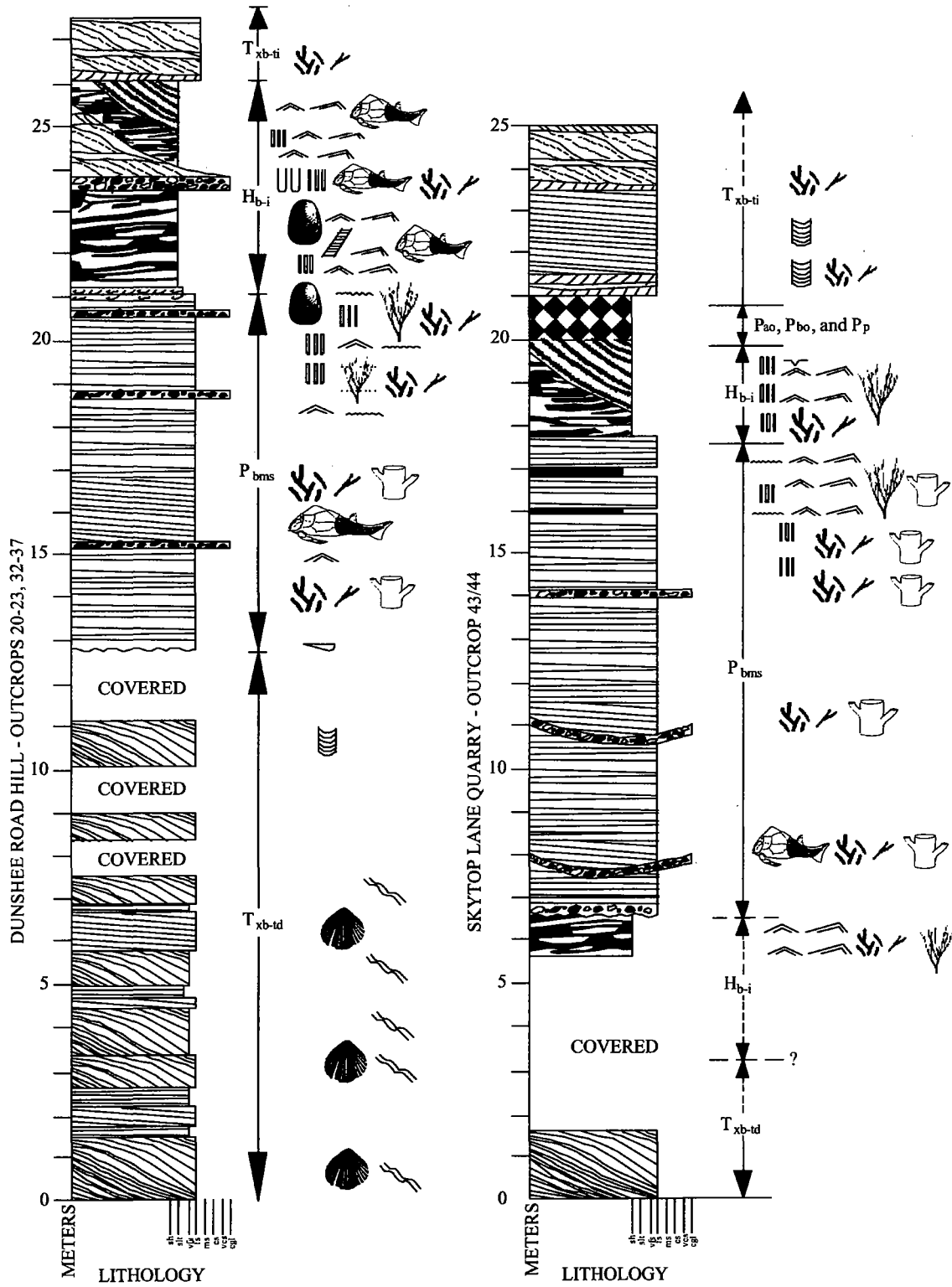


Fig. 4 Simplified stratigraphic columns for outcrops on Dunshee Road Hill and Skytop Lane Quarry (STOP #1).

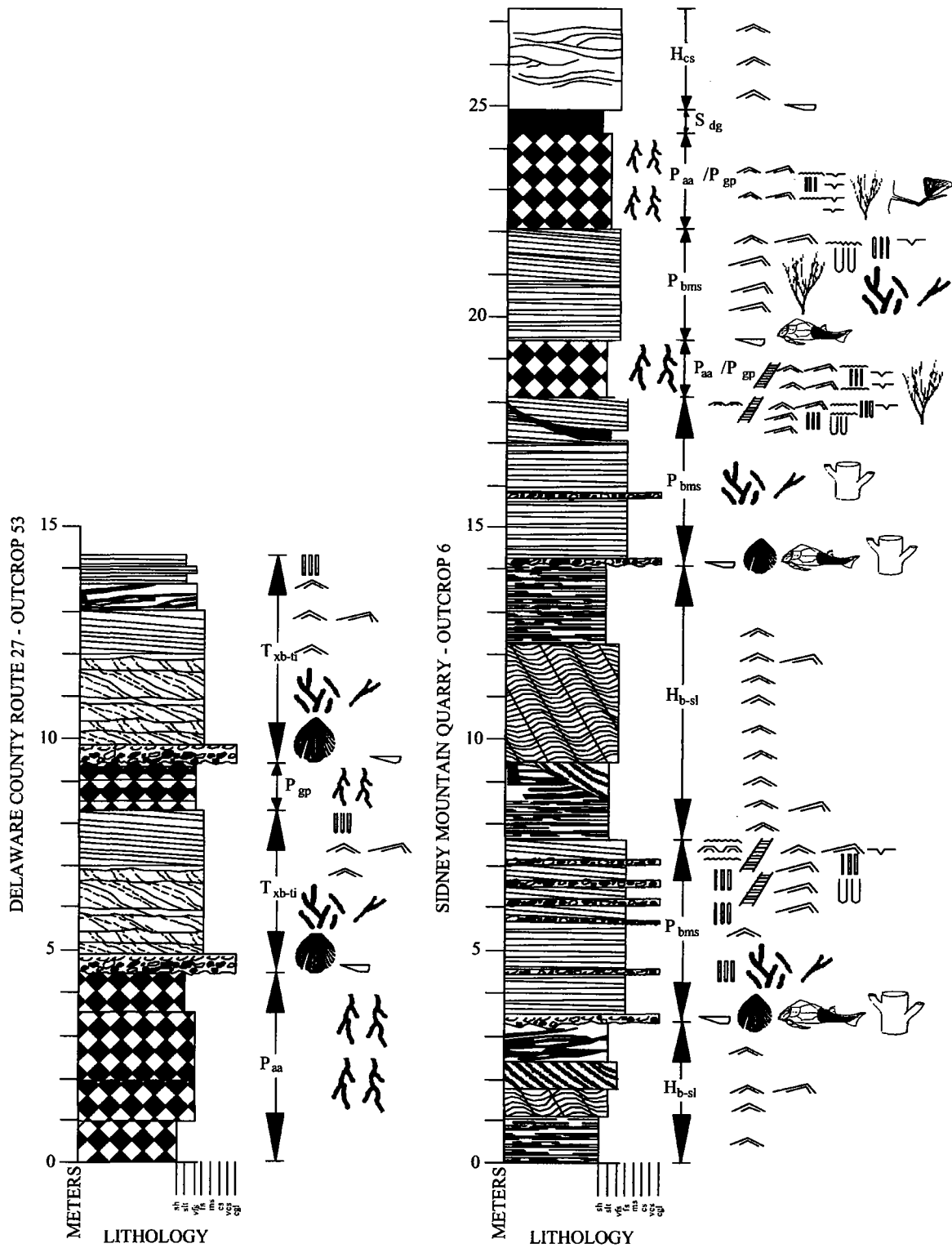


Fig 5. Simplified stratigraphic columns for Outcrop 53 and Sidney Mountain Quarry (STOP #3)

## Explanation for Figures 4 and 5

### LITHOLOGY

sh = Shale/Mudstone

slt = Siltstone

vfs = Very Fine Grained Sandstone

fs = Fine-Grained Sandstone













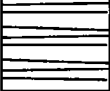





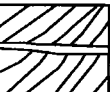














ms = Medium-Grained Sandstone

cs = Coarse-Grained Sandstone

vcs = Very Coarse-Grained Sandstone

cgl = Conglomerate

### SEDIMENTARY & BIOGENIC STRUCTURES

|   |  |   |   |
|---|--|---|---|
|    | Shale/Mudrocks (Mudstone and Siltstone)                  |   | Hummocky Cross-Stratification   |
|    | Streaky Lamination/Lenticular Bedding (Undifferentiated) |   | Sparse Shell Lag<br>Extraformational Pebbles<br>Intraformational Pebbles<br>Erosion Surface |
|   | Flaser and Wavy Bedding (Undifferentiated)               |   | Paleosol  |
|  | Flat Heterolithic Bedding                                |    | Symmetrical ripples   |
|  | Inclined Heterolithic Bedding                            |    | Flat crested ripples  |
|  | Intercalated Inclined and Flat Heterolithic Bedding      |    | Ladderback ripples  |
|  | Planar Lamination  |    | Asymmetrical ripples  |
|  | Subhorizontal Planar Tabular Cross-Strata                |    | Climbing asymmetrical ripples & crinkle laminations (oppose dominant cross-bed direction)   |
|  | Trough Cross-Bedding (Ebb-directed)                      |    | Runzel marks  |
|  | Trough Cross-Bedding (Flood-directed)                    |   | Mudcracks   |
|  | Trough Cross-Bedding with Silt/Mud Drapes (Ebb-directed) |  | Directional Tool Marks  |
|   |  |  | Large diameter <u>Archanodon</u> -like burrows with spreites                                |
|   |  |  | <u>Arenicolites</u> burrows   |
|   |  |  | Vertical <u>Skolithos</u> burrows   |
|   |  |  | <u>Cupularostrum</u> (Rhynchonellid Brachiopod)   |
|   |  |  | <u>Barroisella</u> (Inarticulate Brachiopod)  |
|   |  |  | <u>Bothriolepis</u> Fish Fragments  |
|   |  |  | In Loco Plants (Minimally Transported)  |
|   |  |  | Transported Plant Fragments   |
|   |  |  | Wood, Branch, and Bark Fragments (Transported)  |
|   |  |  | Root Traces and Rhizoliths  |
|   |  |  | <u>Drapanophycus</u> plant remains  |

Carbonized stems, branches and logs are common along erosional bases and are locally abundant throughout the subfacies. Concentrations of *Cupularostrum* and very rare crinoid ossicles occur at and just above erosional bases of two sequences of subfacies  $T_{xb-ti}$  at outcrop 53. No other invertebrates occur in subfacies  $T_{xb-ti}$ . Large caliber, *Archanodon*-like burrows are relatively uncommon, but occur at outcrops 43, 45, and 47. No other trace fossils have been observed.

Subfacies  $T_{xb-ti}$  abruptly overlies paleosols  $P_{ao}$ ,  $P_{bo}$ ,  $P_p$ ,  $P_{aa}$ , and  $P_{gp}$  at several outcrops (Fig. 6). This contact is sharp, planar, and non-erosive (e.g., the A horizon in paleosols are preserved), but is locally minimally-erosive with channel forms. The upper contact of  $T_{xb-ti}$  is seen only at outcrop 53, where it is gradationally overlain by paleosol  $P_{gp}$ . At outcrop 53, both contacts are exposed and subfacies  $T_{xb-ti}$  is about 4-5 meters thick. Regional correlation of closely spaced outcrops indicates that subfacies  $T_{xb-ti}$  may reach thicknesses of 5-10 meters. Subfacies  $T_{xb-ti}$  is laterally traceable for approximately 3.5 miles from the southeast at outcrop 53 in Sidney Center to the west at outcrop 47 in East Masonville.

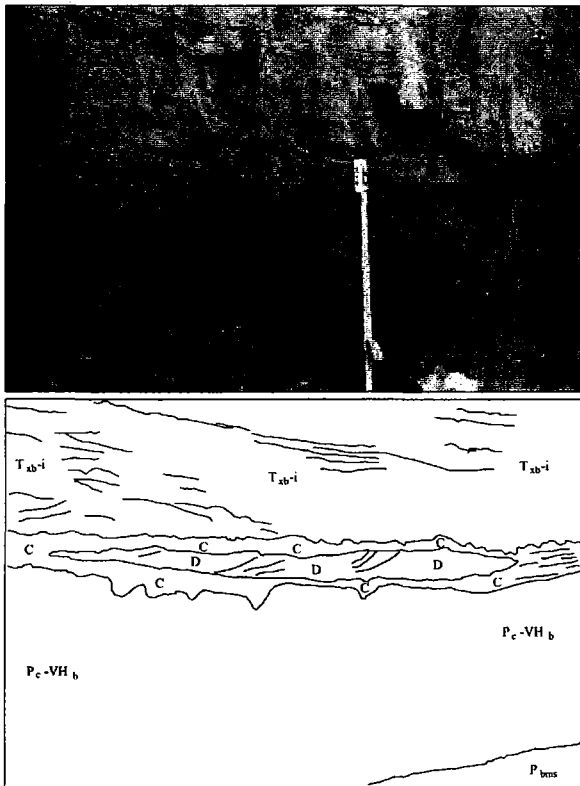


Fig. 6 Field photo and trace showing trough cross-bedded, tidally-influenced facies overlying paleosol (C and  $VH_b$  horizons). Note relict dune (D) within C horizon.

Sedimentary structures in subfacies  $T_{xb-ti}$ , such as sigmoidal trough cross-stratification, asymmetrical ripples migrating up the avalanche face of trough cross-beds, and reformed ripples with small-scale herringbone cross-lamination indicate mild tidal activity. The dominance of a northwest paleoflow is consistent with a very strong to strong asymmetric, ebb-directed tide (de Mowbray and Visser, 1984; Nio and Yang, 1991). The subordinate flood current was weak, since reactivation surfaces, flood-directed ripples, and mud drapes are rare. The presence of *Archanodon*-like burrows in subfacies  $T_{xb-ti}$  further supports the notion that these organisms occupied brackish water settings.

The basal erosion surfaces of subfacies  $T_{xb-ti}$  are inferred to be sequence-bounding unconformities, since subfacies  $T_{xb-ti}$  invariably overlies subaerially exposed lowstand surfaces comprising paleosols and other nonmarine environments. The differential erosion into underlying units observed at the basal contact of subfacies  $T_{xb-ti}$  reveals important clues to the nature of the marine transgressive surface. When traced laterally, the basal contact lacks significant relief indicating gentle onlap of marine waters into a series of antecedent coastal-plain vacuities such as paleosol-bearing interfluves and coastal drainage networks (i.e., drowned valleys and embayed coastline). Where the basal contact is sharp, planar, and non-erosive paleosol-bearing interfluves were apparently drowned in place. Where the contact becomes locally erosive with channel forms up to 0.5-0.75 m deep, the transgressing shoreface preferentially scoured antecedent fluvial and distributary channels of the paleovalley. The resultant erosion surface is therefore characterized by little or no microtopography on the paleolandscape flooding surface and lack of coastal plain incision. Similar marine transgressive surfaces, resulting in drowned, broad shallow valleys or embayed coastlines, have been described by Abbott (1998), Ricketts (1991), and Okazaki and Masuda (1995). No incised valley fill systems, and lowstand systems tract (LST) fluvial deposits and transgressive systems tract (TST) marine deposits associated with them, have been identified in the study area.

The basal erosion surfaces of subfacies  $T_{xb-ti}$  represent ravinement surfaces produced by transgressive flooding. Flute casts on the contact and small-scale trough cross-stratified dune bedforms immediately overlying the contact are oriented consistently to the southeast. This is not the orientation that would be expected if subfacies  $T_{xb-ti}$  were fluvial. Instead, the paleoflow data indicate onshore-directed (flood) tidal currents. The transgressive successions were deposited within and seaward of the erosional shoreface, starting with fields of shell-bearing gravel lags at the basal contact. These shell-bearing gravel lags are interpreted as winnowed and transported lag deposits of LST coastal plain and TST paralic facies that were consumed by erosion at the transgressing shoreface and, at least in part, redeposited offshore. Similar shell-bearing gravel lags are documented by Abbott (1998).

#### **Planar-Bedded Multistory Sandbody Facies (Facies $P_{bms}$ )**

The planar-bedded multistory sandbody facies ( $P_{bms}$ ) consists of moderately to well sorted, very fine- to fine-grained sublitharenites (Fig. 7). Planar beds are horizontal or subhorizontal. Horizontal beds locally attain a thickness of 7 meters. Bedding plane partings average 2 to 6 centimeters in thickness. Internal lamination is subtle and detectable only in thin section and polished slabs. Inversely graded laminae are rare. Upper stage planar laminations are abundant. Structures such as parting lineation and aligned carbonaceous plant fragments show a dominantly NW-SE paleocurrent direction at all localities. Vertical, hematitic burrows of *Skolithos* occur at several locations. Abundant inarticulate brachiopods of *Barroisella campbelli*(?) and *in loco* (i.e., minimally transported) delicate plant leaves and stems occur as a lag on the top surface of facies  $P_{bms}$  at two outcrops. *Barroisella campbelli* (?) occurs with root casts, carbonate-filled and open void rhizoliths, and sandy rhizocretions at outcrops 24 and 25.

Horizontal planar beds are locally replaced laterally by subhorizontal, planar-tabular cross-strata and channel forms measuring 2-10 m wide and 1-2 m thick. Planar-tabular cross-strata are inclined less than 5 degrees and measure tens of meters in length and 1-4 m thick. Dip directions of these cross-strata are predominantly west to northwest. Channel forms are infilled with cross-bedded sandstone and occur near the top 1-3 meters of facies  $P_{bms}$ . Reactivation surfaces with flat shale intraclasts also display valves of *Cupularostrum* and *Platyrachella*, and fragments of



the fish *Bothriolepis*, a fauna indicative of brackish conditions (Thayer, 1974). Highly abraded, arthropod fragments (trilobites or ostracods?) occur locally, as do flaser and wavy bedding with sand-filled mudcracks. Some bedding planes display trains of straight crested, asymmetrical ripples which have rounded to flat, planed-off crests. Other structures include symmetrical ripples, "ladder-back" ripples, and runzel marks.

Bedding planes are commonly strewn with carbonized and pyritized plant fragments, small branches, bark and rare logs (Fig. 8). Plant-rich strata commonly have yellow limonitic and reddish-orange hematitic stains from oxidation of pyrite. Discrete beds of intraformational conglomerate, consisting of flat shale clasts and weathered calcrete clasts occur on some bedding surfaces as well. These conglomerate beds occur more commonly as non-erosive sheets, and less commonly as low and broad erosive reactivation scour depressions measuring several meters wide and long and less than 10 centimeters deep, and channel lags measuring up to 1 meter deep. These scour depressions commonly contain mud drapes and abundant burrows of *Skolithos*(?), *Diplocraterion*(?), and *Paleophycus*(?). Large caliber, *Archanodon*-like vertical burrows are rare, occurring only at only two outcrops. Some localities display *Cruziana* trails, crescent-shaped burrows, *Rhizocorallium* (?), and "figure-8" burrows.

Facies  $P_{bms}$  is typically interbedded with the heterolithic-bedded facies ( $H_b$ ). Vertical exposure of  $P_{bms}$  and  $H_b$  stacks is limited, so the contact of facies  $P_{bms}$  with the underlying facies  $T_{xb}$  is rarely exposed. Observed thicknesses of available outcrop exposures of  $P_{bms}$  and  $H_b$  stacks range from 6-15 meters (1-2 complete stacks present) up to a maximum of 25 meters (4 complete stacks present) in the continuous exposure at Sidney Mountain quarry (outcrop 6).

Facies  $P_{bms}$  is interpreted as upper-flow-regime (UFR) sand flat deposits accompanied by a drainage network of tidal channels and tidal point bars within a central, seaward-flaring estuarine-like funnel of a tidally-influenced delta. Sandy tidal flats occupied the margins of the estuarine-like funnel. The UFR sand flats are represented by the thick accumulations of horizontal planar-laminated beds. The thickness of horizontal planar-laminated beds (i.e., up to 11 meters at outcrop 43) is not surprising, since they were deposited in the zone of turbidity maxima, where tidal and sediment-laden fluvial currents intersected. The tidal channels are recorded as channel forms infilled with cross-bedded sandstone, whereas the tidal point bars are recorded as subhorizontal, planar-tabular cross-strata interpreted as lateral accretion surfaces. Sandy tidal flats are recorded by abundant tidal sedimentary structures found capping facies  $P_{bms}$ .

The sandstone bodies of facies  $P_{bms}$  are organized into erosional-based storeys with lateral-accretion bedding that are similar to other deposits interpreted as tidal bars in laterally migrating estuarine channels. Paleocurrent measurements indicate a dominant southeast to northwest flow, which coincides with the axis of the seaward estuarine mouth of the deltas. The paleoshoreline is oriented normal to this, in a northeast-southwest direction. This orientation is in agreement with past paleogeographic reconstructions of the Catskill Sea (Haeckel and Whitzke, 1984; Barrel, 1913, 1914). Similar ancient examples from other parts of the Catskill Clastic Wedge are reported by Slingerland and Loule (1988), Bridge and Droser (1985), Halperin and Bridge (1988) and, Bridge and Willis (1994). Griffing and others (2000) report similar facies from Gaspé and Dalrymple and others (1992) cite modern examples which appear analogous.

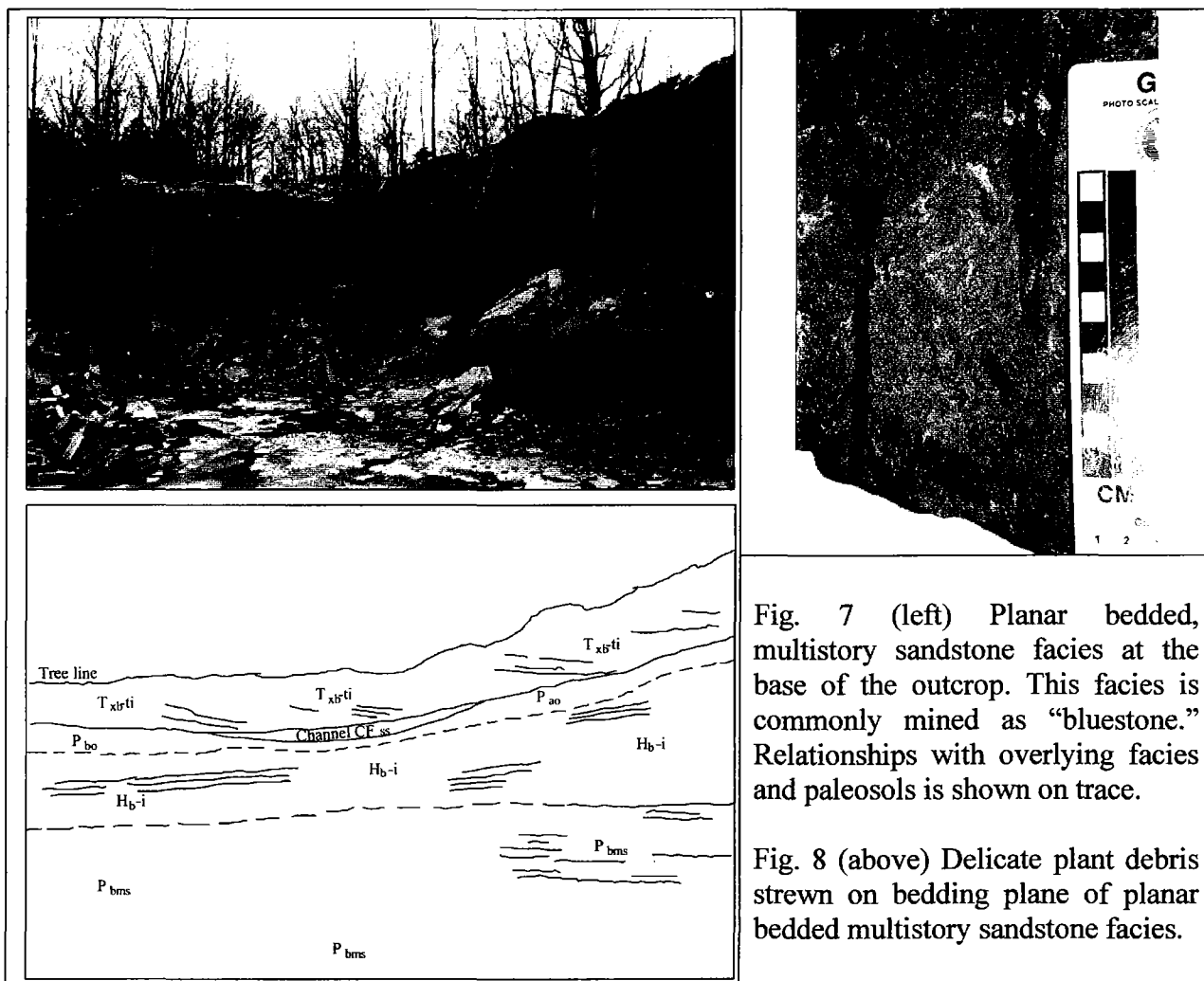


Fig. 7 (left) Planar bedded, multistory sandstone facies at the base of the outcrop. This facies is commonly mined as “bluestone.” Relationships with overlying facies and paleosols is shown on trace.

Fig. 8 (above) Delicate plant debris strewn on bedding plane of planar bedded multistory sandstone facies.

### Heterolithic-Bedded Facies (Facies $H_b$ – with Subfacies $H_b$ -I and $H_b$ -SL)

The heterolithic-bedded facies ( $H_b$ ) is comprised of two subfacies: inclined heterolithic bedded subfacies ( $H_b$ -I) (Fig. 7) and streaky-laminated heterolithic subfacies ( $H_b$ -SL). Facies  $H_b$  is invariably found interbedded as repetitively-stacked sequences with facies  $P_{bms}$ . Contacts with underlying facies  $P_{bms}$  are sharp and planar. This contact is locally erosional with facies  $H_b$  filling channel forms that are incised up to 4 meters into the underlying facies  $P_{bms}$ . Facies  $H_b$  is either gradationally overlain by paleosols or sharply overlain by facies  $T_{xb}$  and  $P_{bms}$ . The contact between  $H_b$  and  $T_{xb}$  is sharp, planar, and non-erosive, but is locally erosive with channel forms that incise up to 0.5-1 meter into the top of facies  $H_b$ . Facies  $H_b$  ranges in thickness from 1.2 to 6.8 meters at individual outcrops.

### *Heterolithic bedded subfacies (H<sub>b</sub>-I)*

Subfacies H<sub>b</sub>-I is characterized by inclined heterolithic stratification (IHS) and inclined stratification (IS) as defined by Thomas and others (1987) and horizontal planar rhythmic bedding (i.e., flat heterolithic bedding or FHB), which bear some similarities to bedding described by Van den Berg (1981), Smith (1985), and Coleman, Gagliano, and Webb (1964). IHS and FHB are commonly intercalated, but one form may dominate at individual outcrops. Average inclination of IHS in the study area is approximately 7° with an overall range from 3-25°. Average inclination of IHS at outcrops where FHB predominates is 3° or less. These stratification types consist of distinct cm-scale coarse- and fine-grained members that together constitute repetitive coarse-fine couplets.

Coarse members of couplets are predominantly homolithic, but may also contain one or more of the following: pebble- to granule-grade intraformational and extraformational conglomerates, carbonized wood and bark debris, and fish plates, bone fragments, and teeth of *Bothriolepis* and possibly *Eustenopteron*. Sandy members are generally consistent in thickness (i.e., 3-5 cm), but have an overall range in thickness from approximately 1-15 cm. Fine members of couplets are invariably heterolithic. Silty fine members are typically comprised of 1-3 cm thick layers, but mm-scale layers are also found. Lithologies are very fine to fine sublitharenite capped by sandy siltstone, siltstone, or mudstone. Fine members may also be intricately interlaminated (1-2 mm scale) with siltstone, mudstone, and carbonaceous plant detritus or, *in loco* plant fragments. These interlaminated fine members resemble horizontal pinstripes similar to the streaky lamination of subfacies H<sub>b</sub>-SL.

Sedimentary structures common in coarse members include planar lamination and small-scale asymmetrical, symmetrical, and climbing ripple cross-lamination. Small- to medium-scale trough and planar cross-lamination is less common. Load and flute casts are common on the sole of sandy coarse members. Fine members are typically parallel-laminated, small-scale symmetrical and asymmetrical ripple cross-laminated, and lenticular-, flaser-, and wavy-bedded. Runzel marks are common on top bedding surfaces of coarse and fine members and are found in close proximity to mudcracks where they occur. Mudcracks occur throughout subfacies H<sub>b</sub>-I, but are rare and closely associated with the top of channel fills. Asymmetrical ripples with superimposed mudcracks and ladder-back ripples are fairly common, and tend to occur near the top of subfacies H<sub>b</sub>-I and within channel fills. Evidence for bidirectional paleoflow is rare, and is limited to ladder-back ripples.

Carbonized, pyritized, and chalcopyritized plant, wood, and bark fragments are commonly strewn on bedding planes in subfacies H<sub>b</sub>-I. Rare carbonized impressions of shrub-like plants have also been found. These plant and woody fragments are typically oriented in a west-northwest to east-southeast paleoflow direction. Body fossils of invertebrates are extremely rare, consisting of sparse occurrences of *Barroisella campbelli*. Vertebrates are restricted to fish plates, bone fragments, and teeth of *Bothriolepis* and possibly *Eusthenopteron* found at the base of sandy IHS coarse members. Concentrations of fish bones average 3-8 centimeters in thickness (Figs. 4, 5). Gordon (1988) and Blicek (1982, 1985) described similar fish bone beds in other

Devonian marginal marine deposits. Two fish bone beds are laterally continuous for at least 100-150 meters. The lowermost bone bed at outcrops 35 and 36 changes laterally into a granule- and pebble-grade conglomeratic sandstone approximately 5-10 centimeters thick and 10-20 meters wide. The layer contains weathered siltstone, shale, and calcrete clasts. The conglomeratic sandstone is in turn traced lateral into a wedge-shaped point bar sandstone. This point bar sandstone is 1-1.5 meters thick and contains long sigmoidal-shaped inclined strata. Abundant, U-shaped *Arenicolites* burrows are found in the basal 30 cm of the sandstone. Two, articulated shells of *Archanodon* were found nearby in the float.

An overall vertical fining upward trend is evident at all IHS-bearing outcrops. Lateral fining away from channel forms (i.e., predominantly IHS-filled) into predominantly FHB is observed at some outcrops. The northeast portion of outcrop 43 contains lateral fining into a large abandoned channel (i.e., 4 meters deep and 20+ meters wide), which was subsequently plugged with predominantly IHS fine members

Subfacies H<sub>b</sub>-I is interpreted as IHS-filled, sinuous, tidally-influenced channels, point bars, and associated overbank tidal flats occupying the middle to upper-portion of the estuarine-like funnel. Based on the conformable nature of subfacies H<sub>b</sub>-I with the underlying facies P<sub>bms</sub>, (i.e., tidally-influenced) and the sedimentary structures contained within it, a tidal origin for subfacies H<sub>b</sub>-I is a reasonable interpretation. Although evidence for bidirectional paleoflow is rare or absent, several structures found in subfacies H<sub>b</sub>-I meet the criteria required for differentiation of tidally-influenced river point bar deposits from wholly fluvial deposits as summarized by Thomas and others (1987):

1. Brackish cosmone-bearing fish and lingulid brachiopods are present.
2. Rhythmically interbedded couplets of sandstone, siltstone, and mudstone (i.e., FHB) constitute much of the point bar deposits.
3. Upper point bar sequences contain wave and current-ripple structures plus linsen and flaser bedding.
4. Bimodal paleoflow indicators are limited to ladderback ripples. However, sandy couplets contain an abundance of other tidal structures.
5. Some point bar sequences are characterized by intense bioturbation.

The abundance of biogenic structures (e.g., *Arenicolites* and *Skolithos*) is regarded as the single-most useful and widely applicable feature for distinguishing tidal creek point bar deposits from those of fluvial origin (Thomas and others, 1987). The low trace fossil diversity and rarity of biogenic activity in areas outside of concentrations of trace fossils suggests stressed environmental conditions related to brackish salinities and/or high concentrations of suspended sediment.

### ***STREAKY-LAMINATED HETEROLITHIC BEDDED SUBFACIES H<sub>B</sub>-SL***

Subfacies H<sub>b</sub>-SL occurs as two distinct units in the Sidney Mountain Quarry (outcrop 6, Fig. 5). Subfacies H<sub>b</sub>-SL comprises a spectrum of small-scale heterolithic structures ranging from streaky-laminated sandy siltstone to sandy lenticules and wavy-bedding with discontinuous ripple cross-lamination to ripple cross-laminated fine sandstone with mud flasers (sensu Reineck

and Wunderlich, 1968). Streaky-laminated sandy siltstone overlies the channel fills and the sharp, planar contact above facies  $P_{bms}$ . The streaky-lamination is characterized by pinstripe lamination as described above. Rarely, sub-millimeter thick lenses of fine sand and sandy-silt drapes are arranged as small-scale climbing oscillatory ripples similar to those described by Kvale and Archer (1991). Similar laminations have been described by Abbott (1998), Mutti and others (1985) and Homewood and Allen (1981).

No invertebrates have been identified in subfacies  $H_b$ -SL. However, concentrations of carbonized plant debris and tree bark are strewn on top bedding surfaces of sandstone lenses in the lowermost 2 meters of the unit. Trace fossils are restricted to locally abundant, short, vertical burrows resembling *Skolithos* and epichnial sinuous trails of *Planolites*.

A combined shallow marine setting of alternating storm-dominated and tidally-influenced conditions appears to be a reasonable origin for subfacies  $H_b$ -SL. Fine-grained prodelta sediments prograded in advance of sandier, tide-influenced deltaic sediments, which likely were affected by landward-directed storm currents, bidirectional (ebb-dominated) tidal currents, and possibly seaward-directed river currents during flood events. Several subaerial features prominent in tidal-flat and estuarine facies models are missing from subfacies  $H_b$ -SL. Bedding and sedimentary structures are more consistent with deposition in a shallow marine, inner-shelf, pro-delta setting. Similar deposits described by Homewood and Allen (1981) were interpreted as representing periods of alternating strong and slack currents. Homewood and Allen (1981) also suggested that millimeter- to decimeter-scale alternations in bedding thickness in facies such as subfacies  $H_b$ -SL can result from fluctuations of wave base caused by spring-neap tidal cyclicity. Alternatively, heterolithic beds such as flaser bedding and wavy-bedded sheet sandstones (i.e., middle portion of subfacies  $H_b$ -SL) may be produced solely by storms alternating with quiescent periods (McCave, 1970).

## Catskill Magnafacies - Paleosols

### *Introduction*

Five types of paleosols have been recognized in the study area. These paleosols overlie the unnamed formation of the Cattaraugus Magnafacies (i.e., caps both facies  $P_{bms}$  and  $H_b$ ). A total of 7 paleosol horizons have been identified in the study area at outcrops 6 (2 horizons), 43, 51, and 53 (3 horizons) (See Figs. 4, 5). Suspected pedogenically-influenced horizons occur at outcrops 24, 34, 35, and 47. Descriptions and interpretations of these paleosols are presented below. A more complete treatment of this topic is forthcoming by Bishuk, Hairabedian and Ebert (in preparation).

### *Classification of Paleosols*

**In an attempt to provide a non-genetic, descriptive account of the paleosols, primary classification will follow the scheme proposed by Mack and others (1993). Secondary names will be assigned using the Duchaufour (1982) classification since it focuses on pedogenic process rather than modern soil properties and recognizes intergradations between classes of paleosols. Tertiary names are assigned to each paleosol in accordance with Retallack (1983) and Soil Survey Staff (1996). Interpreted names of Sonyea paleosols are summarized in Table 1.**

**Table 1: Classification of paleosols in outcrops of the Sonyea Group according to various systems.**

| <b>OUTCROP NUMBER</b> | <b>PALEOSOL SYMBOL</b> | <b>LOCATION</b>  | <b>Mack, James, and Monger Classification (1993)</b> | <b>Duchaufour Classification (1982)</b> | <b>Retallack Classification (1983)</b> | <b>Soil Survey Staff Modern Soil Classification (1996)</b> |
|-----------------------|------------------------|--|--|---|--|--|
| 6                     | P <sub>gp</sub>        | Sidney Mountain Quarry, Delaware County Route 4, Sidney                        | Gleyed Protosol                                      | Slightly Developed Soil                 | Gleyed Inceptisol                      | Aquic Inceptisol   |
| 6                     | P <sub>aa</sub>        | Sidney Mountain Quarry, Delaware County Route 4, Sidney                        | Albic Argillisol                                     | No direct equivalent                    | Albic Argillaceous Inceptisol          | Eutrochrept Inceptisol                                     |
| 43                    | P <sub>ao</sub>        | Skytop Lane Quarry, Skytop Lane, Sidney Center                                 | Plinthite-Bearing Argillic Oxisol                    | Plinthite-Bearing Ferruginous Soil      | Plinthite-Bearing Argillic Oxisol      | Plinthic Acrustox Oxisol                                   |
| 43                    | P <sub>bo</sub>        | Skytop Lane Quarry, Skytop Lane, Sidney Center                                 | Brecciated Oxisol Hardpan                            | Brecciated Ferruginous Soil             | Brecciated Oxisol Hardpan              | Lithic Acrustox Oxisol                                     |
| 43                    | P <sub>p</sub>         | Skytop Lane Quarry, Skytop Lane, Sidney Center                                 | Caliche-bearing Protosol                             | Slightly Developed Soil                 | Caliche-bearing Entisol                | Caliche-bearing Entisol                                    |
| 50*                   | P <sub>aa</sub>        | Unnamed Quarry, Intersection of Olmstead Road and Wilcox Road, East Masonville | Albic Argillisol                                     | No direct equivalent                    | Argillaceous Inceptisol                | Eutrochrept Inceptisol                                     |
| 51                    | P <sub>gp</sub>        | Sheetz Quarry, Cummings Road, East Masonville                                  | Gleyed Protosol                                      | Slightly Developed Soil                 | Gleyed Inceptisol                      | Aquic Inceptisol   |
| 53                    | P <sub>gp</sub>        | Delaware County Route 27, Sidney Center  | Gleyed Protosol                                      | Slightly Developed Soil                 | Gleyed Inceptisol                      | Aquic Inceptisol   |
| 53                    | P <sub>aa</sub>        | Delaware County Route 27, Sidney Center  | Albic Argillisol                                     | No direct equivalent                    | Albic Argillaceous Inceptisol          | Eutrochrept Inceptisol                                     |

**NOTES:**

\* = More research is necessary at this outcrop for positive identification of paleosol; Paleosol names provided are preliminary.

### **Non-Brecciated Argillic Oxisol and Related Units – Paleosols $P_{ao}$ , $P_{bo}$ , and $P_p$ and Facies $CF_{ss}$ at Outcrop 43, Skytop Lane Quarry, Sidney Center**

Many of the paleosols described in Table 1 can be studied at a single outcrop near Sidney Center – the Skytop Lane Quarry (outcrop 43; Figs. 5, 7). Channels that incise paleosols are present as are lateral changes in paleosol type. This outcrop will be the centerpiece of the field trip.

#### *Channel Form Silty Sandstone Facies ( $CF_{ss}$ )*

Facies  $CF_{ss}$  occurs as laterally discontinuous, concave-upward channel forms measuring up to 10 meters wide, which incise up to 0.5 meters into paleosol  $P_{ao}$ . Two channels of  $CF_{ss}$  are present at outcrop 43 (Fig. 7). Alternating beds of siltstone and silty, very fine sandstone fill the channels. Trough crossbeds with dip angles averaging 30-35 degrees are the dominant sedimentary structure within the center of the channels. These crossbeds are commonly contorted by soft sediment deformation. Near the edges of the channels, low-angle laminations generally approximate the concave-upward shape of the channel. Edges of channels pinch out to approximately 10-15 cm thick and grades into horizons  $A_k/A_{ao}$  of paleosol  $P_{ao}$ . No body fossils, trace fossils, or evidence of pedogenesis occur in facies  $CF_{ss}$ .

Facies  $CF_{ss}$  represents ephemeral wash channels incised into the paleosols of stable interfluvies in a semi-arid environment. Soft sediment deformation in the channels suggests rapid deposition during floods. There is no evidence of tidal-influence in facies  $CF_{ss}$ . The geometry of facies  $CF_{ss}$  indicates these wash channels were stable in position for a prolonged period of time. Lateral bank erosion was probably inhibited or completely halted by the resistant nature of parts of the adjacent paleosols. Downcutting of these channels was hampered by low seasonal discharge and the erosional resistance of the paleosols. These ephemeral wash channels offer insight into the nature of the paleocoastal landscape at a major unconformable surface. They also provide evidence that river avulsion caused a drastic decrease in discharge to the paleocoastline, which set up prolonged conditions for the development of paleosols in the interfluvies (Boswell and Donaldson, 1988).

#### *Paleosol ( $P_{ao}$ )*

A moderately well developed, non-brecciated, argillic oxisol (paleosol  $P_{ao}$ ) caps facies  $H_b$  at outcrop 43 (Fig. 9). Paleosol  $P_{ao}$  gradationally overlies facies  $H_b$  and is sharply overlain by facies  $T_{xb-ti}$ . The contact of paleosol  $P_{ao}$  and facies  $T_{xb}$  is predominantly non-erosional, preserving underlying delicate pedogenic features, but is erosional locally with channels incised up to 0.5-1 meter into the paleosol. Rarely, the channel-form silty sandstone facies (facies  $CF_{ss}$ ) occurs between paleosol  $P_{ao}$  and facies  $T_{xb}$ .

Paleosol  $P_{ao}$  ranges in thickness from 75 cm to 95 cm. Four pedogenic horizons exist - a laminar calcrete crust ( $A_k$ ), a plinthite-bearing melanic epipedon ( $A_{ao}$ ), an argillic horizon ( $B_t$ ), and a nodule-bearing horizon with vestigial bedding ( $C_c$ ). Paleosol  $P_{ao}$  can be traced laterally approximately 100 meters across outcrop 43, but appears to change character into another type of paleosol (paleosol  $P_p$ ) to the northeast across the remaining 200-300 meters of the outcrop.

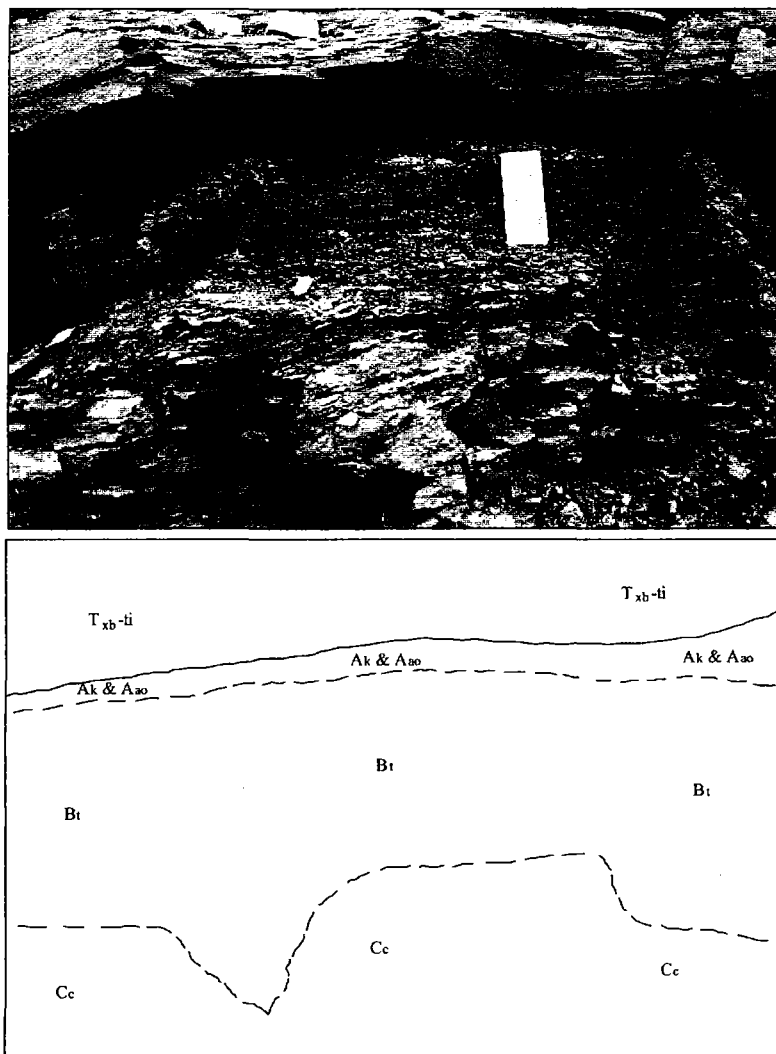


Fig. 9 Field photo and tracing argillic oxisol (paleosol  $P_{ao}$ ) horizonation. Paleosol is abruptly overlain by the tidally-influenced trough cross-bedded facies (outcrop 43).

#### $A_k$ Horizon – Laminar Calcareous Crust

The  $A_k$  horizon is a continuous, 1-5 mm thick, structureless, white to light gray crust (Munsell colors = GLEY1 8/N) that blankets the  $A_a$  horizon, similar to that described by Lattman (1973). The crust is powdery and weakly calcareous. This crust is laterally continuous for a minimum of 20 meters, extending to the edge of the outcrop. Locally, crude microlaminations are discernable. Reaction to hydrochloric acid is weak to moderate. In some places, the layer is replaced laterally by light gray (5Y 7/2), highly plastic, calcareous clay.



### A<sub>ao</sub> Horizon – Plinthite-Bearing Melanized Epipedon

The A<sub>ao</sub> horizon is a dark gray/black to dark reddish gray hardpan cemented by amorphous iron oxide with localized zones of a humus-poor, iron and organic matter complex, which qualifies it as a plinthite (formerly known laterite) layer (Retallack, 2001). Bitumen is abundant on the top surface of the hardpan as well as sparse unidentifiable plants (i.e., humus-poor zone). The hardpan ranges in thickness from 5-10 mm and is laterally continuous for at least 25 meters to the edge of the outcrop. The hardpan is always associated with the A<sub>k</sub> horizon.

The basal portion of the A<sub>ao</sub> horizon ranges in thickness from 4 to 10 cm. It is characterized by a distinct darkening (GLE Y2 3/5 – dark bluish gray) caused by the accumulation of abundant organic matter (bitumen and sparse unidentifiable plant remains) with lesser concentrations of amorphous iron oxide. Numerous glaeboles occur in the form of sesquioxidic, ferruginous nodules, consisting of amorphous iron oxide. The ferruginous nodules show an increase in size upward from the base of A<sub>ao</sub> to the base of the overlying plinthite. These elliptical nodules are so abundant that they form a peppered fabric throughout the thin sections. The nodules commonly disrupt microlaminae and also occur as thin coatings of amorphous iron oxide on microlaminae. The darkest layers of iron oxide anastomose and bifurcate. In the field, the degree of mottling is 30-40 %, with subordinate colors ranging from light gray (5Y 7/2), yellowish brown (10YR 5/6), and pale olive (5Y 6/3). Concentrations of plant stems and leaf fronds occur as a mat-like fabric, which resembles sparse leaf litter.

Crude vestigial laminations of alternating silty, very fine sandstone and medium to coarse sandstone lenses are present. Micro-scale planar laminations and current ripple cross-laminations are readily seen with the naked eye on thin section mounts. The A<sub>ao</sub> horizon grades laterally into silty sandstone beds of facies CF<sub>ss</sub>, thereby establishing a close genetic association between A<sub>ao</sub> and CF<sub>ss</sub>. The original bedding is overprinted by a platy ped fabric. Concentrations of bitumen commonly coat the parting surfaces of peds. No accumulation of clay was observed in the A<sub>ao</sub> horizon.

### B<sub>t</sub> Horizon – Argillic Horizon

The B<sub>t</sub> horizon is characterized by subangular to angular blocky peds composed of siltstone and very fine sandy siltstone. The dominant colors in the B<sub>t</sub> horizon are light greenish gray (GLE Y1 8/10Y) and white (GLE Y1 8/N). Degree of color mottling is 20-25%, with subordinate colors ranging from reddish brown (5YR 5/8 and 5YR 6/8) and dark bluish gray (GLE Y2 3/5PB). The B<sub>t</sub> horizon thickness ranges from approximately 50 to 90 cm and is laterally continuous for approximately 100 meters. Ped structures are commonly infilled with clay argillans and coated with quasiferrans (i.e., two forms of illuviation cutans). The composition of these illuviation cutans is likely illite swelling clay and cryptocrystalline hematite or goethite, respectively. Concentrations of clay argillan material are greatest at the top of the B<sub>t</sub> horizon and decreases downward. Locally, clay argillans are absent and ped surfaces are stained by sesquioxidic quasiferrans. Pedogenic slickensides (i.e., stress cutans) are rare, but locally occur in the mid- to lower-portion of the B<sub>t</sub> horizon. Glaeboles occur in the form of small goethite concretions and homogeneous caliche nodules measuring 0.2-2 cm in diameter, but are relatively uncommon. Roots, rhizoliths, rhizocretions, and pedotubules are absent. Rare lenses of an amorphous iron oxide and darkened organic matter complex, similar in composition and appearance to the A<sub>ao</sub>

horizon, occur within the uppermost 10 cm of the B<sub>t</sub> horizon. These lenses measure 3-5 cm thick and up to approximately 50 cm long.

#### C<sub>c</sub> Horizon – Nodular-Bearing Horizon with Vestigial Bedding

The C<sub>c</sub> horizon is characterized by vestigial heterolithic bedding of facies H<sub>b</sub>. Pedogenic features are restricted to glaebules, which occur in the form of small (0.2-3 cm, but average <1 cm) caliche nodules. These nodules are rare, but tend to occur in clusters (i.e., possible rhizcretion accumulations). No roots, rhizoliths, rhizcretions, or pedotubules were observed. *Skolithos* and *Arenicolites* burrows and *in loco* plant fragments are exceedingly common in vestigial fine sandstone interbeds.

Paleosol P<sub>ao</sub> is interpreted as a moderately-well developed argillic oxisol (Table 1). Pedogenic features present in paleosol P<sub>ao</sub> make a strong case that processes such as hydrolysis, hydration, dissolution, oxidation, leaching, and acidification were either intense, prolonged, or both. The advanced weathering of the oxisol profile is an indicator of great age, which can amount to tens of millions of years (Retallack, 2001). However, a more realistic estimate of the oxisol age is 20-400 thousand years, which is based on rates of modern hardpan development (Williams and Krause, 2000).

#### **OTHER PALEOSOLS AT OUTCROP 43 – PALEOSOL P<sub>BO</sub> AND PALEOSOL P<sub>P</sub>**

A brecciated oxisol (paleosol P<sub>bo</sub>) (Fig. 10) and protosol (paleosol P<sub>p</sub>) are found interdigitated with paleosol P<sub>ao</sub> in the Skytop Lane Quarry. Paleosols P<sub>bo</sub> and P<sub>p</sub> gradationally overlie facies H<sub>b-ti</sub> and are sharply overlain by facies T<sub>xb</sub>. Paleosols P<sub>bo</sub> and P<sub>p</sub> are generally 1 meter or less in thickness. Three pedogenic horizons exist in paleosol P<sub>bo</sub>, including a sesquioxenic-nodular, brecciated horizon (B<sub>co</sub>) with evidence of early lithification, an argillic horizon (B<sub>t</sub>), and a horizon which bears caliche nodules and vestigial bedding (C<sub>c</sub>), which are remnants of heterolithic bedding in facies H<sub>b</sub>. Lateral continuity of discontinuous sections of paleosol P<sub>bo</sub> is poor across outcrop 43, ranging from 5-25 meters.

Fig. 10 Brecciated oxisol (paleosol P<sub>bo</sub>) at STOP #1. See Figs. 4 and 7 for position of this paleosol. DNAG card on the paleosol provides scale.



Paleosol P<sub>bo</sub> is interpreted as a moderately-well developed brecciated oxisol hardpan (Table 1). Degree of pedogenesis in paleosol P<sub>bo</sub> is mild to moderate as compared to features present in paleosol P<sub>ao</sub>. The presence of numerous sub-rounded clasts of variable lithology (i.e., weathered calcrete, siltstone, and shale) indicates that some material was transported rather than *in situ* parent material. The laterally adjacent position of paleosol P<sub>bo</sub> in relation to facies CF<sub>ss</sub> is significant. It may indicate that paleosol P<sub>bo</sub> is a pedogenically-altered remnant of a point bar of the ephemeral wash channels of facies CF<sub>ss</sub>.

No horizonation is discernable in paleosol P<sub>p</sub>. It contains only a nodular-bearing horizon with vestigial bedding (C<sub>c</sub>). Paleosol P<sub>p</sub> is interpreted as a protosol, which was subjected to minor levels of pedogenesis. Similar to the C<sub>c</sub> horizon of paleosol P<sub>ao</sub>, pedogenesis is restricted to caliche nodules (i.e., glaebules). Paleosol P<sub>p</sub> is laterally proximal to a large channel form infilled with tidally-influenced sedimentary structures. It appears likely that frequent tidal flooding diminished or disrupted pedogenesis.

#### **Albic Argillisol—Paleosol P<sub>aa</sub>**

A moderately developed, albic argillisol (paleosol P<sub>aa</sub>) underlies facies T<sub>xb-ti</sub> at outcrop 53 in Sidney Center, and caps facies P<sub>bms</sub> at outcrop 6 (Sidney Mountain Quarry- STOP 3; Fig. 5). At outcrop 6, paleosol P<sub>aa</sub> sharply overlies facies P<sub>bms</sub>. A second occurrence of paleosol P<sub>aa</sub> at outcrop 6 is sharply overlain by marine deposits of facies S<sub>dg</sub> and H<sub>cs</sub>. The contact of paleosol P<sub>aa</sub> and facies P<sub>bms</sub> is minimally-erosional, preserving underlying E horizon features, but is locally erosional with channels incised up to 0.5-1 meter into the paleosol. The contact of paleosol P<sub>aa</sub> and facies S<sub>dg</sub> and H<sub>cs</sub>, occurring at the top of outcrop 6, appears to be minimally-



erosional as well, with localized relief up to 1 meter at the contact on an otherwise sharp and planar contact.

Paleosol  $P_{aa}$  ranges in thickness from 1.7 meters at outcrop 6 to 4.5 meters at outcrop 53. Pedogenic horization is weak. Recognized horizons include a leached albic horizon (E), an argillic horizon ( $B_t$ ), and a rooted sandstone horizon (C). Paleosol  $P_{aa}$  laterally replaces paleosol  $P_{gp}$  at outcrop 6, and occurs twice (i.e., 18-19.5 m and 22-24.5 m on Fig. 5).  $P_{aa}$  is distinguished from  $P_{gp}$  by the absence of relict bedding.

Paleosol  $P_{aa}$  supported mono-specific stands of plants further away from coastal waterways in comparison to paleosol  $P_{gp}$ , but was still influenced by infrequent flooding from them (i.e., likely 100-year and 500-year flood events). This is in marked contrast to paleosols  $P_{ao}$ ,  $P_{bo}$ , and  $P_p$ , which is best explained by differences in relative distances from active paleochannels.

Since the E horizon is directly overlain by marine deposits of facies  $P_{bms}$  at outcrop 6 and facies  $T_{xb-ti}$  at outcrop 53, it is possible that overprinting of pedogenic patterns by early marine physical and chemical processes (i.e., marine hydromorphism) occurred at these locations (Fig. 11). This occurrence bears similarities to features described by Driese, Mora, and Cotter (1993). Chemical analysis or X-ray diffraction is necessary to confirm the presence or absence of chemical signatures of marine hydromorphism in paleosol  $P_{aa}$ .



Fig. 11 Possible marine hydromorphism of paleosol at outcrop 53. Paleosol is overlain by marine sediments of the planar-bedded multistorey sandstone facies

#### Gleyed Protosol— Paleosol $P_{gp}$

A poorly developed, gleyed protosol (paleosol  $P_{gp}$ ) caps facies  $P_{bms}$  at outcrop 51 (Sheetz Quarry- STOP 2), outcrop 6 (Sidney Mountain Quarry- STOP 3), and outcrop 53. Paleosol  $P_{gp}$  sharply overlies facies  $P_{bms}$  and contains vestigial heterolithic bedding from facies  $H_b$ . Paleosol  $P_{gp}$  is sharply overlain by facies  $P_{bms}$  or sharply overlain by marine facies  $S_{dg}$  and  $H_{cs}$ . The contact of paleosol  $P_{gp}$  and facies  $P_{bms}$  is minimally-erosional, preserving underlying E horizon features, but is locally erosional with a channel incised up to 0.5-1 meter into the paleosol.

Paleosol  $P_{gp}$  ranges in thickness from 0.4 meters at outcrop 53 to 1.7 meters at outcrop 6. Paleosol  $P_{gp}$  can be traced to the lateral extents of each outcrop; several hundred meters at outcrops 6 and 51 and less than 50 meters at outcrop 53. At outcrop 6, paleosol  $P_{ap}$  occurs interdigitated with paleosol  $P_{gp}$ . Pedogenic horizonation is poor; therefore abundant primary sedimentary structures are preserved. Recognized horizons are limited to a leached albic horizon (E) and a pedogenically-influenced, vestigially bedded horizon ( $C_c$ ). An abrupt horizon of red-green mottling occurs at the contact of vestigial heterolithic bedding with facies  $P_{bms}$ , and generally decreases in intensity upward in paleosol  $P_{gp}$ . Ped fabrics are very weak to absent, exhibiting a cryptic platy framework. The most common features at outcrops 51 and 53 are drab root haloes, which consists of 0.5-5 mm diameter rhizoliths (i.e., filled and unfilled varieties) with boundaries of bluish gray to greenish gray haloes extending out into the paleosol matrix (See also Retallack, 2001). Mat-like rhizomatous networks occur in mudrocks between successive sandstone beds within vestigial heterolithic bedding. One possible ornamental leaf-like structure with associated node-like basal attachment, tentatively identified as *Drapanophycus*, has been collected from these horizons (Fig. 12). Rarely, large diameter tap roots measuring up to 0.5-1 cm are found, as are rare caliche nodules.



Fig. 12 Leaf-like structure tentatively identified as *Drapanophycus*.

The color mottling exhibited by this paleosol is a diagnostic feature of waterlogged soils and is a conspicuous feature of groundwater gley (Retallack, 1997). Mottling is also characteristic of frequent exposure to redox fluctuations (Driese and others, 1997). Based on the shallow, dominantly horizontal, mat-like fabric of root traces,  $P_{gp}$  is considered the product of groundwater gley associated with a shallow groundwater table along coastal waterways.

Paleosol  $P_{gp}$  is interpreted as a gleyed protosol, which supported mono-specific stands of flood-resistant plants on the banks of fresh-water channels on the distal alluvial plain at outcrop 51 and tidally-influenced deltaic channels at outcrops 6, 53 and possibly 51. Evidence for tidal-influence is clear at outcrops 6 and 53, suggesting that plant stands along the banks of delta channels and tidal flats were tolerant of brackish water conditions. The exact type of plants living in these coastal habitats is unknown, since *in situ* aerial portions of plants attached to roots are absent. However, the horizontal, mat-like rhizomatous fabric is reminiscent of fungi (i.e., calcified fungal mycelia or possibly lichen rhizomes) characterized by Driese and others (2001). Driese and others (1992) found that these fungi lived in moist, pedogenically modified deposits of

coastal mudflats and marshes. Alternatively, the rhizomatous fabric is similar to the reconstruction of the root system of the early lycophyte *Drapanophycus spinaeformis* (Driese and others 2001). The presence of *in situ* lycophytes is a possibility, since *Drapanophycus*-like structures were found. The identity of plants that produced the large diameter tap roots is unknown.

#### FACTORS AFFECTING SONYEA PEDOGENESIS

Paleosols of the Sonyea Group developed on coastal-margin interfluves and show a wide variability that can be explained by sea level position/fluctuation, drainage characteristics of adjacent and underlying sediments, an avulsion-controlled style of coastal-margin basin filling and tectonically induced subsidence. The sequence of events leading to Sonyea paleocoastline development is summarized in Fig. 13. Based on the consistent stratigraphic position of these paleosols (Figs. 4, 5), well developed paleosols record pedogenesis during periods of reduced or negative accommodation (i.e., slight base level fall). Repetition of the following sequences of processes is interpreted to have occurred: 1) sea-level rise (transgressive system tract), 2) peak sea-level rise slows to a pause (highstand system tract), 3) river avulsions and slight lowering of sea-level (falling stage to lowstand systems tracts) with coeval pedogenesis, 4) episodic subsidence associated with epeirogenic downflexure of the lithosphere (Quinlan and Beaumont, 1984), and 5) coastal drowning associated with renewed eustatic sea-level rise. Differences in paleosol characteristics are attributable to the time allowed for pedogenesis, during both autocyclic and high order (i.e., 4<sup>th</sup> and 5<sup>th</sup> order) eustatic fluctuations.

Changes in the Sonyea shoreline are characterized as “mixed shift” (Boswell and Donaldson 1988), where transgression is displayed in some parts of the shoreline, and regression along others. Mixed shifting is the result of changes in the distribution or supply of sediments as controlled by autocyclic processes such as river avulsions. Boswell and Donaldson (1988) determined that major rivers had relatively fixed positions in the Famennian portions of the Catskill clastic wedge, which created large, long-lived interfluve areas where tidal processes dominated. It is clear from this investigation that similar conditions existed in the study area during Frasnian time.

Sonyea paleosols formed near coastal sandflats and mudflats. During initial stages of soil development, the water table was probably close to the surface most of the time, and reducing conditions prevailed (e.g., paleosols P<sub>gp</sub>, P<sub>aa</sub>, and P<sub>p</sub>). The rate of overbank deposition of the tidally-influenced coastal waterways controlled the degree of pedogenesis. Some areas of the Sonyea coast experienced major river avulsions, starving some interfluves of sediment and creating new interfluves. Avulsion probably resulted in soils with lower water tables on newly created interfluves (i.e., paleosols P<sub>ao</sub>, and P<sub>bo</sub>), whereas areas that continued to receive sediment along active distal rivers remained waterlogged with high water table conditions (i.e., paleosols P<sub>gp</sub>, and P<sub>aa</sub>). The post-avulsion paleolandscape was probably a somewhat barren, semi-arid desert shrubland, possibly dominated by Aneurophytes, which also dominated marine bayfill/prodelta deposits and coastal-margin paleosols of the early Frasnian Oneonta Formation in the Catskill clastic wedge (Scheckler and others, 2000).

*In loco* plant remains of *Archeopteris* in paleosol P<sub>ao</sub> indicates that vegetated stream banks were localized upstream in fresher water habitats, such as river courses. These fresh-water habitats

must have been close to the brackish waterways. *Archeopteris* populations probably decreased rapidly after avulsions when reduced discharge and lower water tables caused habitat loss.

### *Comparison of Sonyea paleosols with other Devonian paleosols in the Appalachian Basin*

Interest in coastal-margin paleosols in the Devonian Catskill clastic wedge has increased over the past decade (Driese, Mora, and Cotter, 1993a and 1993b; Cotter, Mora, Fastovsky, and Driese, 1993; Driese, Mora, and Elick, 1997; Scheckler and others, 2000; Schieber, 1999; and Driese, Mora, Cotter, and Foreman, 1992). Paleosols reported show wide variability, but there is a general consensus of a semi-arid paleocoastal environment. These variations, as well as the oxisols and argillisols reported in this study, are consistent with a semi-arid paleoclimate. The coastal-margin gleysols and protosols reported by Cotter and others (1993) and Driese and others (1997) appear to be the closest matches to paleosols  $P_{gp}$ ,  $P_{aa}$ , and  $P_p$ . Paleosols similar to  $P_{ao}$  and  $P_{bo}$  have not been reported in other portions of the Catskill clastic wedge.

### Conclusions

Portions of the Frasnian Sonyea Group in northwestern Delaware County are comprised of the Cattaraugus Magnafacies, which represents nearshore and shoreline environments in the Catskill clastic wedge. These facies record a mosaic of tidally-influenced subenvironments in an overall estuarine setting. Facies that record tidally-influenced marginal marine environments succeed one another laterally and vertically and are complexly interleaved with fully marine facies and terrestrial paleosols. In many locations, paleosols developed directly on tidally-influenced facies and are overlain by facies of obvious marine or marginal marine origin. These stratigraphic relationships indicate a "mixed shift" shoreline in which progradation and retrogradation could be contemporaneous in adjacent areas. Avulsion and sediment supply were primary controls on which segments of the Sonyea shoreline prograded or retrograded during the overall Frasnian transgression. Episodes of tectonically induced subsidence and pulses of eustatic rise accelerated onlap, especially in areas from which sediment supply was diverted by avulsion. The newly recognized paleosols in the Sonyea Group have proven to be critical clues to understanding the dynamics of the Catskill shoreline.

Several types of paleosols developed on Sonyea coastal-margin interfluves under a semi-arid climatic regime. The formation of individual paleosols was terminated by marine inundation produced by decreased sediment supply via avulsion and/or high order eustatic fluctuations. Therefore, differences in paleosol characteristics resulted from differing durations of pedogenesis before inundation. The relative position of the water table and proximity to brackish channels were also important factors influencing the development of Sonyea paleosols. Some paleosols were vegetated, but it is likely that the flora was restricted to areas adjacent to channels where the water table was high. Paleosols with limited degrees of pedogenesis (e.g., gleysols and protosols) have been reported from similar paleoenvironments elsewhere in the Catskill clastic wedge. However, paleosols with greater degrees of pedogenesis such as the oxisols and argillisols in the Sonyea Group have not been reported elsewhere. These more advanced stages of pedogenesis may indicate a subtle increase in the aridity of the climate in the Late Devonian.



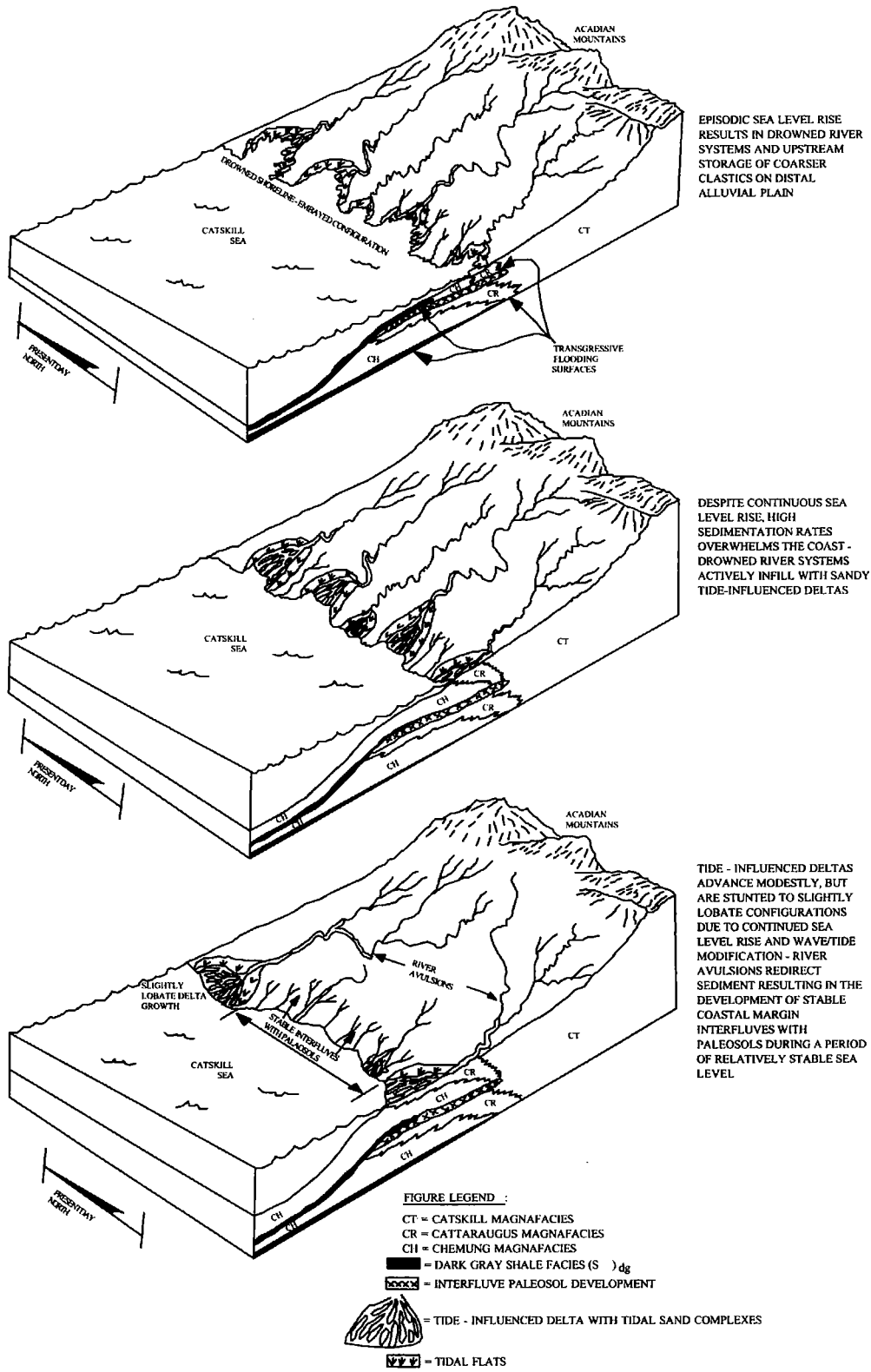


Fig. 13 Schematic, three-dimensional reconstruction of sedimentary environments and progressive development of shoreline configuration and depositional cyclicity. Note the intimate association of paleosols with fully marine and tidally-influenced paleoenvironments.

## ACKNOWLEDGEMENTS

This paper constitutes a major expansion upon parts of the M.A. thesis of D. Bishuk, completed at SUNY Oneonta. We thank Dr. Edward Cotter for helpful field discussions at several outcrops and follow-up correspondence during early versions of the manuscript, which helped refine our facies and paleosol interpretations. We are greatly indebted to Sherri Pullar of EA Science and Technology for her outstanding AutoCad drafting services. This project would not have reached its current level of completeness without the undergraduate research contributions of Ann McSpedon, which consisted of thin section preparation and analysis, and preliminary interpretations of paleosol micromorphology. Much love and appreciation is extended to Mary Bishuk for enduring support and understanding over the 3-year research period.

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**FIELD TRIP ROAD LOG**

| <u>Cumulative Mileage</u> | <u>Miles From Last Point</u> | <u>Route Description</u>  |
|---------------------------|------------------------------|---|
| 0.0                       | 0.0                          | Mileage begins from the Hunt Union parking lot on the SUNY Oneonta campus. Make a right out of the parking lot and proceed to stop sign (bear left).  |
| 0.1                       | 0.1                          | Make a left onto Ravine Parkway and proceed 0.6 miles to stop sign.   |
| 0.7                       | 0.6                          | Make a left onto West Street. Proceed down hill for 0.7 miles to traffic light.   |
| 1.4                       | 0.7                          | Make a left at traffic light onto Chestnut Street. Proceed 0.3 miles to second traffic light.   |
| 1.7                       | 0.3                          | Make a right at traffic light onto Main Street. Proceed 0.5 miles to Junction I-88.   |
| 2.2                       | 0.5                          | Make a right onto Interstate I-88 Exit 14 on-ramp heading westbound. Proceed 17.3 miles to exit 10.   |
| 19.5                      | 17.3                         | Take exit 10 off I-88 at Unadilla.  |
| 20.1                      | 0.6                          | As you descend off of long off-ramp, bear left. Make your first left onto River Road located just before bridge over Susquehanna River. Proceed on River Road to the second road on the left. |
| 21.1                      | 1.0                          | Make a hard left onto Delaware County Route 23 heading eastbound. Proceed to the third road on the right.   |
| 22.3                      | 1.2                          | Turn right onto Dunshee Road. Proceed up mountain for entire length of Dunshee Road to stop sign.   |
| 23.9                      | 1.6                          | Turn right onto Delaware County Route 35. Proceed up hill to first road on the left.  |

| <u>Cumulative Mileage</u> | <u>Miles From Last Point</u> | <u>Route Description</u>  |
|---------------------------|------------------------------|---|
| 24.3                      | 0.4                          | Turn left onto Skytop Lane and proceed to dead end.   |
| 24.6                      | 0.3                          | At dead end, park in dirt parking area near gate entrance to <b>STOP #1- SKYTOP LANE QUARRY (outcrop 43)</b> . See text for descriptions and interpretations of facies $T_{xb-td}$ , $T_{xb-ti}$ , $P_{bms}$ , and $H_b-i$ , and paleosols $P_p$ and $P_{ao}$ . |
| 24.9                      | 0.3                          | Return to intersection of Skytop Lane and Delaware County Route 35, Turn right onto CR 35. Proceed 1.5 miles on Delaware County Route 35 to Village of Sidney Center .  |
| 26.4                      | 1.5                          | Turn left onto Maywood Lane (Historical Depot Drive).   |
| 26.5                      | 0.1                          | Proceed up hill to Maywood Historical Depot parking area. Eat lunch and attend optional tour of Historical Railroad Depot.  |
| 26.6                      | 0.1                          | Return to intersection of Maywood Lane and CR 35. Turn right onto CR35. Proceed 3.0 miles on CR 35.   |
| 29.6                      | 3.0                          | You will see a small lake on your left. Turn left onto Cummings Road and proceed 0.6 miles.   |
| 30.2                      | 0.6                          | Turn right onto gravel access road to <b>Sheetz Quarry (outcrop 51) = STOP #2</b> . Follow gravel access road 0.5 miles to gate   |
| 30.7                      | 0.5                          | Park at gate and proceed on foot to quarry  |
| 31.2                      | 0.5                          | Return to Cummings Road and turn left. Proceed 0.6 miles on Cummings Road.  |
| 31.8                      | 0.6                          | Turn left onto CR 35 and proceed 3.5 miles to intersection with NYS Route 206.  |

| <u>Cumulative Mileage</u> | <u>Miles From Last Point</u> | <u>Route Description</u>   |
|---------------------------|------------------------------|--|
| 35.3                      | 3.5                          | Make a right onto NYS Route 206. Proceed 1.4 miles to the Hamlet of Masonville   |
| 36.7                      | 1.4                          | Make a right at stop sign onto NYS Route 8 North. Proceed 3.2 miles on NYS Route 8 to Delaware County Route 4.                                       |
| 39.9                      | 3.2                          | Make a right onto CR 4. Proceed to stop sign at intersection with Cole Road spur.  |
| 40.0                      | 0.1                          | Continue straight up hill to Sidney Mountain Quarry access road.   |
| 40.2                      | 0.2                          | Bear left onto gravel access road and follow to entrance gate for Sidney Mountain Quarry. Park on side of access road and proceed on foot to quarry. |
| 40.3                      | 0.1                          | <b>STOP #3 Sidney Mountain Quarry (outcrop 6)</b>  |

Return to intersection of CR4 and NYS Route 8 and turn right onto 8 North to reach I-88.



### APPENDIX I: LOCATIONS OF SONYEA OUTCROPS USED IN THIS STUDY

| OUTCROP NUMBER | LOCATION  | QUADRANGLE | LATITUDE   | LONGITUDE  | ELEVATION (Ft AMSL) | ASSIGNED FACIES  | MAGNAFACIES/FORMATION                          |
|----------------|---|------------|------------|------------|---------------------|--|--|
| 1              | Interstate-88 at Exit 8, Bainbridge                     | Sidney     | 42°17' 37" | 75°27' 51" | 1180                | HS <sub>cs</sub>   | Chemung/Triangle                               |
| 2              | Route 8, Sidney   | Sidney     | 42°17' 14" | 75°23' 52" | 1420                | HS <sub>cs</sub> /S <sub>dg</sub>  | Chemung/Glen Aubrey-Sawmill Creek              |
| 3              | Thorpe Road at Route 8, Sidney                          | Sidney     | 42°16' 50" | 75°23' 51" | 1600                | HS <sub>cs</sub> /S <sub>dg</sub> /HS <sub>cs</sub> -A   | Chemung/Glen Aubrey-Sawmill Creek              |
| 4              | Route 8 near Bundy Hollow Road, Sidney                  | Sidney     | 42°15' 46" | 75°23' 56" | 1400                | HS <sub>cs</sub>   | Chemung/Glen Aubrey-Sawmill Creek              |
| 5              | Delaware County Route 4, Sidney                         | Sidney     | 42°17' 14" | 75°23' 29" | 1460                | HS <sub>cs</sub>   | Chemung/Glen Aubrey                            |
| 6 = STOP #3    | Sidney Mountain Quarry, Delaware County Route 4, Sidney | Sidney     | 42°16' 35" | 75°23' 20" | 1779                | P <sub>bms</sub> /H <sub>b</sub> -sl/H <sub>b</sub> -i/P <sub>gp</sub> /P <sub>aa</sub> /S <sub>dg</sub> /HS <sub>cs</sub> | Cattaraugus/Unnamed Fm and Chemung/Glen Aubrey |
| 7              | Pine Hill Road, Sidney                                  | Unadilla   | 42°17' 30" | 75°21' 50" | 1851                | T <sub>xb</sub> -td  | Cattaraugus/Lower Walton                       |
| 8              | Dunshee Road Quarry, Sidney Center                      | Unadilla   | 42°17' 06" | 75°17' 08" | 1720                | HS <sub>cs</sub> -A  | Chemung/Glen Aubrey                            |
| 9              | Dunshee Hill, Steele Pasture Quarry, Sidney Center      | Unadilla   | 42°17' 13" | 75°16' 58" | 1728                | HS <sub>cs</sub> /S <sub>dg</sub> /HS <sub>cs</sub> -A   | Chemung/Glen Aubrey-Sawmill Creek              |
| 10             | Dunshee Hill, Outcrop A, Sidney Center                  | Unadilla   | 42°17' 32" | 75°17' 02" | 1745                | T <sub>xb</sub> -td  | Chemung/Glen Aubrey                            |
| 11             | Dunshee Hill, Outcrop B, Sidney Center                  | Unadilla   | 42°17' 30" | 75°16' 57" | 1760                | T <sub>xb</sub> -td  | Cattaraugus/Unnamed Fm                         |
| 12             | Dunshee   | Unadilla   | 42°17' 29" | 75°16' 55" | 1787                | T <sub>xb</sub> -td  | Cattaraugus/                                   |

|                                     |  |                        |                      |                       |                                     |  |  |
|-------------------------------------|--|------------------------|----------------------|-----------------------|-------------------------------------|--|--|
|                                     | Hill,<br>Outcrop C,<br>Sidney<br>Center            |                        |                      |                       |                                     |  | Unnamed Fm                             |
| 13                                  | Dunshee<br>Hill,<br>Outcrop D,<br>Sidney<br>Center | Unadilla               | 42°17' 32"           | 75° 16' 49"           | 1790                                | T <sub>xb</sub> -td                      | Cattaraugus/<br>Unnamed Fm             |
| 14                                  | Dunshee<br>Hill,<br>Outcrop E,<br>Sidney<br>Center | Unadilla               | 42° 17'<br>35"       | 75° 16' 48"           | 1748                                | T <sub>xb</sub> -td                      | Cattaraugus/<br>Unnamed Fm             |
| 15                                  | Dunshee<br>Hill,<br>Outcrop F,<br>Sidney<br>Center | Unadilla               | 42° 17'<br>25"       | 75° 16' 38"           | 1768                                | T <sub>xb</sub> -td                      | Cattaraugus/<br>Unnamed Fm             |
| 16                                  | Dunshee<br>Hill,<br>Outcrop G,<br>Sidney<br>Center | Unadilla               | 42°17' 32"           | 75° 16' 44"           | 1850                                | T <sub>xb</sub> -td                      | Cattaraugus/<br>Unnamed Fm             |
| <b>OUTCRO<br/>P<br/>NUMBE<br/>R</b> | <b>LOCATIO<br/>N</b>                               | <b>QUADRANG<br/>LE</b> | <b>LATITUD<br/>E</b> | <b>LONGITU<br/>DE</b> | <b>ELEVATIO<br/>N<br/>(Ft AMSL)</b> | <b>ASSIGNE<br/>D<br/>FACIES</b>          | <b>MAGNAFACI<br/>ES/<br/>FORMATION</b> |
| 18                                  | Dunshee<br>Hill,<br>Outcrop I,<br>Sidney<br>Center | Unadilla               | 42°17' 33"           | 75° 16' 43"           | 1775                                | T <sub>xb</sub> -td                      | Cattaraugus/<br>Unnamed Fm             |
| 19                                  | Dunshee<br>Hill,<br>Outcrop J,<br>Sidney<br>Center | Unadilla               | 42° 17'<br>34"       | 75° 16' 44"           | 1769                                | T <sub>xb</sub> -td                      | Cattaraugus/<br>Unnamed Fm             |
| 20                                  | Dunshee<br>Hill,<br>Outcrop K,<br>Sidney<br>Center | Unadilla               | 42°17' 27"           | 75° 16' 36"           | 1774                                | T <sub>xb</sub> -td                      | Cattaraugus/<br>Unnamed Fm             |
| 21                                  | Dunshee<br>Hill,<br>Outcrop L,<br>Sidney<br>Center | Unadilla               | 42° 17'<br>29"       | 75° 16' 36"           | 1820                                | T <sub>xb</sub> -td/<br>P <sub>bms</sub> | Cattaraugus/<br>Unnamed Fm             |
| 22                                  | Dunshee<br>Hill,<br>Outcrop M,<br>Sidney<br>Center | Unadilla               | 42° 17'<br>21"       | 75° 16' 20"           | 1795                                | T <sub>xb</sub> -td                      | Cattaraugus/<br>Unnamed Fm             |
| 23                                  | Dunshee<br>Hill,<br>Outcrop N,<br>Sidney<br>Center | Unadilla               | 42° 17'<br>25"       | 75° 16' 36"           | 1775                                | T <sub>xb</sub> -td                      | Cattaraugus/<br>Unnamed Fm             |
| 24                                  | Delaware<br>County                                 | Unadilla               | 42°16' 56"           | 75° 16' 52"           | 1880                                | P <sub>bms</sub>                         | Cattaraugus/<br>Unnamed Fm             |

|    |   |          |                |             |      |  |                            |
|----|---|----------|----------------|-------------|------|--|----------------------------|
|    | Route 35,<br>Outcrop O,<br>Sidney<br>Center                       |          |                |             |      |  |                            |
| 25 | Delaware<br>County<br>Route 35,<br>Outcrop P,<br>Sidney<br>Center | Unadilla | 42° 16'<br>53" | 75° 16' 52" | 1875 | P <sub>bms</sub>                       | Cattaraugus/<br>Unnamed Fm |
| 26 | Delaware<br>County<br>Route 35,<br>Outcrop Q,<br>Sidney<br>Center | Unadilla | 42° 16' 40"    | 75° 17' 00" | 1880 | P <sub>bms</sub>                       | Cattaraugus/<br>Unnamed Fm |
| 27 | Dunshee<br>Hill,<br>Outcrop R,<br>Sidney<br>Center                | Unadilla | 42° 17'<br>15" | 75° 16' 47" | 1762 | HS <sub>cs</sub> -A                    | Chemung/Glen<br>Aubrey     |
| 28 | Dunshee<br>Hill,<br>Outcrop S,<br>Sidney<br>Center                | Unadilla | 42° 17' 15"    | 75° 16' 44" | 1756 | HS <sub>cs</sub> -A                    | Chemung/Glen<br>Aubrey     |
| 29 | Dunshee<br>Hill,<br>Outcrop T,<br>Sidney<br>Center                | Unadilla | 42° 17'<br>18" | 75° 16' 48" | 1773 | T <sub>xb</sub> -ti                    | Cattaraugus/<br>Unnamed Fm |
| 30 | Dunshee<br>Hill,<br>Outcrop U,<br>Sidney<br>Center                | Unadilla | 42° 17'<br>16" | 75° 16' 51" | 1760 | HS <sub>cs</sub> -A                    | Chemung/Glen<br>Aubrey     |
| 31 | Dunshee<br>Hill,<br>Outcrop V,<br>Sidney<br>Center                | Unadilla | 42° 17'<br>14" | 75° 16' 55" | 1739 | HS <sub>cs</sub> -A                    | Chemung/Glen<br>Aubrey     |
| 32 | Dunshee<br>Hill,<br>Quarry 1,<br>Sidney<br>Center                 | Unadilla | 42° 17' 21"    | 75° 16' 52" | 1799 | P <sub>bms</sub>                       | Cattaraugus/<br>Unnamed Fm |
| 33 | Dunshee<br>Hill,<br>Quarry 2,<br>Sidney<br>Center                 | Unadilla | 42° 17' 24"    | 75° 16' 48" | 1825 | P <sub>bms</sub>                       | Cattaraugus/<br>Unnamed Fm |
| 34 | Dunshee<br>Hill,<br>Quarry 3,<br>Sidney<br>Center                 | Unadilla | 42° 17'<br>25" | 75° 16' 44" | 1845 | H <sub>b</sub> -i/ T <sub>xb</sub> -ti | Cattaraugus/<br>Unnamed Fm |
| 35 | Dunshee<br>Hill,<br>Quarry 4,<br>Sidney                           | Unadilla | 42° 17'<br>26" | 75° 16' 41" | 1836 | P <sub>bms</sub> / H <sub>b</sub> -i   | Cattaraugus/<br>Unnamed Fm |

|                           |  |                   |                 |                  |                                |   |   |
|---------------------------|--|-------------------|-----------------|------------------|--------------------------------|---|---|
| 36                        | Center<br>Dunshee Hill,<br>Quarry 5,<br>Sidney Center                                    | Unadilla          | 42°17' 29"      | 75° 16' 39"      | 1836                           | P <sub>bms</sub> / H <sub>b-i</sub>   | Cattaraugus/<br>Unnamed Fm              |
| 38                        | Delaware County<br>Route 35,<br>Outcrop V,<br>Sidney Center                              | Unadilla          | 42° 17'<br>14"  | 75° 16' 55"      | 1739                           | HS <sub>cs</sub>  | Chemung/Glen<br>Aubrey                  |
| <b>OUTCROP<br/>NUMBER</b> | <b>LOCATION</b>  | <b>QUADRANGLE</b> | <b>LATITUDE</b> | <b>LONGITUDE</b> | <b>ELEVATION<br/>(Ft AMSL)</b> | <b>ASSIGNED<br/>FACIES</b>  | <b>MAGNAFACIES/<br/>FORMATION</b>       |
| 40                        | Delaware County<br>Route 35,<br>Outcrop X,<br>Sidney Center                              | Unadilla          | 42° 16'<br>24"  | 75° 17' 01"      | 1880                           | HS <sub>cs</sub> / S <sub>dg</sub>  | Chemung/Glen<br>Aubrey-Sawmill<br>Creek |
| 41                        | Roof Road,<br>Outcrop Y,<br>Sidney Center  | Unadilla          | 42°16' 33"      | 75° 15' 35"      | 1480                           | HS <sub>cs</sub> / S <sub>dg</sub>  | Chemung/Glen<br>Aubrey-Sawmill<br>Creek |
| 42                        | Roof Road,<br>Outcrop Z,<br>Sidney Center  | Unadilla          | 42°16' 21"      | 75° 16' 09"      | 1720                           | T <sub>xb-ti</sub>  | Cattaraugus/<br>Unnamed Fm              |
| <b>43 =<br/>STOP #1</b>   | <b>Skytop<br/>Lane<br/>Quarry,<br/>Skytop<br/>Lane,<br/>Sidney<br/>Center</b>            | Unadilla          | 42°16' 53"      | 75° 16' 20"      | 1851                           | P <sub>bms</sub> / H <sub>b-i</sub> /<br>P <sub>ao</sub> /P <sub>bo</sub> /P <sub>p</sub> /<br>T <sub>xb-ti</sub> | Cattaraugus/<br>Unnamed Fm              |
| 44                        | Skytop<br>Lane,<br>Outcrop<br>AA, Below<br>Skytop<br>Lane<br>Quarry,<br>Sidney<br>Center | Unadilla          | 42° 16'<br>59"  | 75° 16' 26"      | 1750                           | T <sub>xb-td</sub>  | Cattaraugus/<br>Unnamed Fm              |
| 45                        | Skytop<br>Lane Spur,<br>Outcrop<br>BB, Sidney<br>Center                                  | Unadilla          | 42° 16'<br>30"  | 75° 16' 23"      | 1680                           | H <sub>b-i</sub> / T <sub>xb-ti</sub>   | Cattaraugus/<br>Unnamed Fm              |
| 46                        | Skytop<br>Lane Spur,<br>Outcrop<br>CC, Sidney<br>Center                                  | Unadilla          | 42°16' 28"      | 75° 16' 24"      | 1720                           | T <sub>xb-td</sub>  | Cattaraugus/<br>Unnamed Fm              |
| 47                        | Delaware<br>County<br>Route 35,  | Unadilla          | 42°16' 16"      | 75° 17' 14"      | 1880                           | T <sub>xb-ti</sub> /P <sub>p</sub>  | Cattaraugus/<br>Unnamed Fm              |



|                         |   |          |                |             |      |                                 |                            |
|-------------------------|---|----------|----------------|-------------|------|---------------------------------|----------------------------|
|                         | Outcrop<br>DD, East<br>Masonville   |          |                |             |      |                                 |                            |
| 48                      | Unnamed<br>Quarry,<br>Wilcox<br>Road at<br>Olmstead<br>Road, East<br>Masonville | Unadilla | 42° 16'<br>32" | 75° 19' 03" | 1860 | $P_{bms}/P_{aa}$                | Cattaraugus/<br>Unnamed Fm |
| 49                      | Turner<br>Quarry,<br>Olmstead<br>Road, East<br>Masonville                       | Unadilla | 42° 16' 30"    | 75° 18' 44" | 1880 | $P_{bms}$                       | Cattaraugus/<br>Unnamed Fm |
| 50                      | Stanton<br>Quarry,<br>Olmstead<br>Road, East<br>Masonville                      | Unadilla | 42° 16' 44"    | 75° 18' 26" | 1720 | $T_{xb-td}/$<br>$P_{bms}$       | Cattaraugus/<br>Unnamed Fm |
| <b>51 =<br/>STOP #2</b> | <b>Sheetz<br/>Quarry,<br/>Cummings<br/>Road, East<br/>Masonville</b>            | Unadilla | 42° 15' 10"    | 75° 18' 29" | 1909 | $P_{bms}/H_b-i$                 | Cattaraugus/<br>Unnamed Fm |
| 52                      | Cardi<br>Quarry,<br>Cummings<br>Road, East<br>Masonville                        | Unadilla | 42° 14'<br>51" | 75° 18' 09" | 1960 | $P_{bms}$                       | Cattaraugus/<br>Unnamed Fm |
| 53                      | Delaware<br>County<br>Route 27,<br>Sidney<br>Center                             | Unadilla | 42° 15'<br>39" | 75° 15' 52" | 1932 | $T_{xb-ti}/$<br>$P_{aa}/P_{gp}$ | Cattaraugus/<br>Unnamed Fm |

## **TRIP A3**

### **Glacial regime and depositional environments along the retreating Laurentide Ice Sheet, northeastern Appalachian Plateau, New York**

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#### **ABSTRACT**

Deglacial environments along the retreating Laurentide Ice Sheet throughout of the upper Susquehanna region were primarily controlled by an ice front configuration draped across the high relief, Appalachian Plateau terrain. Valley ice tongues extended 20 km beyond upland positions commonly terminated in ice-contact lakes. Active ice flow nourished by the ice sheet persisted within through valleys, where as ice tongues in non-through valleys became detached and stagnated. Temperate conditions and hydrologic connection with the ice sheet supported highly turbid englacial and subglacial tunnel discharge directly into through valley ice-contact lakes. Subaerial streams favoring positions lateral to ice tongues transported sand and gravel outwash to prograding deltas. Thus, bottomset and toeset sediments interfinger with lacustrine silt and sand. Post-glacial dam incision followed by lake emptying led to the modern landscape. A finer facies of thick silt and sand lies beneath the modern floodplain and is typically bound by deltaic terraces (commonly called kame terraces) consisting of the coarser facies sand and gravel.

In contrast, downwasting over non-through valley headward divides led to loss of nourishment and near separation of the ice tongue from the ice sheet, thus causing the combined effect of hydrologic disconnection and stagnation. Consequently, non-through valley environments of deposition were less uniform, characterized by a) local ponding behind ephemeral ice-cored dams, b) kilometer-size, detached ice masses partially covered by outwash and lake sediments, and c) meltwater sources limited to the volume of the detached ice tongue. These collectively describe an environment significantly different from through valleys. Non-through valley sediments are far more diverse, less well sorted, and lack the laterally uniform subsurface conditions characteristic of through valleys. In addition, the landform assemble also differs through the lack of a continuous lacustrine plain, segmented terraces graded to local base levels, isolated small esker-forms, kame and kettle topography unrelated to systematic moraines (called kame fields), and mega-kettles that span the valley floor (known as dead ice sinks).

The Bering Glacier, Alaska, depicting sedimentary conditions in ice-contact lakes and along peripheral drainage systems is cited as an analog for the retreating Laurentide Ice Sheet in central New York.

#### **INTRODUCTION**

The Susquehanna River and its tributaries occupy glacially modified valleys that form an asymmetric, dendritic pattern of considerable antiquity on the coarsely dissected Appalachian Plateau (Figure 1). Major north-south valleys are oriented parallel to the general direction of overriding ice movement thus forming broad, U-shaped troughs separated by wide, low-relief divides. Many valley segments show the effects of glacier flow in oversteepened slopes, and several have asymmetric cross-

Figure 1. Index map.



sectional profiles, with steeper, west-facing and less steep, east-facing slopes, the cause of which appears unrelated to glacial erosion. Furthermore, the gently dipping (<10 degrees SW) Middle Devonian bedrock strata only have subtle expression in the topography, thus explanation for asymmetry remains elusive. Valleys oriented orthogonal to glacier movement show lesser effects of glacial erosion, although many have steep slopes assumed inherited from pre-glacial fluvial incision (Fleisher 1977). Local relief ranges between 250-300 m, but valley water wells (Randall, 1972) frequently penetrate 100-125 m of Quaternary sediment, thereby indicating that bedrock relief is considerably greater, averaging 300 m and reaching 400 m in some locations. The ice-marginal configuration of a temperate glacier draped across terrain was strongly influenced by this topographic relief. Depositional environments at the snout of ice salients within each valley were in significant contrast with those on adjacent uplands. The resulting landforms quite different from classic mid continent features formed along regionally extensive ice lobes.

This region received little published attention between Fairchild's 1925 description of the glacial landscape (kame and kettle topography, pitted plains, terraces, proglacial lakes and hanging deltas) and Coates' definition of the till shadow effect in 1966. Woodfordian facies of "bright" and "drab" drift were the subject of study in the Binghamton/Elmira area, (Denny, 1956; Moss and Ritter, 1962; Coates, 1963), but consideration of depositional environments along the main Susquehanna Valley remained unreported until described by Melia (1975) and Fleisher (1977a). Fleisher (1984, 1985), used driller's logs (water wells) and landform expression as indicators of proglacial ice-contact lakes, proposed a late glacial lake chain throughout the upper Susquehanna drainage basin. Interpretation of chronology is hampered by the lack of datable materials. However, Krall (1977) interpreted the Cassville-Cooperstown Moraine (5 km south of Otsego Lake) to be topographically correlative with moraines in the Hudson Valley, thereby suggesting a readvance circa 14,000 years BP. However, Ridge et al. (1991) tentatively correlated the Cassville-Cooperstown Moraine with the West Canada Readvance of Late Wisconsinan, pre-Valley Heads age (circa 15.5 ka) based on elevation projections of ice-marginal deposits in the Mohawk Valley and the limit of readvance inferred from subsurface stratigraphy (Fleisher, 1986).

Coates (1974) defines the "Through Valley Section" of the Appalachian Plateau to be characterized by the valley of Otsego Lake at the head of the Susquehanna drainage basin. In this context a through valley may be depicted as a scoured glacial trough, oriented parallel to the regional ice flow direction, and "open-to-the-north" as a hanging valley on the northern plateau escarpment. Unlike non-through valleys that conventionally rise to upland divides, the head of the through valley drainage basin is on the valley floor where it separates streams flowing north out of low, poorly drained valley bogs and swamps from those that leave the same bogs and swamps but flow southward. The open through valley trough extends northward beyond the limits of the water divide. The valley walls do not converge, but rather remain on opposite sides of the valley where they gradually separate to join the escarpment slopes that define the northern boundary of the Appalachian Plateau (Figure 1).

As Laurentide ice moved southward from the southern Adirondack terrain and across the Mohawk Valley, it passed from crystalline terrain to an Appalachian Plateau substrate dominated by fine-grained, lower Paleozoic siltstone and shale. The combined influence of high topographic relief and a deformable substrate led to the development of valley ice tongues 20 km long (Fleisher, 1993).

MacNish and Randall (1982) used water well and boring log data to establish Quaternary aquifer properties, and diagrammatically illustrate generalized subsurface information and landforms to interpreted rate (rapid versus slow) and mode of retreat (active versus stagnant). Fleisher (1991 a, 1991b, 1993, 1986a) established a link between glacial regime and physiographic settings by recognizing that landforms produced by downwasting and ice-lobe collapse within non-through valleys differ significantly from those formed during active ice backwasting in through valleys, thereby supporting the MacNish and Randall association of retreat regime with landforms and subsurface stratigraphy.

The primary purpose of this paper is to present potential sediment sources, transport mechanisms, environments of sediment accumulation, and landform development using analog conditions of contemporary processes at selected glacial ice margins in Alaska.

### **VALLEY ICE-LOBE MODEL**

The traditional view of the deglacial history of this portion of the northern Appalachian Plateau suggest regional backwasting of a more or less continuous ice margin of the Laurentide Ice Sheet during pre-Valley Heads (~18-15.5 ka) recession, with isolated areas of downwasting (Krall, 1977) . The ice front configuration is portrayed to consist of relatively short "ice-lobes" in pre-existing valleys (Cadwell, 1972; Fleisher and Cadwell, 1984; Fleisher, 1986a). Originally proposed by Moss and Ritter (1962), the ice-lobe model illustrates an ice margin in the form of topographically-controlled, short, valley lobes (a few kilometers in length) extending from upland recessions (Figure 2).

Using a combination of driller's logs from water wells and test borings and glacial landforms, MacNish and Randall (1982) developed diagrammatic illustrations of ground water distribution that combined rate of retreat (rapid or slow) with mode of ice-lobe activity (backwasting or downwasting). Fleisher (1991a, b) recognized that landforms produced by downwasting and ice-lobe collapse within non-through valleys differ significantly from those formed during active ice backwasting in through valleys, thereby supporting the MacNish and Randall association of retreat regime with landforms and subsurface stratigraphy.

#### **Shortcomings of the short ice-lobe model**

The early ice-lobe model has been useful in the development of ideas and concepts related to the origin of glacial landforms and environments of deposition. However, it fails to account for several puzzling aspects of regionally extensive deposits and does not portray sufficient details pertaining to depositional environments. For example, it does not address the following observations; 1) different landform assemblages in different valleys, 2) kame-moraines lose topographic expression on valley walls and across divides, which defies confident correlation from one valley to the next, and 3) moraines predominantly consist of crudely-sorted and stratified sand and gravel (some till) that interfinger with thick lacustrine silt, yet well-developed strandline features are uncommon. Furthermore, the model fails to identify sediment sources and transport mechanisms, and does not offer a basis for interpretation of anomalous landforms and stratigraphy. For these reasons, Fleisher (1993) proposed a modified ice-tongue model that takes into consideration the influence of a deformable bed on the ice surface gradient, and consequently the length of the ice tongue.

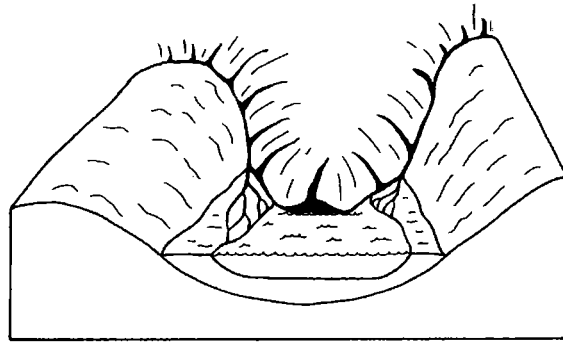


Figure 2. Short Ice-lobe model.

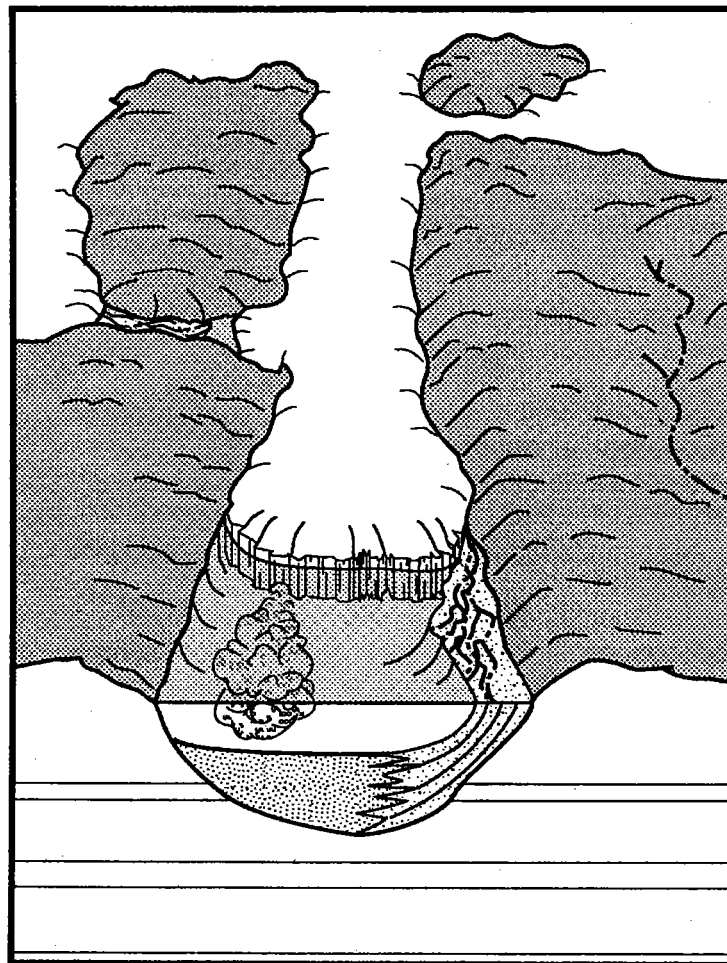


Figure 3. Modified ice-tongue model.

#### **Extended ice-tongue model**

As Laurentide ice advanced southward from the crystalline terrain of the southern Adirondacks, it crossed the Mohawk Valley and ascended the north-facing escarpment of the Appalachian Plateau where 450 to 600 m of Ordovician and Silurian shale and siltstone (60-

70% of the stratigraphic column) are exposed. Within the eastern Susquehanna drainage basin, lower and middle Devonian lithologies contain an additional 485 m of shale and siltstone that dip gently to the south-southwest (Rickard and Zenger, 1964).

Boulton and Jones (1979) present evidence to suggest "the glacier profile is related to the hydraulic and strength properties of potentially deformable bed material". Theoretical modeling predicts the strain response of a glacial substrate as critical shear stress exceeds sediment strength and subsequent initiation of sediment deformation. Shear stress and confining pressure are resisted by sediment strength (i.e., cohesion and frictional strength) (Benn and Evans, 1998; Boulton and Hindmarsh, 1987; Alley, 1989; Tulaczyk, 1999).

Water conditions at the sole of temperate glaciers provides maximum pore-water pressure and diminished intergranular friction in the presence of maximum shear stress. For fine-grained sediment, from which water does not readily escape, pore-water pressure builds, thus causing diminished strength. At a critical strain rate, dilation leads to ductile flow and pervasive deformation (Alley, 1989; Benn and Evans, 1998). Conversely, coarse, well-drained sand and gravel facilitate water escape thus reducing or precluding the potential for deformation (Boulton et al., 1974; Boulton and Hindmarsh, 1987).

The glacial substrate on the Appalachian Plateau would have been charged with saturated, fine sediment of low permeability derived from the erosion of lower Paleozoic shales and siltstones as the ice moved into the Susquehanna drainage basin. As modeling indicates, with low permeable materials, water pressure builds and bed deformation follows. Thus, the development of positive pore pressure within saturated bed material facilitated deformation and the development of a lower equilibrium profile (flatter ice surface gradient). Consequently, the ice-tongues may have reached lengths in excess of 20 km (Figure 3) (Fleisher, 1993). Calculations leading to this assume specific quantitative conditions that may be more precise than accurate for the purpose of developing a conceptual glacial model. Therefore, the most important aspect of the modified model is not the finite length of ice tongues, but rather the presence of valley ice adjacent ice-free upland slopes on which inwash process may have been active.

## **GLACIAL REGIME AND LANDFORM ASSEMBLAGES**

### **Glacial regime**

The Susquehanna River drains the eastern Appalachian Plateau in an asymmetric, dendritic pattern of glacially enlarged north-south oriented valleys and less well developed east-west tributaries (Figure 1). Gently undulating upland divides separate broad through and non-through valleys with local topographic relief of 200-330 m. Regional water well data (Randall, 1972) indicate that valley floors are underlain by stratified drift that commonly exceeds 100 m in thickness, which means the bedrock relief approaches 400 m. This magnitude of relief influenced the configuration of the retreating Laurentide ice margin, thus forming 20 km long ice-tongues in valleys oriented parallel to the ice flow direction (NNE-SSW). The dynamics of long ice-tongues in this terrain would favor stagnation and the development of detached remnant ice masses where nourishment from the ice sheet was restricted or limited. Evidence of this may be found in a) through valleys where active ice flow persisted during backwasting, in b) non-through valleys where nourishment to ice-tongues was restricted by thinning ice across upland divides, thus leading to downwasting, and in c) transverse valleys deprived of ice tongues by virtue of their orientation. This regime yielded two distinctly different landform assemblages, one representative of active ice retreat in through valleys and another depicting widespread stagnation and downwasting in non-through valleys and transverse valleys.

### Through valley flow regime and landform assemblage

Observed detachment of ice masses along steeply rising internal structures (taken to be deep seated thrusts or shears) within the termini of several Alaska glaciers (Mulholland, 1982; Fleisher, 1993) lead to the concept of an ice-marginal "cleat" upon which active ice ramps, thus serving as a mechanism for ice mass separation within active ice retreat. It is suggested that similar ice-marginal cleats measuring ~70 m thick, 1 km wide and 3 km long formed semi-continuously during backwasting (Figure 4). Subsequent burial by both outwash and inwash would yield remnant ice masses and ice-cored valley train.

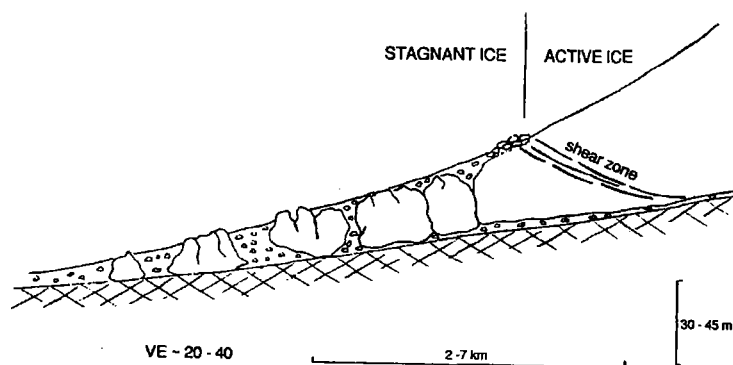


Figure 4. Detached and stagnant ice. (modified from Mulholland, 1982)

Through-valley landforms primarily consist of kame moraines, pitted and dissected valley train, kame terraces, kame fields, lacustrine plains, and occasional dead-ice sinks (Figure 5). Kame moraines typically consist of poorly-sorted and crudely-stratified, matrix-supported silty gravel, with ice-contact collapse structures. Boulders are well rounded and lack the effect of glacial abrasion. Data from wells and borings indicate that moraine gravels interfinger with lacustrine silts on the stoss side of moraines, which suggest that moraines served as dams for proglacial lakes. Moraines rise 25-30 m above the floodplain and modern streams that breach them. Their conspicuous kame and kettle topography typically reaches 10-15 m of local relief and occupies the full width of the valley floor. However, their use as morphostratigraphic units in regional correlation is limited because their topographic expression fades significantly on valley walls and is absent from uplands.



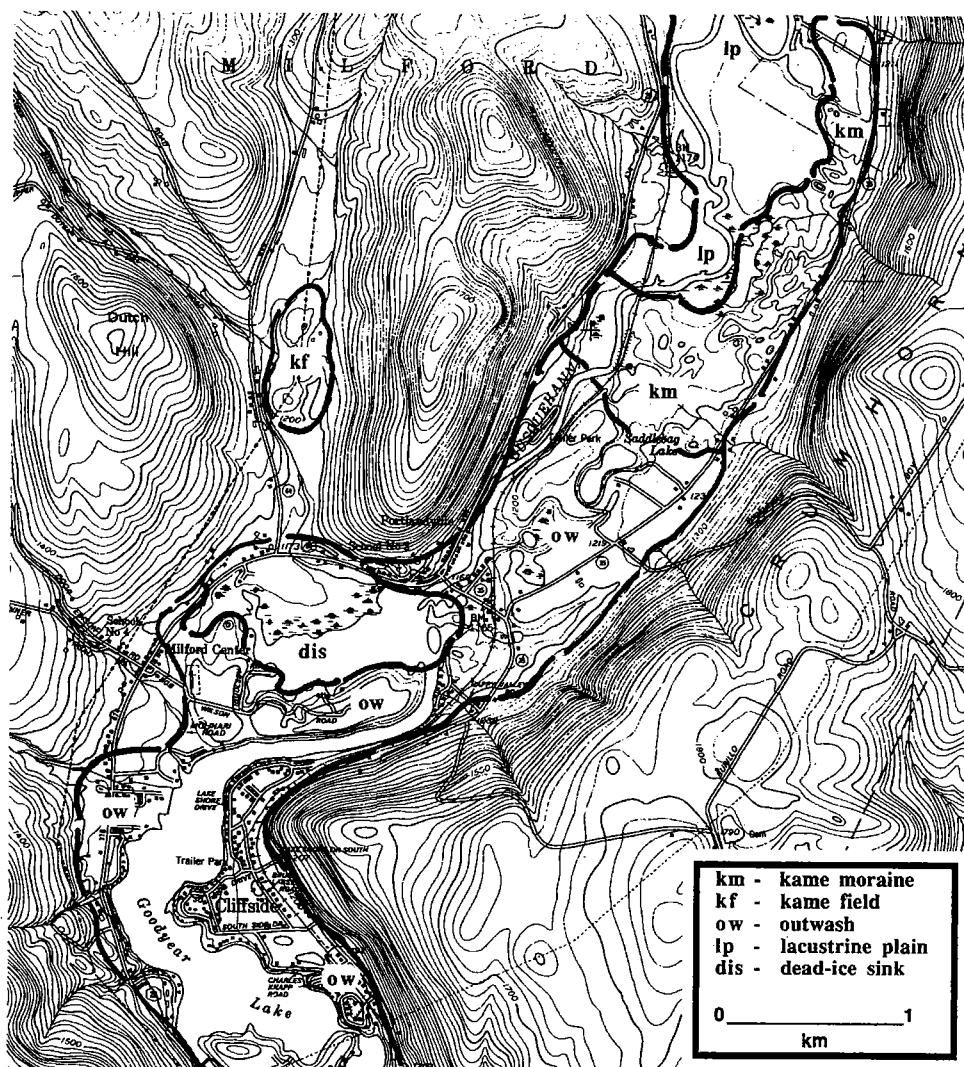


Figure 5. Through valley landform assemblage.

Pitted and dissected valley trains are often graded to kame moraines, but are also known to exist independently. They are found 20-25 m above the modern floodplain and represent the highest planar deposits in the valleys. Therefore, they have been referred to as high outwash as a means of distinguishing them from similar deposits at lower elevations which are called low outwash (Fleisher, 1986a). Most outwash is well-sorted, clast-supported, coarse, sandy gravel interstratified with layers and lenses of less well-sorted, matrix-supported silty gravel.

Kame terraces stand 25-30 m above the modern floodplain in paired and non-paired segments and may be topographically indistinguishable from dissected valley train. They are

typically less than 2 km long and commonly contain well developed deltaic internal structure, although associated lacustrine plains are not consistently present. Topsets are rather poorly-sorted and may show only the most subtle indication of imbrication, channel structures, and cross bedding. Deltaic foresets are massive in scale, occupy the full height of the terrace beneath topsets, and indicate progradation toward the axis of the valley. Texture ranges from coarse, pebbly sand to lenses of very coarse, matrix-supported gravel. A sandy and silt-rich toeset facies is occasionally exposed in borrow pits. Kame terrace segments found at tributary mouths are at grade with dissected hanging deltas.

Portions of most main valley floodplains contain broad and extensive, low gradient reaches across which exaggerated meanders flow. This topographic setting characterizes lacustrine plains throughout the region. Water well data indicate the terrace sands and gravels interfinger laterally at depth with 50-120 m of sand, coarse silt (locally called quicksand) and lesser amounts of clay, all of which are assumed to be of lacustrine origin (Randall, 1972). However, the sparse occurrence of well developed hanging deltas and paucity of other strandline features seem inconsistent with the extent of lakes suggested by thick silts beneath extensive lacustrine plains.

The term dead-ice sink has genetic significance by simultaneously implying a condition of stagnation (e.g. dead-ice) and subsequent sediment accumulation (e.g. sink). It is a valley-floor feature characterized by an anomalously broad floodplain confined, upvalley and down, by valley train terraces. It originates in much the same way as a kettle, by collapse over buried ice. However, in this case, the buried ice is sufficiently massive to occupy the entire width of the valley, thus the dead-ice sink is a large-scale-kettle. Unlike a conventional kettle that forms a depression within a landform, the dead-ice sink is large enough to be a landform. Formation involves the initial detachment and burial of a large, remnant ice mass (several kilometers in diameter and upwards of 100 m thick) beneath an insulating veneer of outwash, which serves to retard melting during the deglacial processes (Figure 6). As pointed out by Mulholland, ice-block detachment may be a normal part of the backwasting mechanism. Should complete melting of the detached ice occur prior to retreat from the Appalachian Plateau, the developing dead-ice sink would be filled by outwash sediment as it develops, thus precluding the formation of a surface depression. However, buried ice masses that survive retreat from the Appalachian Plateau and the diversion of meltwater to the Mohawk drainage continue to slowly melt, leading to gradual subsidence and collapse without the addition of significant overlying sediment. The resulting valley-floor depression serves as a sediment sink for late glacial and post-glacial accumulation, which is reflected by subsurface data in the logs of water wells and test borings (Fleisher, 1986a).

#### **Non-through valley flow regime and landform assemblage**

Fleisher (1986) proposed stagnation of an entire non-through valleys ice-tongue by the progressive development of a negative ice budget due to restricted flow in thinning ice on headward divides. Figure 7 illustrates how ice-tongue starvation developed in non-through valleys, whereas active ice movement within through valleys (open-to-the-north) was sustained.

Non-through valleys contain a significantly different landform assemblage dominated by kame fields, isolated kames, and segments of discontinuous, remnant gravel plains, eskers, and dead-ice sinks, as noted along Otego Creek (Figure 8). The topographic expression of a kame field is similar to that of a kame moraine, but with significant differences. A kame field has limited lateral extent and is typically found to

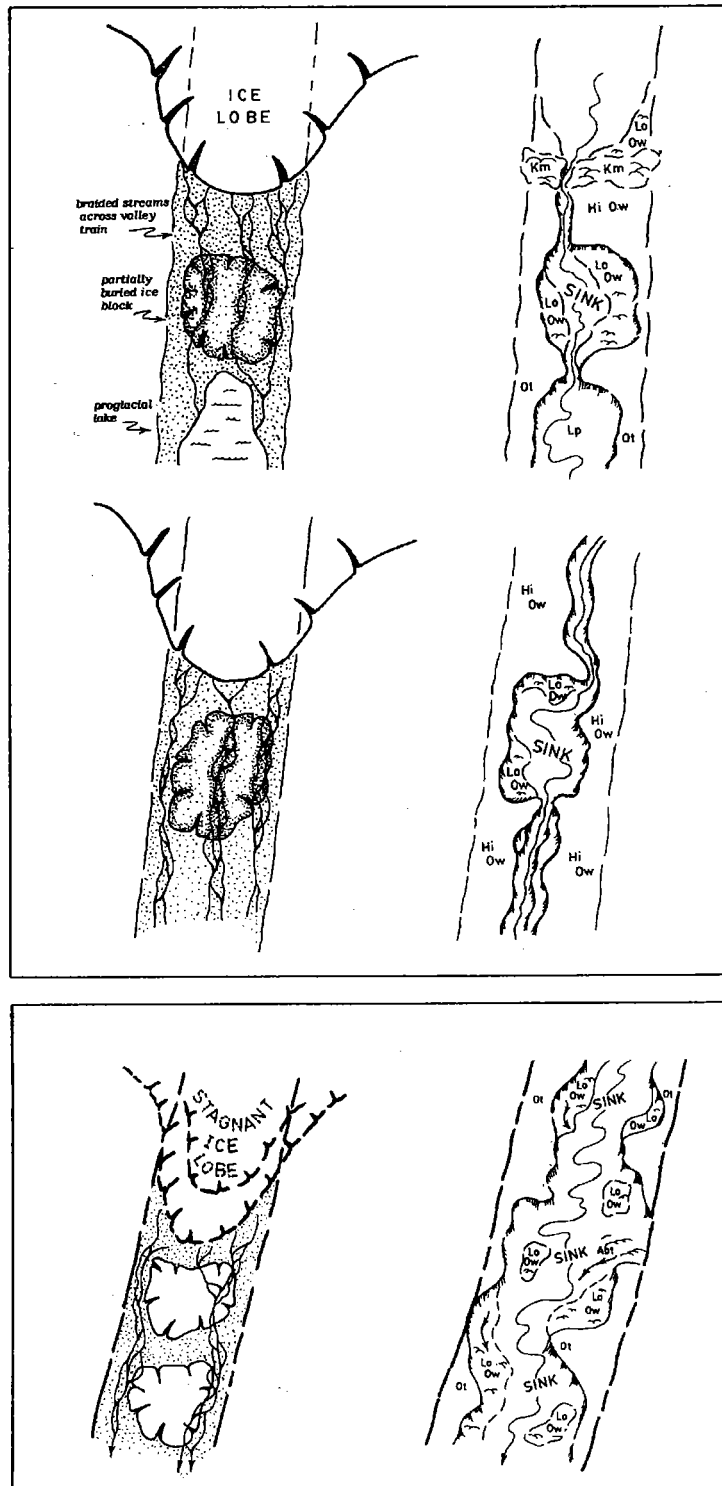


Figure 6. Development of dead-ice sinks.  
 A - active ice in through valley.  
 B - stagnant ice in non-through valley.

be lacking association with a valley train. Although kame fields are far more common in non-through valleys, they also occur in through valleys. They consist of poorly-sorted, silt-rich, matrix-supported gravel interstratified with lenses of diamict, well sorted gravel and sand beds, which are often disturbed by small scale collapse structures (Fleisher, 1984, 1986a). Kame fields are commonly found at the confluence of upland tributaries and main valleys, as are other landforms of ice-cored origin.

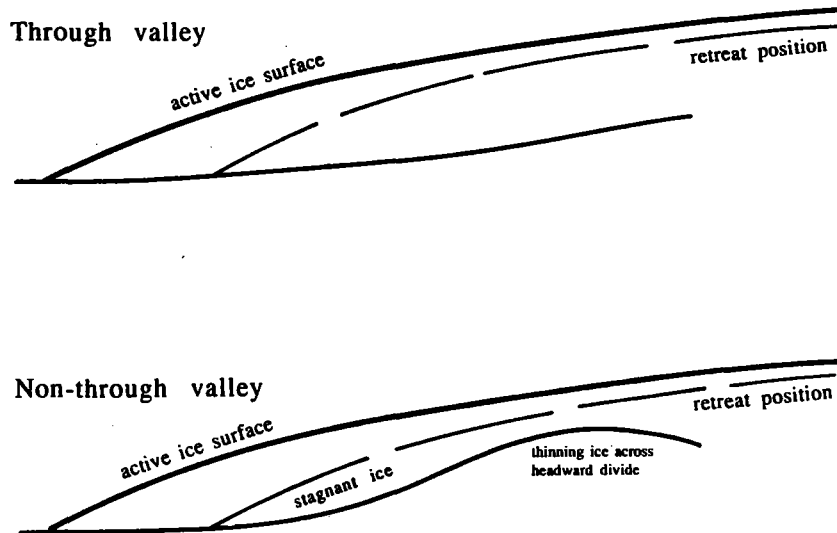


Figure 7. Schematic longitudinal profiles in through valley and non-through valley.

### Transverse valley landform assemblages

Valleys of this type are recognized by their transverse to sub-transverse orientation to the regional ice flow direction. They are prominent components of the regional stream pattern, but show no evidence of having been fed by active ice-tongues. Although some tend to be significantly smaller than through and non-through valleys, they have comparable bedrock relief and are significantly larger than upland tributaries.

Deglacial thinning over up-glacier divides would cause depleted nourishment and eventual separation of ice masses hundreds of meters thick in transverse valleys. Burial beneath outwash and inwash would retard melting and lead to the formation of various late glacial and early post-glacial landforms of ice-cored origin, without additional sedimentation by proglacial outwash.

Three different landform assemblages have been identified in transverse valleys. They are 1) kames, kame fields and limited gravel plain segments, amidst small lake plains (short-lived, local base level) and dead-ice sink complexes, as noted along the West Branch Delaware River between Delhi and Hobart, and the valley of Schenevus Creek. 2) large lake plains, small hanging deltas, alluvial fans and occasional kames have been mapped along Charlotte Creek between Davenport Center and Butts Corner, and 3) till plugs, isolated kames, small lake plains and discontinuous pitted terraces as may be seen in the valley of Ouleout Creek.

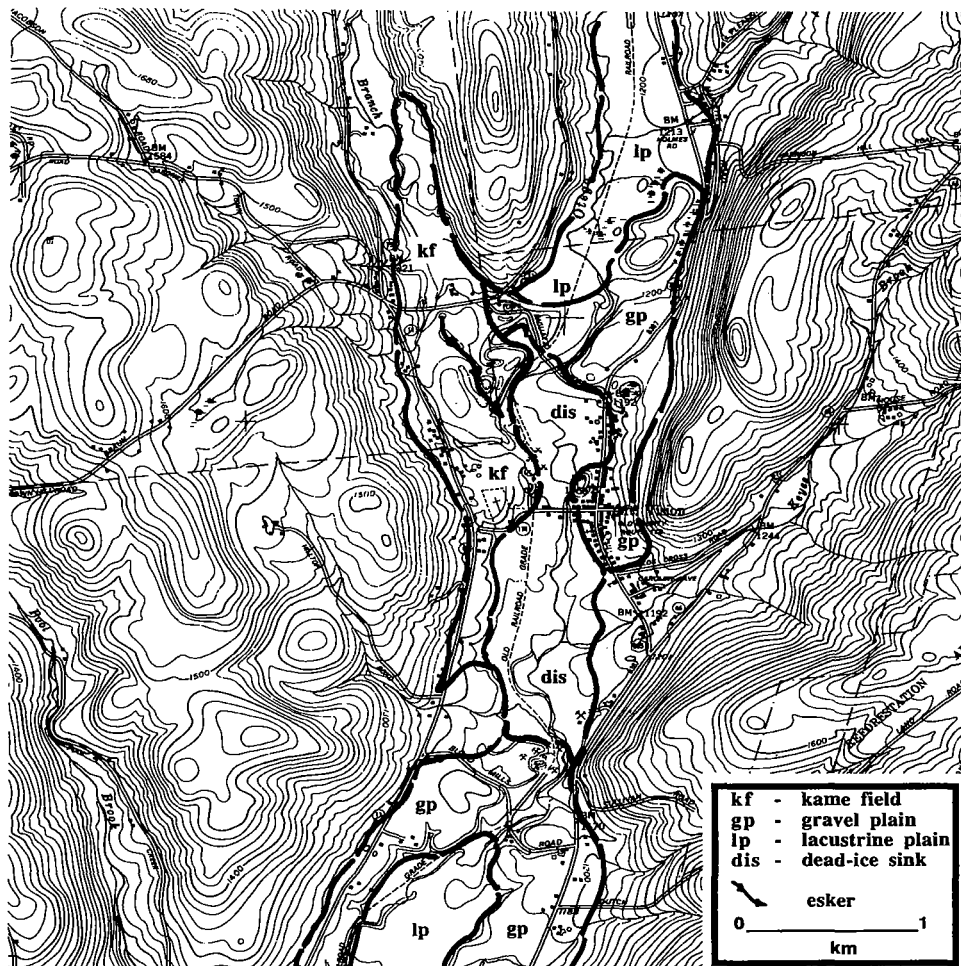


Figure 8. Non-through valley landform assemblage.

#### Pebble count Analysis - Method

Samples from borrow pits in kame terraces were taken by random selection of pebble and cobble-size clasts from multiple stratigraphic units at depth. Upland samples were obtained from hand-dug holes. After the samples were washed, screened, a representative number (in most cases more than one hundred) were broken, lithologic identification was made under a binocular microscope. The lack of a universally applied, standard sampling procedures may result in diverse pebble count data from multiple authors (MacClintock and Apfel, 1944; Merritt and Muller, 1959; Moss and Ritter, 1962; Coates, 1963; Denny and Lyford, 1963; Randall, 1978), thus limiting comprehensive interpretation. For example, in this study limestone and chert are technically not exotic materials because they are exposed in the headward reaches of

most through valleys. Yet, they are grouped with crystalline exotics because both indicate glacial transport from remote areas. Although the lack of uniform sampling and analytical procedures diminishes the value of pebble count data, they do lend themselves to basic generalizations.

### Summary of pebble count data

A summary of pebble count data arranged by sample site location appears in Table 1. Rock type categories represent contrasting lithologic suites. Percentage values vary broadly and include anomalous highs that tend to skew averages, but not beyond useful limits,

|                          |                  | Through Valleys     |                    |                   | Transverse Valley      |                        |
|--------------------------|------------------|---------------------|--------------------|-------------------|------------------------|------------------------|
|                          |                  | Unadilla Valley     | Susquehanna Valley | Oaks Creek Valley | Cherry Valley          | Charlotte Creek Valley |
| number of sites sampled* | (12)             | (12)                | (6)                | (7)               | (8)                    |                        |
|                          | %                | %                   | %                  | %                 |                        | %                      |
| local lithologies*       | 73.5             | 85.9                |                    | 51.5              | 80.0                   | 98.0                   |
| quartz-rich lithologies  | 9.0              | 1.1                 |                    | 4.8               | 0.8                    | —                      |
| limestone and chert      | 17.0             | 12.3                |                    | 40.6              | 17.5                   | 1.0                    |
| crystalline lithologies  | 0.5              | 0.7                 |                    | 3.1               | 1.7                    | 1.0                    |
|                          |                  | Non-through Valleys |                    | Uplands           |                        |                        |
|                          | Butternut Valley | Otego Creek Valley  |                    | Unadilla Upland   | Charlotte Creek Upland |                        |
| number of sites sampled* | (7)              | (6)                 |                    | (13)              | (9)                    |                        |
|                          | %                | %                   | %                  | %                 | %                      |                        |
| local lithologies*       | 95.9             | 93.1                |                    | 95.6              | 99.8                   |                        |
| quartz-rich lithologies  | 1.2              | 2.1                 |                    | 2.9               | —                      |                        |
| limestone and chert      | 2.0              | 1.5                 |                    | 1.3               | 0.1                    |                        |
| crystalline lithologies  | 0.9              | 3.3                 |                    | 0.0               | 0.1                    |                        |

# a minimum of 100 pebbles were counted at each sample site  
 \* includes shale, siltstone, mudstone, and graywacke

### **Generalizations derived from pebble count data**

Several generalization may be derived from these data.

1. Through valley stratified drift contains a conspicuously higher percentage of exotic lithologies than drift in all other locations. This is primarily due to the greater abundance of limestone/chert, and light colored crystalline rocks, thus giving rise to the term "bright" drift. This is in contrast to non-through valleys and uplands, where "drab" drift prevails due to the concentration of clasts from local sources.
2. Data from through valleys show greater variation than non-through valleys and uplands.
3. Drift within non-through valleys is only slightly less "drab" than upland drift, containing a significantly higher proportion of local lithologies than brighter, through valley drift.
4. Virtually all upland drift appears derived from local bedrock sources.
5. The lithologic suites of transverse and non-through valleys are similar to upland drift.

### **Pebble count anomalies**

Limestone outcrops at the heads of through valleys had a significant influence as a sediment source. This is demonstrated by exceptionally high pebble count values (27.2% and 56.4%) within 15-20 km of Onondaga Limestone outcrops in the valleys of Oaks Creek and the Susquehanna River. However, conspicuously fewer pebbles of limestone and chert in Cherry Valley suggest that factors other than distance from sediment source are also significant. Similar variations in pebble count data are also noted in Unadilla Valley near New Berlin. Five pebble counts from two sites depict a "bright" valley train (31.7% exotics) in the vicinity of New Berlin, whereas seven counts from the kame field at the mouth of Tallette Creek (a major tributary 6 km upvalley) are significantly less bright (8-16% exotics) (Yukinevicz Master's thesis). Local variations in data are also noted at the confluence of Oaks Creek and the Susquehanna River south of Cooperstown where exotics vary by 24% at two sample sites a few kilometers apart on the same moraine in adjacent valleys. Pebble count values for exotics in a few isolated locations along the Susquehanna River and Otego Creek are anomalously high. And, local lithologies in Charlotte Creek Valley (transverse valley) occur to the virtual exclusion of exotics.

### **Provenance**

Pebble count data from the Cooperstown area of the upper Susquehanna Valley provides evidence in support of glacial transport during kame moraine formation. Here, the Cassville-Cooperstown moraine, situated 5 km south of Cooperstown, contains 41.5 % limestone (Melia, 1975). The nearest limestone source is 20 km upvalley where outcrops of Onondaga Limestone are exposed at an elevation 100 m higher than the moraine and 175 m above the bedrock valley floor beneath the moraine. With the ice-tongue margin at the moraine and an assumed basal shear stress between 0.5 and 1.0 bar, the surface elevation of the glacier would be 250-400 m higher than the limestone outcrops, which precludes a superglacial source. Neither downvalley transport by meltwater flow nor post-glacial fluvial processes could account for these deposits because sediment movement through the Otsego Lake basin would not have been possible following retreat from the moraine. Two possible alternative are 1) limestone plucking followed by transport as basal load to the ice front where rising shears incorporated it in supraglacial debris at the moraine site or 2) transport by pressurized water moving through englacial and subglacial conduits exiting the glacier at the moraine site. The high concentration of limestone exotics in the Cassville-Cooperstown moraine not only indicates that bright drift originates from upvalley sources, as suggested by Moss and Ritter (1962), but also demonstrates the importance transportation of glacier sourced material within the ice.

Consistent with observations by Moss and Ritter south of the Valley Heads Moraine, data obtained from drift along Oaks Creek reveals a rapid downvalley decrease of limestone pebbles, which indicates that nearly all (97%) glacial transport is limited to distances of 10-15 km from the bedrock source, as suggested by Halter et al., (1984). However, 1-3 % exotics within upland drift

throughout the region (including crystallines that must have originated 40 to 100 km or more to the north) were either carried much farther or derived from re-worked older drift. Holmes (1952) relates distance of transport to lithologic resistance to abrasion and crushing, noting the virtual lack of shale pebbles and cobbles beyond 6-7 km of a bedrock source, whereas well lithified sandstones remain as conspicuous till components 130 km from their outcrops.

Additional sediment source information is derived from a kame field and associated remnants of a pitted surface that occupy the Unadilla Valley at the confluence of Tallette Creek, a major tributary 5 km north of New Berlin. The lithologic suite of Tallette Creek is typical of upland, drab drift, which is in sharp contrast with bright valley drift. Here, and at many tributary confluences throughout the region, the kame field is thought to have developed from downwasting of ice-cored inwash. The drab drift of Tallette Creek yields abruptly to bright valley drift at the kame field, which indicates its upvalley source. However, the normally high percentage of exotics found in two sites 6 km down valley appears diluted by drab inwash from Tallette Creek, which is consistent with an inwash origin for the kame field material.

Additional evidence in support of the inwash mechanism as a source of drift is noted in a hanging delta at the mouth of Kortright Creek, a primary upland tributary to Charlotte Creek. Here, collapsed topset and foreset beds and large kettles in the delta surface indicate a near-ice or ice-contact origin. In contrast with pebble count data from through valleys of the Susquehanna system, exotic lithologies are very sparse and occur no more frequently than in upland drift. This unique landform must consist of tributary inwash that was prograded across grounded ice in Glacial Lake Davenport (Fleisher, 1991b). An associated kame field, with a similar lithologic suite, could not have derived sediment from outwash sources in the Susquehanna and is also thought to be of inwash origin.

With the exception of valleys oriented semi-parallel to the ice margin, from which ice tongues would have been excluded, it appears that most coarse valley drift consists of different proportions of two basic components; 1) re-worked drab drift off the uplands and b) re-sedimented outwash from upvalley ice-tongue sources. The relative proportion of each would be a function of specific depositional conditions in each individual valley. For example, Cherry Valley data includes an abrupt contrast in limestone (from 34.4 to 8.1 %) in sample sites from opposite ends (upvalley and down) of a lacustrine plain. This suggests an interruption of glaciofluvial transport by a proglacial lake basin, similar to that proposed north of the Cassville-Cooperstown moraine. Likewise, anomalous data from another areas, such as Charlotte Creek valley, indicate that local processes had defining influence on the source of sediment.

In summary, pebble counts reveal that non-through valley drift consistently contains a lower percentage of exotics, which supports the notion that upland sources dominated. High exotic values are typical of bright drift thought to have been derived from upvalley glacial and outwash sources within through valleys. Very high values are restricted to areas within 10-15 km of a limestone bedrock source, but local exceptions include a few distant, isolated occurrences in both through and non-through valleys.

## **DISCUSSION**

### **Transportation mechanisms and sediment sources**

Fundamental to the modified ice-tongue model are 20 km long extensions of the ice front within all through valleys and large non-through valleys, with adjacent ice-free upland slopes. Under these conditions, Evenson and Clinch (1987) suggest that there are two fundamental sediment sources; 1) materials overridden and picked up by the glacier (glacier sourced) and 2) material brought to the glacier by fluvial and slope processes from adjacent ice-free upland sources (inwash sourced). The glacier is the transport agent for material incorporated as bed load, as well as that which accumulates as superglacial debris. All remaining material is transported by fluvial



processes within englacial and subglacial conduits (Gustavson and Boothroyd, 1988), subsequently accumulating in a variety of depositional environments at the glacier snout. Lawson (1979) maintains that only 5 % of glacially derived sediment is actually deposited from the ice. The remainder is attributed to a process called re-sedimentation involving transportation and deposition by various forms of mass movements and fluvial processes (Lawson, 1979; Evenson and Clinch, 1987).

Landform mapping and water well logs from the upper Susquehanna region indicate that sand and gravel are concentrated within moraines, kame fields, valley trains and kame terraces, whereas fine sand and silt occur beneath lacustrine plains. Although fine and coarse sediments interfinger laterally and are the result of single-stade deglaciation (Fleisher, 1987), they represent two distinctly different environments involving more than one transport mechanism and several possible sources. Although definitive conclusions may not be drawn solely from existing data, systematic and uniform differences in lithologic suites suggest that the dominant transport mechanisms in through valleys differ from those in non-through valleys. For example, active ice-tongue flow would favor glacial transport accompanied by high meltwater discharge. Mechanical attrition of poorly consolidated local, drab lithologies by glaciofluvial transport, as speculated by Holmes (1952), combined with enrichment of exotics from upvalley sources would brighten through valley drift.

However, in non-through valleys, reduced ice flow over northern divides led to ice-tongue collapse, local stagnation, diminished meltwater discharge, and reduced downvalley transport. In effect, non-through valleys were ultimately isolated from bright drift source areas. As deglaciation progressed, inwash from ice-free uplands became increasingly significant.

#### **Sources of Silt**

The silt-rich upland drift must be considered as a potential source of lacustrine silt found within most valley fill. The total volume of valley fines was calculated from well data and compared with the area of adjacent uplands to determine the degree of upland denudation that would have been required to supply an equivalent volume of silt. The main valleys of the entire upper Susquehanna contain approximately 14 cubic miles of fine sediment fill, which when distributed as a uniform blanket over the adjacent uplands (1720 square miles) would add an average of 16 m to the existing upland drift mantle. No field evidence exists to support this degree of general dissection. Furthermore, winnowing by upland runoff would have produced abundant residual boulder lag, which is also lacking. Yet uplands must have been subject to late glacial and early postglacial erosion as indicated by incision along primary tributaries. Although slope wash, mass wasting, and re-sedimentation by tributary inwash certainly account for some of the fines, most of the silt must have been derived from the only remaining source - the glacier.

Boulton and Hindmarsh (1987) argue that because fine-grained, deformable, subglacial sediment would not be capable of draining a constant meltwater influx, subglacial channels and conduits form to transmit excess water. As water discharge increases, so does the piezometric gradient, which in turn raises water pressure values within the sediment to equal the ice overburden pressure. As Boulton and Hindmarsh suggest, this leads to sediment liquefaction in the glacier terminal zone, which, in turn, causes a "flow of liquefied sediment into the proglacial environment". Liquefaction of fine sediment in combination with subglacial migration of the saturated substrate under pressure would force fines into the subglacial hydrologic system of conduits that exit at the glacier snout, and in this case directly into ice-contact lakes. This is proposed as the primary source and transport mechanism of lacustrine silt (see Figure 3).

#### **Sources of sand and gravel**

Well data indicate sand and gravel represent only 25 % of the total valley fill, the remainder being silt and quick sand. Although dark, coarse sands are known from exposures thought to be at depths near or at the spring high water table, most sand and gravel consists of interbedded bright sand and very coarse, clast-supported gravel. Based on pebble count data reported here and by

Moss and Ritter (1962), this drift is much too bright to have had an upland source, and is, therefore, thought to have originated from upvalley sources that ultimately derived debris directly from glacier ice.

Sedimentary structures (ripple marks, cross-bedding, graded-bedding, cut-and-fill) and particle properties (size, shape, sorting) indicate fluvial transport was fundamental to downvalley movement of bright sand and gravel. Aggradation in the form of deltaic kame terraces confirm discharge of meltwater streams directly into proglacial lakes via lateral channels along ice-tongue margins (see Figure 3).

Re-sedimentation of ice-cored glacial debris is seen in matrix-supported and poorly sorted silty, sand and gravel of moraines and kame fields. However, the limited occurrence of well developed kame moraines and associated valley train suggest few stable ice-marginal positions existed during retreat. Yet, bright drift in kame moraines must have been derived from upvalley bedrock sources and remote regions, which indicates glacial flow was an essential debris transport mechanism and re-sedimentation accompanied depositional processes. Although landform distribution suggests kame fields received inwash from upland sources, pebble count data show a strong affinity of bright drift to upvalley sources. Many valleys contain a few anomalous pebble counts that deviate from general trends, but the distinction between upland and valley drift is clear. In summary, the data suggest more than one sediment source for sand and gravel, and downvalley movement involved multiple transport mechanisms.

## **ANALOG ENVIRONMENTS OF SEDIMENTATION**

### **The Bering Glacier paradigm**

Conditions peripheral to Bering piedmont glacier, Alaska, include examples of depositional environments that existed during late glacial retreat in central New York State. Fed by the Bagley Ice Field, Bering Glacier spreads on a coastal lowland where it coalesces with the Steller Glacier. Combined, they form a 30 km wide piedmont lobe that fronts in several ice-contact lakes. Tsvat and Tsiu lakes basins flank Weeping Peat Island along the eastern sector (Figure 9). As a large, warm-based glacier it simulates lobate conditions along the retreating margin of the Laurentide Ice Sheet, and embraces environments of deposition analogous to conditions of ice tongue retreat from the Appalachian Plateau.

### **Suspended sediment**

A well established subglacial conduit system discharges, highly turbid meltwater directly into ice-contact lakes. Multi-year measurements of suspended sediment load and rates of sedimentation derived from annual bathymetric surveys and stratigraphic evidence provide a comprehensive data base for evaluating comparable conditions in the late Pleistocene glacial lakes of central New York State (Gardner, et al., 1993; Casamento, et al., 1997; Dell and Fleisher, 1998; Fleisher, et al., 1998; Fleisher, et al., 2000; Fleisher, et al., 2003). Furthermore, Bering Glacier is known to surge periodically (Post, 1972; Muller and Fleisher, 1995), most recently in 1993-95. Among the many abrupt and noteworthy changes brought about by the surge was a six-fold increase in suspended sediment (Figure 10). This is taken to indicate the closing of meltwater tunnels forcing broader meltwater distribution at the sole of the glacier and increased access to subglacial sediment. A mid surge outburst (jokulhlaup) marked the re-establishment of conduits to convey meltwater, thus turbidity began to return to pre-surge values.

## **Bathymetry**

Bathymetric surveys conducted during the 1991-2000 decade in Tsivat and Tsiu Lakes captured changes in lake basin morphology, including net accumulation of sediment from which rates of sedimentation are derived. Cumulative bathymetric changes include two pre-surge years (1991-1992), followed by two years involving the surge (1993-1995), and ending with four post-surge years (1996-2000). Distinctly different sedimentary conditions prevailed within each lake basin and from year-to-year, thus leading to relatively rapid changes in lake basin morphology. These changes are attributed to four main causes: 1) ice-front advance, 2) fluctuation in suspended sediment load, 3) migration of meltwater vents, and 4) retreat.

Excluding delta aggradation by heavily loaded inflowing subglacial streams, vertical settling is assumed to be the primary cause of sediment accumulation and bathymetry change. Therefore, changes in water depth detected from sequential bathymetric surveys are used as a proxy for the amount of accumulated sediment. Rates of sedimentation are interpreted from this information, as summarized in Table 2.

Table 2 contains a summary of processes that had an influence on the rate of sediment accumulation. Several relevant generalizations are: 1) uniform conditions did not prevail from year-to-year, 2) sedimentary processes within each lake changed with time, 3) rates of sediment accumulation due to vertical suspension settling vary with time, 4) rates of glaciolacustrine sedimentation related to vertical settling of suspended sediment increased significantly from 0.6-1.2 m yr<sup>-1</sup> prior to the surge to 3.1 - 3.3 m yr<sup>-1</sup> during the surge, where they remained for five years following the surge, 5) post-surge turbidity returned to pre-surge values, yet the rate of sedimentation remained high, thus suggesting adjustment related to the re-establishment of equilibrium conditions involving the redistribution of sediment already in the system.

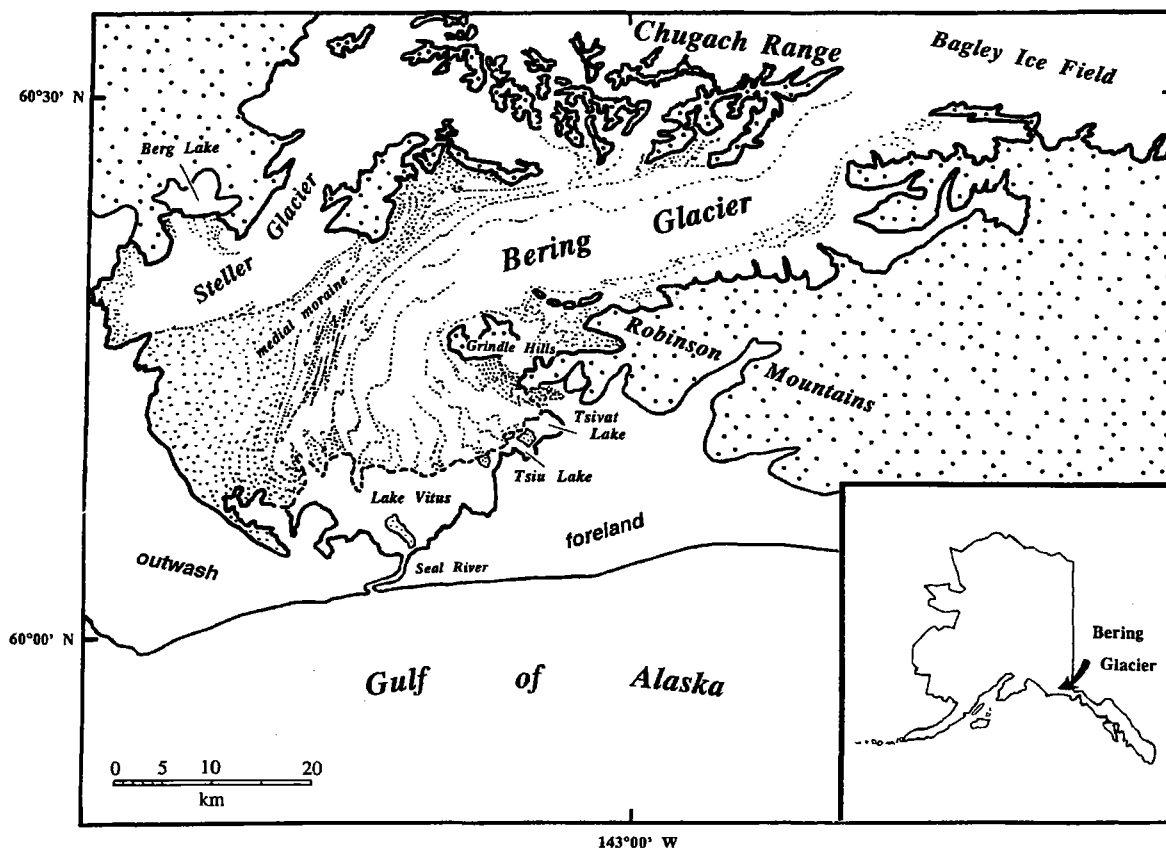
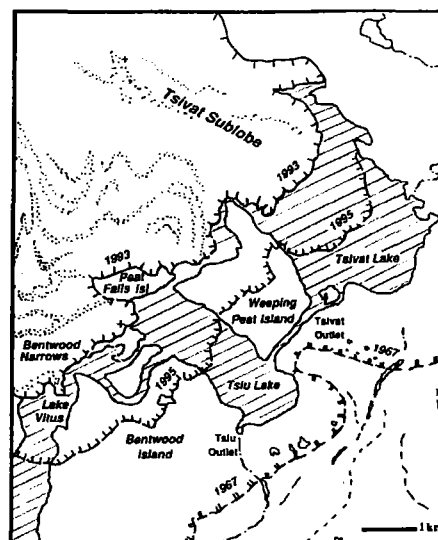


Figure 9. Bering Glacier location map and eastern sector lakes and islands.

Ice-contact Tsiat and Tsiu Lakes flank Weeping Peat Island.



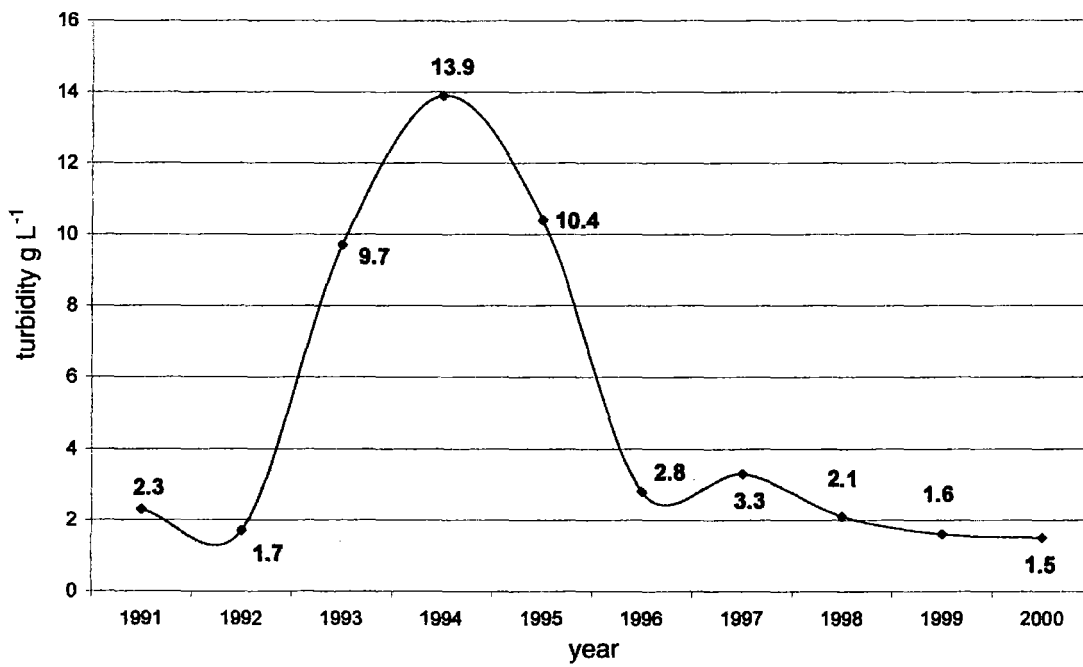


Figure 10. Average turbidity, Tsiu Lake, 1991-2000.

**TABLE 2. TIMING, EVENTS, PROCESSES AND RATES OF SEDIMENTATION**

|                              | <b>Timing of significant Events and processes</b>  | <b>Glaciolacustrine rates<br/>Suspension settling</b>                                    |
|------------------------------|--|--|
| Pre-surge<br>1990/91         | Upwelling vents on Tsivat ice front provide increased suspended sediment; also sediment from englacial and subglacial portals                          | 0.6 to 1.2m yr <sup>-1</sup> increased to 9.7 m yr <sup>-1</sup> with input delta growth |
| Early surge<br>1993          | Subglacial tunnels close; highly turbid leakage from base of advancing ice front; continued sediment settling  | No data  |
| Full surge<br>1994/95        | Outburst into sandur forms in Tsivat Lake basin; six-fold increase in turbidity; accelerated sediment settling; push moraine forms on Tsiu Lake floor. | 2.2 to 3.1 m yr <sup>-1</sup> average from 5-year total.                                 |
| Early post- surge<br>1996/97 | Outburst into Tsivat Lake basin; rapid growth of Tsiu Delta; reduction of basin volume; turbidity reduced  | 3.1-3.3 m yr <sup>-1</sup> with minimal input from ice front or delta growth.            |
| Late post-surge<br>1997/98   | Expansion of outburst sandur; sand fills Tsivat Lake basin; sediment bypassing to Tsiu Lake increases rate of Tsiu Delta growth.                       | Data from all sites affected by input from ice front and/or delta growth.                |
| Post-surge<br>1999/2000      | Tsiu Delta occupies 60% of Tsiu Lake basin. Growth rate exceeds rate of ice front retreat. Pervasive input by delta rate growth.                       | 3.0 m yr <sup>-1</sup> from single site.   |

### **Stratigraphic evidence**

A semi-continuous aerial photo record (1978 to 1991) documents retreat positions of the eastern Bering piedmont lobe and the development of a shallow embayment of Tsiu Lake on Weeping Peat Island during the summer of 1986. An observed breakout in 1989 abruptly dropped lake level 17 m causing water to withdraw from the embayment, thereby exposing the net accumulation of three years of lacustrine sedimentation that accumulated within 100 m of the ice front. Three observation trenches were excavated to expose a short stratigraphic column containing three annual couplets with a total thickness of 54 cm. This continuous record of accumulation during a documented 3-year period yields an average, net accumulation rate of 18 cm/year. However, turbidity flows and undercurrents known to be common in this type of environment may have on occasion interrupted sedimentation by scouring, as indicated by notable diastems. Therefore, varve thickness here is taken to represent minimum values for annual accumulation.

The lower portion of each annual couplet averages 11.5 cm in thickness and consists of fine, gray silt laminae, each approximately 0.7 mm thick. These grade upward into interlaminated light and dark gray silt and tan, very fine sand beds, each about 0.5 cm thick, with an average cumulative thickness of 6.5 cm. The sand is commonly cross bedded and contains subtle graded bedding. Because the uppermost interlaminated unit was the last to be deposited prior to the breakout, it must represent summer-season sediments. This interpretation is consistent with intermittent higher summer discharge and associated currents that periodically introduce sand and hold silt in suspension. Lower energy conditions beneath frozen lakes during winter months favor quieter water and yield a thicker accumulation of uniform silt.

### **Application to Susquehanna Drainage**

Applying these elevated rates of sedimentation to similar deposits beneath the Susquehanna River floodplain, and other central New York valleys, has interesting implications.

For example, within the valleys of the Susquehanna Valley it is common to find thick lake silts (100-125 meters) associated with sand and gravel terraces. Bathymetric information from ice-contact lakes at Bering Glacier indicate rates of sediment accumulation on the order of 0.6 to 1.2 m/year. Applying these rates to the Susquehanna Valley suggests that lakes at the front of the retreating Laurentide Ice Sheet would have been relatively short-lived, existing for approximately 100 to 200 years. Using an average annual rate of 0.18 m/year derived from measured varves formed at the ice front on Weeping Peat Island, the duration of accumulation would have been approximately 700 years. Regardless of which analog rate is used, it is amply clear that Late Pleistocene lakes on the Appalachian Plateau persisted for a much shorter period of time than might be implied by applied annual rates based on varves a few centimeters thick.

These conditions changed significantly as the ice pulled back from the northern drainage divide of the plateau and meltwater was diverted from the Susquehanna to the Mohawk Valley (Ridge et al., 1991). Subsequently, reduced fluvial discharge, fed only by meteoric runoff, carrying a severely diminished sediment load to remaining lake basins. Such was the case in Otsego Lake where Holocene deposition was determined by Yuretich (1981) to have been less than 8 m. The eventual failure of most lake dams and events associated with subsequent release of water are virtually unknown.

#### **SUMMARY**

1. The margin of retreating Laurentide ice consisted of valley ice tongues 20 km long and adjacent ice-free uplands.
2. The landform assemblage produced by active ice backwasting in through valleys differs from that in non-through valleys, where stagnation and downwasting occurred.
3. The sediment sources, transport mechanisms and formative processes for moraines and kame fields are different. Kame moraines mark the position of active ice margins, whereas kame fields indicate the concentration of inwash on stagnant valley ice and downwasting.
4. The primary source of lacustrine silt, which constitutes 75 % of all stratified drift, was subglacial meltwater flow through conduits that discharged saturated, fine sediment directly into proglacial, ice-contact lakes. Pebble count data indicate that most outwash sand and gravel was originally transported by glacier movement from upvalley bedrock sources, then reworked by meltwater flow at or near the ice margin.
5. The exotic pebble content of through valley drift is primarily responsible for its bright appearance, whereas inwash-derived drab drift from upland sources is a significant component of non-through valley deposits.
6. The through valley landform assemblage and subsurface stratigraphy have been interpreted to depict rapid retreat of active ice tongues from a proglacial, ice-contact lake environment. Such conditions would favor subglacial conduit discharge of highly turbid water and subsequent rapid lacustrine deposition. In addition, surface meltwater streams entering lakes from positions lateral to valley ice tongues would account for active deltaic aggradation. Indeed, well data indicate the interfingering of delta foresets with penecontemporaneous lake sediments, which means that lacustrine sedimentation and subaerial deposition were synchronous, and both occurred very rapidly. Analog conditions of rapid deposition at the margin of Bering Glacier, Alaska, suggest that the accumulation of glaciolacustrine fine sand, silt and clay may have been on the order of 0.6 to 1.2 m/year. At these rates, the average glaciolacustrine sequence of 100-125 m would have accumulated in just 100 to 200 years. At a rate of 0.4 m/year, as derived from exposed varves, the duration of accumulation may have been 700 years, which is much more rapid than varves of conventional thickness would suggest. Consequently, lakes in the Susquehanna Valley during Laurentide retreat were relatively short-lived.

7. Remnant ice and associated dead-ice sedimentation were common during regional deglaciation. Entire ice tongues stagnated and downwasted in non-through valleys due to restricted flow across headward divides. Local stagnation of detached ice masses is recognized to have accompanied backwasting in through valleys.

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As is customary for NYSGA field trips, a Road Log of planned stops is prepared months in advance. However, the ephemeral nature of Quaternary exposures generally leads to changes made only days prior to the trip. Therefore, the stops described below may not correspond to stops made during this field trip.

## ROAD LOG

This Road Log begins at the I-88, Rt. 23, Rt. 7 interchange at Exit 13, west of Oneonta.

| <u>Miles from last point</u> | <u>Cumulative Miles</u> |   |
|------------------------------|-------------------------|---|
| 0.0                          | 0.0                     | Proceed west on I-88 from Exit 13. I-88 parallels the Susquehanna River for the next 2.4 miles. Active floodplain aggradation mantles lake sediments of Glacial Lake Otego that was dammed during retreat by the Wells Bridge Moraine, our first stop.  |
| 5.6                          | 5.6                     | The highway rises above the valley floor and provides a good view of the modern flood plain and the abrupt change in valley trend that is a remnant of a preglacial engrown meander. A bedrock promontory on the horizon to the right (north) protrudes into the valley along the inside of the meander bend. |
| 0.2                          | 5.8                     | STOP 3 is on the left (south), but we won't stop now. Access to this area is possible from County Rd. 48 (locally referred to as the Otego-Wells Bridge Rd.), which we will take on our return to the Oneonta area from Wells Bridge.   |
| 3.3                          | 9.1                     | Continue west on I-88 past Rt. 7 & Otego exit.  |
| 1.3                          | 10.4                    | Good view to the west of the valley plug formed by the Wells Bridge Moraine.  |
| 1.7                          | 12.1                    | View to the right (north) across the valley includes the back of the Wells Bridge Moraine and an associated kame terrace. The next 1.2 miles provides excellent overviews of the moraine and the gap cut by the Susquehanna River.  |
| 1.2                          | 13.3                    | Rest Area Exit.   |

**STOP 1. Wells Bridge Moraine (Franklin and Unadilla Quadrangles).** The hummocky relief of this moraine is common for other moraines in the upper Susquehanna drainage basin. This moraine completely blocked the valley following glacier retreat thus forming the dam for Glacial Lake Otego that extended about 25 km upvalley to Oneonta. Logs of water wells from

the floor of the valley penetrate more than 400 feet of silt without encountering bedrock. Rates of sedimentation from similar ice-contact lakes at Bering Glacier, Alaska, related to vertical settling of suspended sediment between 0.6 and 1.2 m/yr. Therefore, Glacial Lake Otego may have been relatively short-lived.

The configuration of the Laurentide ice front during retreat is commonly thought to have consisted of valley ice tongues. Early work portrayed relatively short ice tongues, where as more recent evidence suggests that they may have reached 20 km in length. This raises implications of sediment source areas and transport mechanisms, while opening the topic of mechanisms of moraine formation. Considering that most moraines in the region consist of stratified drift and lack topographic expression other than in valleys, alternatives to conventional moraine formation may be entertained.

The Wells Bridge Moraine is assumed to have been emplaced about 15,000 years BP and breached about 14,000 years BP. We will consider the field evidence for an 1140 feet lake level at stop 3. Field work in the Unadilla and Sidney areas indicates that the Upper Susquehanna Lake Chain has greater downvalley extent than will be covered in this road log.

|     |      |  |
|-----|------|--|
|     |      | Return to I-88.  |
| 1.5 | 14.8 | Cross Ouleout Creek.   |
| 0.5 | 15.3 | Leave I-88 at exit for N.Y. 357, Franklin and Unadilla. Turn right on Rt. 357-West.  |
| 1.2 | 16.5 | Cross Susquehanna River and turn right on Rt. 7- East. Highway parallels the river for one mile.   |
| 1.7 | 18.2 | Railroad overpass. Highway climbs onto outwash terrace near mouth of Sand Hill Creek.  |
| 1.8 | 20.0 | Highway drops into Sand Hill Creek incision of outwash and immediately climbs to follow the contact of the Wells Bridge Moraine and outwash .  |
| 0.6 | 20.6 | Crest of moraine on the left, breach on the right. It was within the breach of this moraine that archeological excavations uncovered charcoal in silt covering a buried river point bar. This was dated at 13,000 to 14,500 years BP, thus providing limiting age of Lake Otego.   |
| 0.4 | 21.0 | Village of Wells Bridge. Turn right (south), cross Susquehanna and turn left (east) at the end of the bridge on Otego-Wells Bridge Rd., which becomes Otsego County Rd. 48. Road parallels river for 0.6 miles before rising onto an outwash terrace. A correlative terrace can be seen across the valley to the north at an elevation of about 1140 feet. |

- |     |      |  |
|-----|------|--|
| 3.0 | 24.0 | Cross over I-88. White house across the valley to the north is situated on a terrace at about 1120 feet. Other planar landforms can be found along the valley that suggest a second level for Lake Otego below 1140 feet. Entering Otego Quadrangle. |
| 0.6 | 24.6 | Turn right into parking lot of Gus's Diner.  |

**STOP 2. Massive red sandstone outcrop on south side of parking lot is covered with glacial abrasional features on joint faces that parallel the valley wall. Note that striae and grooves are inclined in the downvalley direction, thus indicating topographically controlled ice flow consistent with valley ice tongue model. Basal flow of overriding ice would not have produced such features. Several joint block surfaces that do not parallel the valley trend are also polished, but not by ice flow. These very smooth and polished surfaces are the product of abrasion by highly turbid water flow and support the notion of a highly charged subglacial hydrologic system.**

- |     |      |   |
|-----|------|---|
| 0.7 | 24.7 | Access to I-88 on left. Continue straight ahead.  |
| 0.6 | 25.3 | Fork in road, bear left.  |
| 0.3 | 25.6 | Pass under I-88.  |
| 0.1 | 25.7 | Intersection at end of bridge, turn right remaining on County Rd. 48.   |
| 0.2 | 25.9 | Gravel excavation on right contained deltaic foreset and topset beds indicating a downvalley current direction.                           |
| 0.2 | 26.1 | Similar exposure in excavation on the left.   |
| 0.6 | 26.7 | Road drops to modern flood plain, which is superimposed on Lake Otego lacustrine plain. Lacustrine plain continues to the left and right. |
| 1.1 | 27.8 | Kame on the right. Prior to construction of I-88 a bit smaller feature could also be seen to the northeast.                               |
| 0.2 | 28.0 | Stop is situated to the right, across I-88, on the lower valley wall below "treeline". Proceed to I-88 overpass.                          |
| 0.5 | 28.5 |   |

**STOP 3. Park on the right beyond the overpass and walk south between I-88 and the forested slope. About 300 meters south of overpass and upslope from the fence, at an elevation of approximately 1120-1140 feet (2/3 the way up the forest-free slope), pebbly coarse sand, fine sand, silt, and a few clay seams were exposed during construction. Fluvial channel structures with small scale foreset beds inclined into the slope and down valley rested**

upon finely laminated, rippled and cross bedded silt and fine sand. These sands are interpreted to have formed along the strandline of Lake Otego by wave generated currents that moved into this valley wall alcove. The sand about 1140 feet lacks pebbles, is considered to be of eolian origin, blown up slope from the beach.

|     |      |   |
|-----|------|---|
| 1.3 | 29.8 | Continue east on County Rd. 48.<br>Road drops back down to the lacustrine plain at an elevation of 1060 feet. Entering Oneonta Quadrangle.  |
| 1.5 | 31.3 | Turn left on access to I-88, cross river and I-88, and proceed to intersection with Rt. 7. This is where the Road Log began.  |
| 0.5 | 31.8 | Go straight through traffic light at Rt. 7 intersection. The highway traverses an outwash/alluvial bench between 1100 and 1120 feet at the confluence of Otego Creek and the Susquehanna River. |
| 0.8 | 32.6 | Continue straight through traffic light.  |
| 0.3 | 32.9 | Junction of Rt. 23 from the right. Continue straight on Rt. 205-23.   |

**Take note that the Road Log mileage tally starts anew at the I - 88, Exit 13 intersection with State Routes 7 and 205, West Oneonta, so that this segment of the trip may be run independent of the preceding Road Log.**

| <u>Miles from last point</u> | <u>Cumulative Miles</u> |   |
|------------------------------|-------------------------|---|
| 0                            | 0                       | Intersection of Rt. 7 and 205, proceed north on Rt. 205.  |
| 1.1                          | 1.1                     | Traffic light at Rt. 23 / 205 intersection.   |
| .7                           | 1.8                     | Bear left on Rt. 23.  |
| 9.2                          | 12.9                    | Intersection of Rt. 23 and 51 in Village of Morris; continue straight through intersection on County Rt. 13 to New Berlin |
| 7.2                          | 20.1                    | Dead-ice sink on right  |
| .8                           | 20.9                    | Unadilla River  |
| .2                           | 21.1                    | Intersection of County Rt. 13 and Rt. 8. Turn sharply left (south) on Rt. 8   |
| .4                           | 21.5                    | Pull off right into gravel quarry operation, that has been inactive for several years                                     |

**STOP 4 - Through valley landforms (New Berlin South Quadrangle)- high terrace (1180') consisting of deltaic valley train marking the ice margin (no moraine) at the head of which is a dead-ice sink. Although excavation in this quarry has been inactive for several years, large-scale, deltaic foresets**

are still expressed through the colluvium. The nearest base level to which these deposits may be graded is 22 km downvalley at Rock Wells Mills (Guilford Quadrangle), where kame and kettle, morainic topography may have served as a drainage plug in a constricted segment of the valley. A partially preserved hummocky, kame terrace at White Store (16 km downvalley) also suggests ice burial.

Earlier exposures here included interbedded, very coarse, clast supported boulder gravel and pebbly, coarse sand. A carbonate cement in some units is related to limestone clast weathering, leaching, and reprecipitation at depth, thus indicating the "bright" nature of this outwash.

The Unadilla River is incised within a valley train and flows from a dead-ice sink upvalley to a lacustrine plain downvalley. These landforms and their internal features are indicative of active ice retreat with glaciofluvial sediment transport into an ice-contact lake. Partial burial of a detached, large remnant ice mass caused retarded melting until after meltwater sedimentation ceased, thus allowing collapse and topographic preservation of a dead-ice sink. Water well logs within the sink verify the lack of subsurface, stratigraphic continuity.

Turn around; proceed north on Rt. 8 through Village of New Berlin

|     |      |   |
|-----|------|---|
| .7  | 22.2 | Intersection of Rt. 8 and 80 (traffic light), proceed north on Rt. 8 and 80.  |
| 1.3 | 23.5 | Continue north on Rt. 8   |
| .5  | 24.0 | Road rises on gravel terrace above lacustrine plain   |
| 1.4 | 25.4 | A high-level terrace at 1280' to 1300' on west side of valley continues for 1.5 miles west of South Edmeston. It appears to be an older kame terrace with no apparent downvalley base level, thus suggesting ice served that purpose. |
| 2.2 | 27.6 | Road rises on pitted planar gravel and kame field   |
| .9  | 28.5 | Columbus Quarters at intersection of Rt. 8 and Chenango County Rt. 41.  |

**STOP 5 - Inwash sediment source and topographic expression of kame field, Columbus Quarter (New Berlin North Quadrangle).** The kame field and associated pitted plain may have served as a local, temporarily dam (1220') for a lake upvalley. Pebble counts from upland "drab" drift now resting on the "bright" drift of the kame field suggest tributary inwash from Tallette Creek as a source of sediment (see plot of pebble count data, after Yuchniewicz, 1996).

Turn around, proceed south on Rt. 8.

|     |      |   |
|-----|------|---|
| 1.9 | 30.4 | At Lambs Corners turn left (east) onto Chenango County Rt. 255 (which changes to Otsego County Rt. 20 at Unadilla River), proceed east across lacustrine plain.                             |
| 4.2 | 34.6 | Road descends onto the pitted and discontinuous valley train of Butternut Creek.  |
| .6  | 35.2 | Intersection of County Rt. 20 and Rt. 80, proceed east on Rt. 80.   |
| 3.0 | 38.2 | Rt. 51 enters from the right, continue east on Rt. 80.  |
| .4  | 39.3 | Rt. 51 turns left, continue east on Rt. 80.   |
| 6.7 | 46.0 | Rt. 205 enters from the right, continue east on Rt. 80 and 205.   |
| 1.7 | 47.7 | Road descends onto the kame and kettle topography of the Oaksville Moraine.   |
| .2  | 47.9 | Intersection with Rt. 28, turn right, proceed south on Rt. 80 and 28.   |
| .5  | 48.4 | Road crosses the crest of the moraine and continues on and off the moraine for next 1.5 miles through the villages of Oaksville and Fly Creek.  |
| 1.7 | 50.1 | Turn right (south) in Fly Creek on County Rt. 26, then bear left at fork.   |
| 1.2 | 51.3 | Quarry operation in sand and gravel valley train with classic assemblage of glaciofluvial sedimentary structures on the left (east). The road continues along the floor of a dead-ice sink. |
| 2.3 | 53.6 | Pull off on right shoulder, park and walk to crest of moraine.  |

**STOP 6 - Cassville-Cooperstown moraine, at Index (5 km south of Cooperstown) (Cooperstown Quadrangle). This classic through valley landform assemblage (a moraine linked with an extensive valley train) represents active ice deposition. Old excavations in a nearby borrow pit exposed stratified sand and gravel. Test borings indicate that the gravel of the moraine interfingers with silt upvalley, toward Cooperstown, before yielding entirely to silt (120+ feet thick, Fleisher, 1992). The classic kame and kettle expression of this moraine loses all topographic expression as it rises to the uplands, as is characteristic of moraines in central New York State. This has implications for a moraine forming mechanism that would involve sediment transport dominated by glaciofluvial processes. Based on topographic criteria, Krall (1977) correlates the Cassville-Cooperstown Moraine with moraines in the Hudson Valley, thereby suggesting a readvance circa 14,000 years BP. Ridge, et al., (1991) makes a tentative correlation with the West Canada Readvance of Late Wisconsinan, pre-Valley Heads age (circa 15.5 ka) based on elevation projections of ice-marginal deposits in the Mohawk Valley and the limit of readvance inferred**



from subsurface stratigraphy (Fleisher, 1986).

Although high enough, this moraine did not dam Glacial Lake Cooperstown, although landforms and stratigraphy suggest that a proto-Lake Cooperstown may have extended to the moraine by "swamping" a dead-ice sink

|     |      |  |
|-----|------|--|
| .2  | 53.8 | Proceed eastward over the moraine to the intersection with Rt. 28, turn left (north).                            |
| .2  | 54.0 | Dead-ice sink between Cassville-Cooperstown Moraine and Cooperstown occupies the valley floor for the next mile. |
| 1.7 | 55.7 | Junction Rt. 80 and 28, proceed east on Rt. 80.  |
| .3  | 56.0 | Traffic light intersection with Main St., Cooperstown; continue straight through intersection east on Rt. 80.    |
| .1  | 56.1 | Stop sign, junction Lake St. with Rt. 80, turn right onto Lake St.   |
| .2  | 56.3 | Park at Intersection of Lake and River St.; stairway to Council Rock to the left.                                |

**STOP 7 - Lake Front Park on Doubleday Ice Margin (Cooperstown Quadrangle). Well data from the crest of the moraine (Bassett Hospital) indicate that the morainic dam for Otsego Lake consists of 180 feet of bouldery silt. Subtle hummocky terrain in the village of Cooperstown may be traced northwestward, where it is well expressed in a golf course at the lake shore. Hanging deltas along the western lake shore and lake clay on the northwestern shore indicate that while the moraine dammed Glacial Lake Cooperstown at an elevation of 1250'. The view northward includes a conspicuous cross-sectional valley asymmetry, which played an important role in determining the primary sediment source for Glacial Lake Cooperstown.**

**A seismic reflection survey shows that Otsego Lake occupies a rock basin that has been eroded as much as 132 m (433') lower than present lake level and infilled with up to 88 m (289') of sediment. Sediment thickness is greatest in the southern half of the lake basin and thins by ~50% to the north where maximum water depths occur. As an alternative to the traditional view of deglaciation, which depicts continuous backwasting of an active ice margin with short valley lobes, subsurface and land-based evidence suggests backwasting retreat of 20 km long ice-tongues that remained in the valleys, subject to collapse while adjacent upland regions were ice-free. Proglacial sediments were largely transported to the south end of the lake basin as alluvium from ice-free western tributary streams (Fleisher, et al., 1992).**

Return to Lake Street, turn right (west)

|    |      |   |
|----|------|---|
| .5 | 56.8 | Turn left onto Chestnut St. (Rt. 28 and Rt. 80 West). |
|----|------|---|

|     |      |  |
|-----|------|--|
| .5  | 57.3 | Junction Rt. 80 west, proceed south on Rt. 28.   |
| 2.3 | 59.6 | Highway traverses Cassville-Cooperstown moraine.   |
| 1.1 | 60.7 | Hamlet of Hyde Park; from here south for several miles, planar gravels appear related to upland tributary inwash sources.  |
| 5.4 | 66.1 | Junction Rt. 166 at village of Milford (traffic light), continue south on Rt. 28. Valley floor consists of lacustrine clays from here south to a moraine at Portlandville at 68.9. |
| 6.0 | 72.1 | Turn right on Clover Hill Road in Milford Center and park  |

**STOP 8. Goodyear Lake overview at Milford Center (Milford Quadrangle). Pitted valley train and dead-ice sink. These landforms indicate local stagnation of the valley ice-tongue during active retreat. However, the valley fill consists of 60+ ft. of ice-contact, sand and gravel outwash over 300+ ft. of silt interpreted to be of lacustrine origin. Does the stratigraphy indicate two stades (readvance) or is there a single-stade environment that accounts for both stratigraphic units? Aggradation over ground ice islands is suggested.**

Continue south on Rt. 28

|     |      |  |
|-----|------|--|
| 1.8 | 73.9 | Junction with Rt. 7, continue south on Rt. 28 to I-88 interchange.   |
| .8  | 74.7 | Rt. 7 overpass, dead-ice sink right and left.  |
| .5  | 75.2 | Pass under I-88 and continue straight on Gersoni Road (unmarked)   |
| 1.0 | 76.2 | Stop sign, turn right on Dead End road leading to Seward's Gravel Quarry.  |
| .4  | 76.6 | Approval for entrance to quarry must be obtained in Office. Mileage into, around and out of quarry not included in road log. |

**Stop 9. Moraine at margin of Dead Ice Sink (West Davenport Quadrangle). Many of the landforms that characterize the through valley assemblage may be found in the vicinity of this location. They include a kame moraine and associated outwash features, dead ice sink, and lacustrine plain. During the many years of operation, a variety of ice-contact features have been exposed. Upper- most units are flow tills (debris flows) that randomly interfinger with interbedded sand and medium gravel, frequently including silt and sand of local significance. Blocks of lodgment till have been observed in piles with sub-meter boulders that come off screening equipment. Massive foreset beds of gravel that range in size from bouldery gravel, through cobble and**

pebble gravel, to pebbly, coarse sand are commonly found beneath the overlying flow till at an intermediate, vertical position through the operation. Well sorted, laminated lake sand is very common on the downvalley, lower margin of the quarry.

The hummocky, kame and kettle topography common to the quarry area fades downvalley to join with the surface of extensively developed outwash terraces. Over the years, quarry operations in these terraces have uniformly exposed massive gravel foreset beds indicative of meltwater sedimentation in an ice-contact lake that expanded upvalley to follow the retreating ice front. MacNish and Randell (1982) describe such a setting to depict slow, active-ice retreat.

These extensive paired terraces formed as "lateral deltas" into the expanding ice-contact lake, thus growing in the upvalley direction as space became available during retreat. Episodes of slow retreat would accommodate growth until delta terraces from either side of the valley coalesced to fill the valley, thus forming a valley train with internal deltaic structure.

|     |      |  |
|-----|------|--|
| .4  | 77.0 | Leave the quarry and return to the stop sign intersection. Turn right (west) onto Hemlock Road (and unmarked continuation of Gersoni Road) |
| .4  | 77.4 | Entrance to Broe Pit, Cobbelskill Products quarry. Time permitting, we will drive in to view massive foresets of delta terrace.            |
| 1.2 | 78.6 | Stop sign intersection with County Rt. 47. Turn left   |
| 1.2 | 79.8 | Delaware County line (Otsego County Rt. 47 becomes Delaware County Rt. 11).  |
| .9  | 80.7 | Turn left in Davenport Center and continue past the Post Office on left. Road parallels Charlotte Creek through the moraine on the right.  |
| 1.9 | 82.7 | Turn right into parking lot for Hartwick College's Pine Lake Camp. Walk downhill to pavilion.  |

**STOP 10 -Davenport Center. Pine Lake Dead-ice Sink Complex. These kames and sinks formed in an ice-cored terrain at the mouth of Charlotte Creek where this transverse valley was clogged by remnant ice. These landforms are typical of valleys in which the stagnated ice was subject to inwash burial and many local ponds and lakes existed. Here, the Davenport Moraine dammed Charlotte Creek valley at 1280 ft. Many shoreline landforms along the valley are graded to this lake.**

Return to road, turn right.

|    |      |  |
|----|------|--|
| .3 | 83.0 | Right hand fork goes downhill and across the lacustrine plain of Glacial Lake Davenport. |
|----|------|--|

|     |      |  |
|-----|------|--|
| .6  | 83.6 | Turn left at intersection with Rt. 23 in Davenport Center.   |
| .2  | 83.8 | Town of Davenport Center quarry on right (no longer active). Past exposures contained well-developed, large-scale deltaic foreset beds graded to 1280 ft. water level. Forty foot deep kettles indicate progradation that incorporated grounded ice. |
| 3.7 | 87.5 | Town of Davenport. Proceed east on Rt. 23.   |
| 1.5 | 89.0 | Clark Company Stone Products. Mileage into, around and out of quarry not included in road log.   |

**Stop 11. Hanging delta at the confluence with Middle Brook (Davenport Quadrangle). Although quarrying activity continues, much of this original hanging delta into Glacial Lake Devenport (1280 to 1300 ft.) has been excavated. Remnants of evidence supporting this interpretation may still be seen. Exposed at the upper quarry level are several meters of well stratified, strongly imbricated topset gravel. Discontinuous exposure to lower positions eventually lead to multiple packages of lacustrine sand and silt interspersed with small-scale deltaic infill and cut and fill structures.**

Return to Rt. 23.

|     |      |   |
|-----|------|---|
| .2  | 89.2 | Turn left onto Delaware County Rt. 9 to Fergusonville.  |
| .3  | 89.5 | Road crosses lacustrine plain   |
| 1.0 | 90.5 | Entering Fergusonville.   |
| .7  | 91.2 | Turn right on Olive Branch Road (Dead End) at Brandow's Trailer Sales. Access by permission of Brandow Family. Mileage into, around and out of quarry not included in road log. |

**Stop 12. Very localized hummocky topography reaching elevations of 1360 ft. Recent excavations exposed several, 2-3 m thick, alternating units of cross bedded coarse sand, tan laminated lake sand, and sand and fine gravel, each with unconformable contacts. The entire sequence was overlain by 2-3 m of moderately well sorted pebble gravel. Isolated from potential tributary sources, this ice-contact landform appears to have originated as meltwater infill of localized ponding on ice-cored terrain, much as would be anticipated in contact with a remnant ice masses in a transverse valley.**

|     |       |   |
|-----|-------|---|
| 2.2 | 93.4  | Backtrack to Rt. 23.<br>Turn right (west) onto Rt. 23.              |
| 1.9 | 100.9 | Highway traverses moraine at West Davenport.                        |
| 2.3 | 103.2 | Terrain to the left is what may be the only lateral moraine in this |

area.

- |     |       |  |
|-----|-------|--|
| .7  | 103.9 | Road descends to Susquehanna valley lacustrine plain.  |
| 1.1 | 105.0 | Highway enters the breach of the Oneonta Moraine (significantly altered by urban development). |
| 1.3 | 106.3 | Traffic light intersection I-88, 28 N/S, 23W; continue straight on 28 south.                   |
| .6  | 106.9 | Turn right toward I-88 west.   |
| .3  | 107.2 | Turn left onto I-88 west.  |
| 2.0 | 109.2 | Take Exit 13 (Morris and Route 205).   |
| .3  | 109.5 | Turn right (north) onto Route 205.   |
| .2  | 109.7 | This brings you back to the start of this trip at intersection of Rt. 7 and Rt. 205.           |

**END OF ROAD LOG**

## OPTIONAL - field trip up the non-through valley of Otego Creek.

| Miles from<br>last point | Cumulative<br>Miles |   |
|--------------------------|---------------------|---|
| 0                        | 0                   | Start at traffic light intersection of Rt. 23 west (to West Oneonta and Morris) and Rt. 205 (Laurens) north, proceed north on Rt. 205.            |
| 4.0                      | 4.0                 | Winnie Hill Road joins Rt. 205 from the right. Pull over and walk up Winnie Hill Road 0.1 mile for downvalley view south across local kame field. |
| .7                       | 4.7                 | Turn left on County Rt. 11A (unmarked) toward Laurens.  |
| .3                       | 5.0                 | Turn left on County Rt. 11 through Village of Laurens   |
| .4                       | 5.4                 | Bear left at fork (sign to Oneonta).  |
| .8                       | 5.8                 | Turn left into Town of Laurens Quarry adjacent to Maple Grove Cemetery. Mileage into, around and out of quarry not included in road log.          |

**OPTIONAL STOP 1. Village of Laurens Quarry (Vision Quadrangle). This is a good example of inwash sourced aggradation associated with a stagnant, downwasting ice tongue that occupied this non-through valley and other like it. At various times, excavation has exposed tilted and truncated sequences of interstratified gravel and pebbly sand with fluvial sedimentary structures and local silt layers of lacustrine origin. The basic feature is an ice-contact, hanging delta (Mt. Vision Quadrangle). Landform expression and large-scale foreset and topset beds indicate deposition into an ice-contact lake. Initial aggradation here was into a lake graded to a moraine dam at West Oneonta, 5 km downvalley. A local inwash source provided adequate sediment for additional topset, alluvial aggradation above lake level. Otego Creek passes through a narrow breach that interrupts the continuation of this landform at the same elevation across the valley floor. The localized concentration of accumulated sediment here is attributed to inwash from Lake Brook. Similarly, other tributaries fed sediment from ice-free slope to local lakes and ponds distributed along the valley and gravel plains are graded to these water bodies.**

Return to road, turn right (north), backtrack through Village of Laurens.

|     |      |  |
|-----|------|--|
| .8  | 6.6  | Junction 11A and 11, continue north on Maple Street.   |
| 2.1 | 8.7  | Gravel plain remnant for next half mile  |
| 1.3 | 10.0 | Junction Rts. 11 and 11B, continue straight on 11. Road rises onto pitted planar gravel remnant. |
| 1.0 | 11.0 | Junction Rts. 11 and 15; bear right continuing on Rt. 11, road                                   |

traverses terrain of ice-cored origin including a small esker out of sight to the right.

- .1      11.1      Park on right shoulder and walk through the pasture to the crest of an esker on the near, right horizon.

**OPTIONAL STOP 2. Esker in dead-ice terrain (Mt. Vision Quadrangle).** Once again we see concentrated aggradation at a stream confluence, here where West Branch joins Otego Creek. Landforms typical of a non-through valley assemblage include a kame field (differs from a kame moraine only in that the kame and kettle topography is not linked with planar outwash, as would be anticipated at an active ice margin), gravel plain, dead-ice sink, and an esker. Eskers are rarely preserved, partly because conditions favoring formation were not common, but also because they are excellent sources of sand and gravel, thus they are typically removed by gravel mining. Here, the esker is indicative of stagnant ice conditions. Lack of excavation precludes investigation of internal features.

Proceed on Rt. 11.

- .2      11.3      Turn right onto Angel Road that traverses kame field.
- .6      11.9      Junction Angel Road and Rt. 205; turn left (north) onto Rt. 205 toward Hartwick.
- .6      12.5      Pull off on left shoulder.

**OPTIONAL STOP 3. Overview of meltwater channel through discontinuous gravel plain remnant and terrain of ice-cored origin.**

- .5      13.0      Road descends to a local lacustrine plain (dead-ice sink?)
- .7      13.7      Gravel excavation to the left removed a landform thought to be an esker.
- 2.5      16.2      County Rt. 45 enters from the right, pull off on right shoulder.

**OPTIONAL STOP 4. Overview of discontinuous, planar gravel surface and terrain of ice-cored origin bordering a local, upvalley lacustrine plain.**

Proceed north on Rt. 205.

- 1.2      17.4      Village of Hartwick, Junction Rt. 205 and County Rt. 11, proceed north on Rt. 205.
- 1.3      18.7      Pull off on right shoulder.

**OPTIONAL STOP 5. Overview of headward area in non-through valley. Valley floor contains a few isolated kames. Unlike through valleys, the valley floor gradually rises to upland slopes across which ice flow diminished due to thinning during glacial retreat, thus resulting in downvalley ice-tongue**

**starvation and ultimate stagnation.**

**END OF ROAD LOG**

Continue on Rt. 205 (north) 4 miles to intersection with Rt. 80 or return on Rt. 205 (south) 16.5 miles to the Junction of Rts. 23 and 205 where this optional trip began.



## ***TRIP A4***

### **KARST OF THE SCHOHARIE VALLEY, NEW YORK**

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### **INTRODUCTION**

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

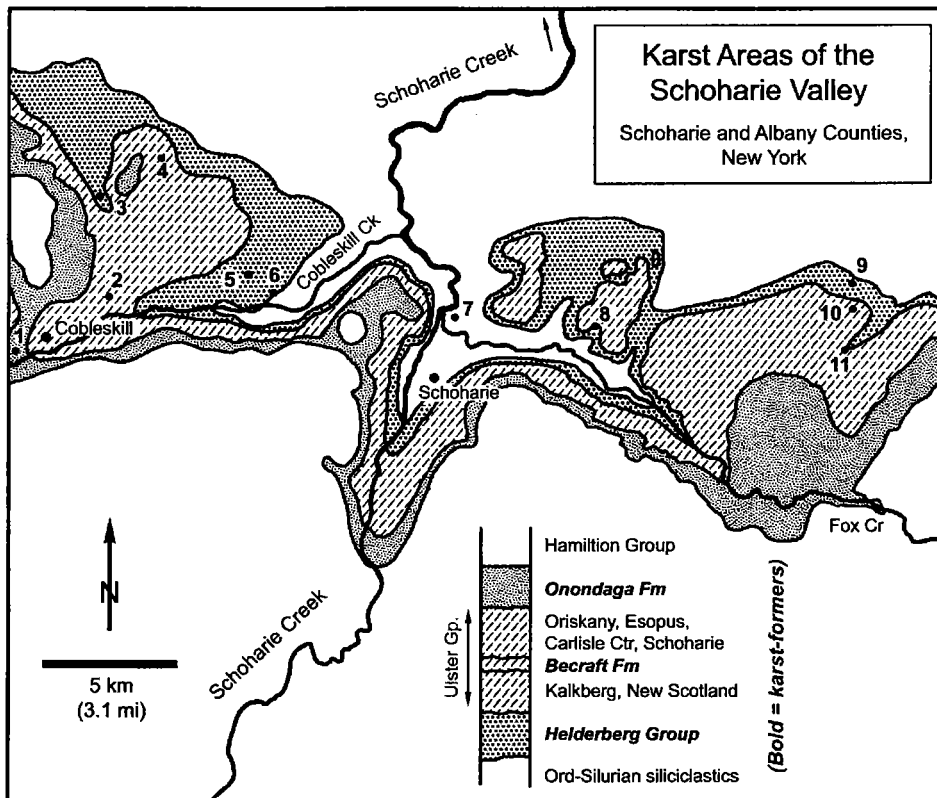
This field trip illustrates some of the major surface karst features of New York, including both surface karst and caves. Three different plateaus are visited, each with a different geomorphic setting and subsurface drainage pattern. No special equipment is needed for this field trip. However, many of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

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

## GEOLOGIC SETTING

### Stratigraphy

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



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

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

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

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

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

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

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

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

### **Geologic structure**

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

### **Geomorphology and Glacial Geology**

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

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

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

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

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

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

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

## KARST PRINCIPLES

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

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

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

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

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

## Effects of Glaciation on Karst

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

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

## Points to Ponder

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

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

## Recent Clues

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

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

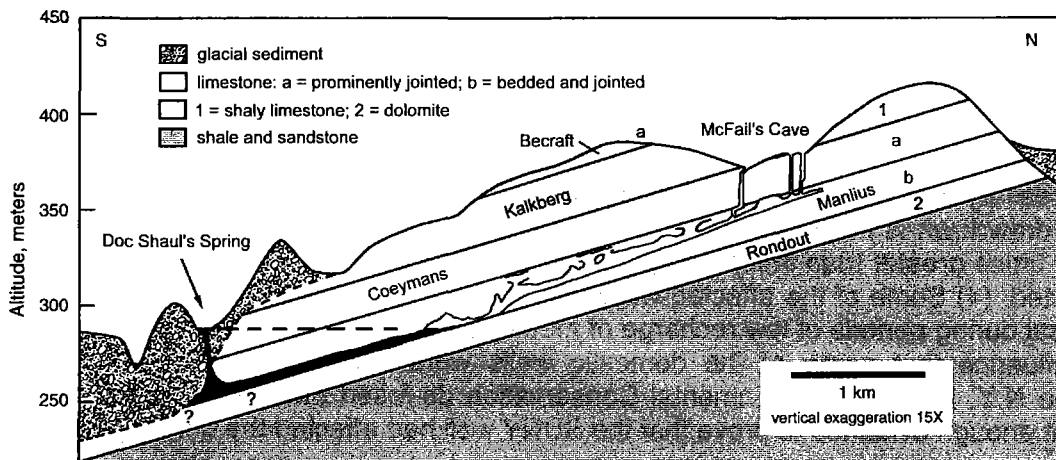
## **DESCRIPTION OF FIELD-TRIP STOPS**

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

### **Cobleskill Plateau**

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





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

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

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

### STOP 1 – WELL FIELD ON CAMPUS OF SUNY COBLESKILL

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

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

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

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

## **STOP 2 – DOC SHAUL’S SPRING**

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

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



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

### **STOP 3 – BROWN'S DEPRESSION**

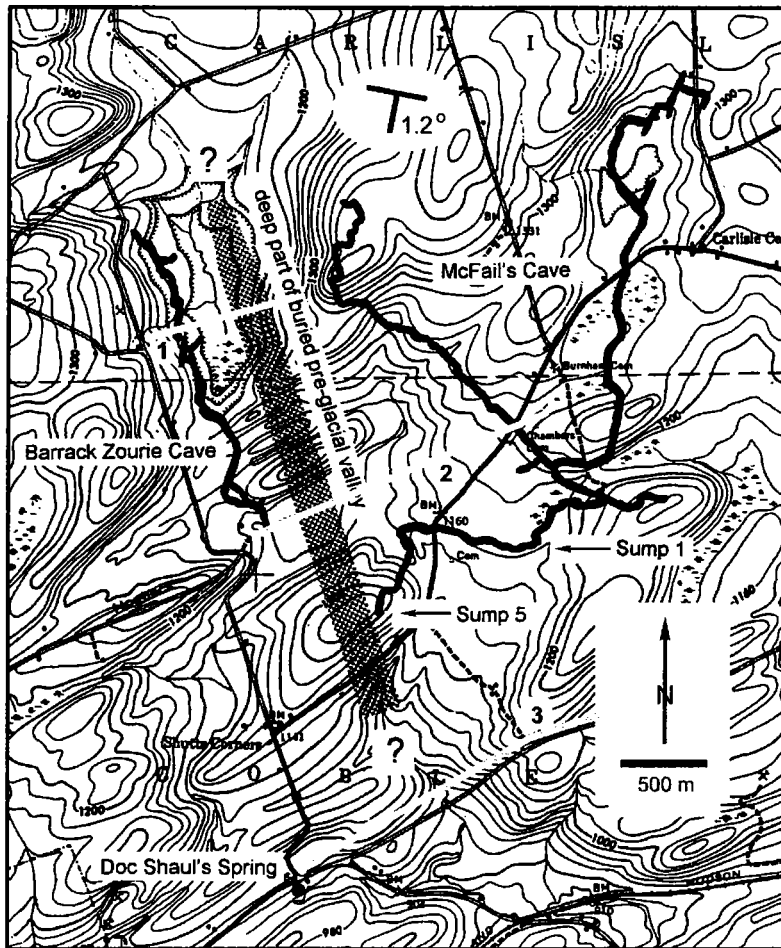
Because of the difficulty of parking and land access, this "stop" actually consists of short views from the bus at several locations, accompanied by geologic discussions. Most of the depression is on private property where visitors are not welcome.

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

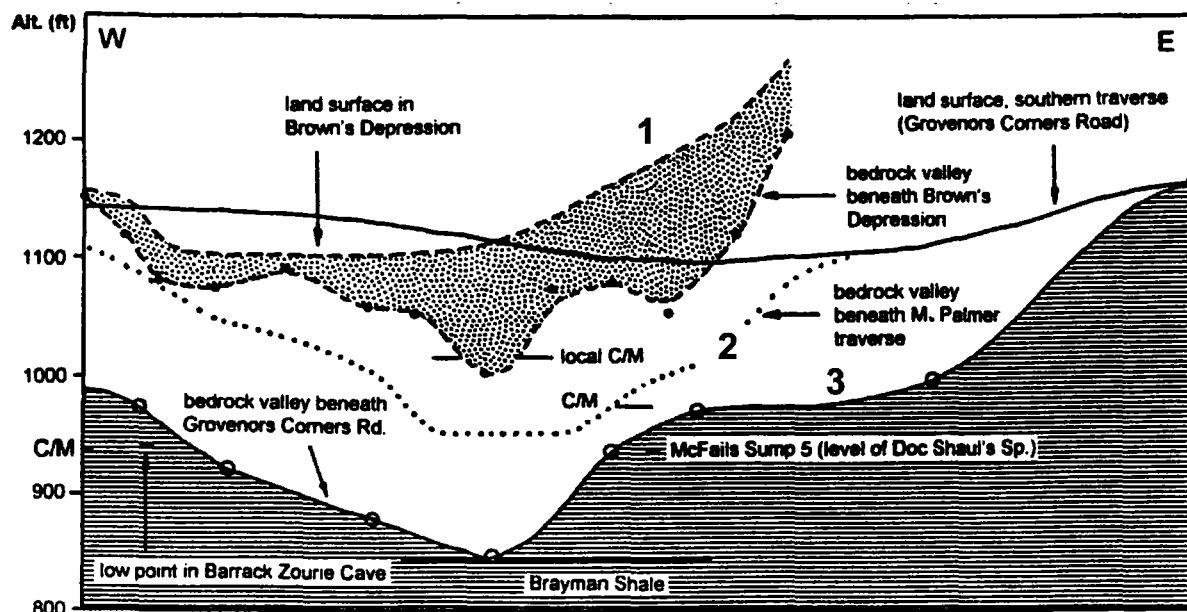
Surface drainage was deranged by glaciation in many parts of the Cobleskill Plateau. Prior to glaciation, a prominent north-south stream valley extended through the western part of the plateau, reaching downward through almost the entire Helderberg Group in places. Deflection of glacial ice by the Helderberg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

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

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



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



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

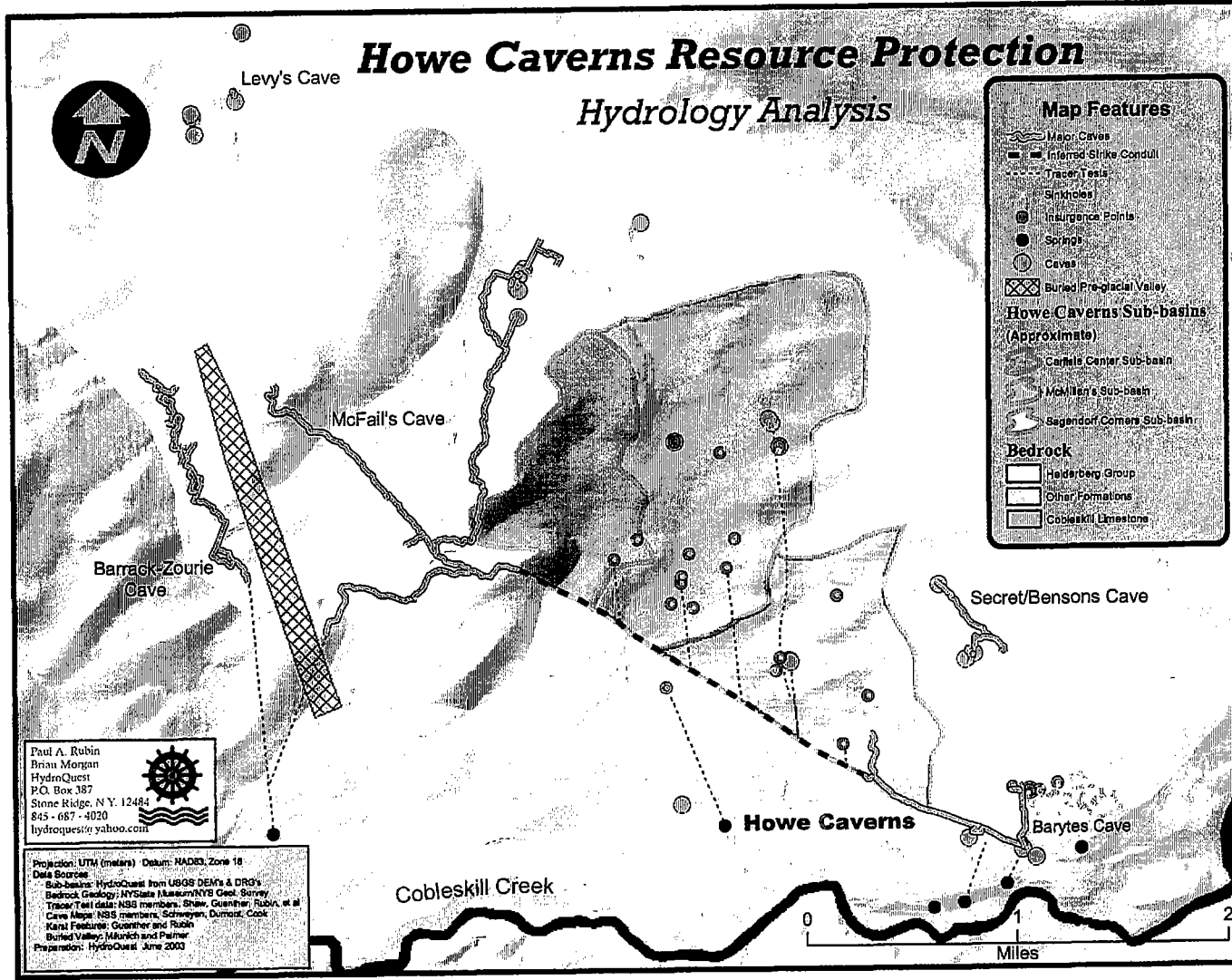


Figure 7. Howe Caverns Resource Protection hydrology analysis. Source: HydroQuest. Funding provided by Howe Caverns, Inc.

#### **STOP 4 -- MCFAIL'S CAVE ENTRANCES**

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

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

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

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

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



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

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

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





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

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

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

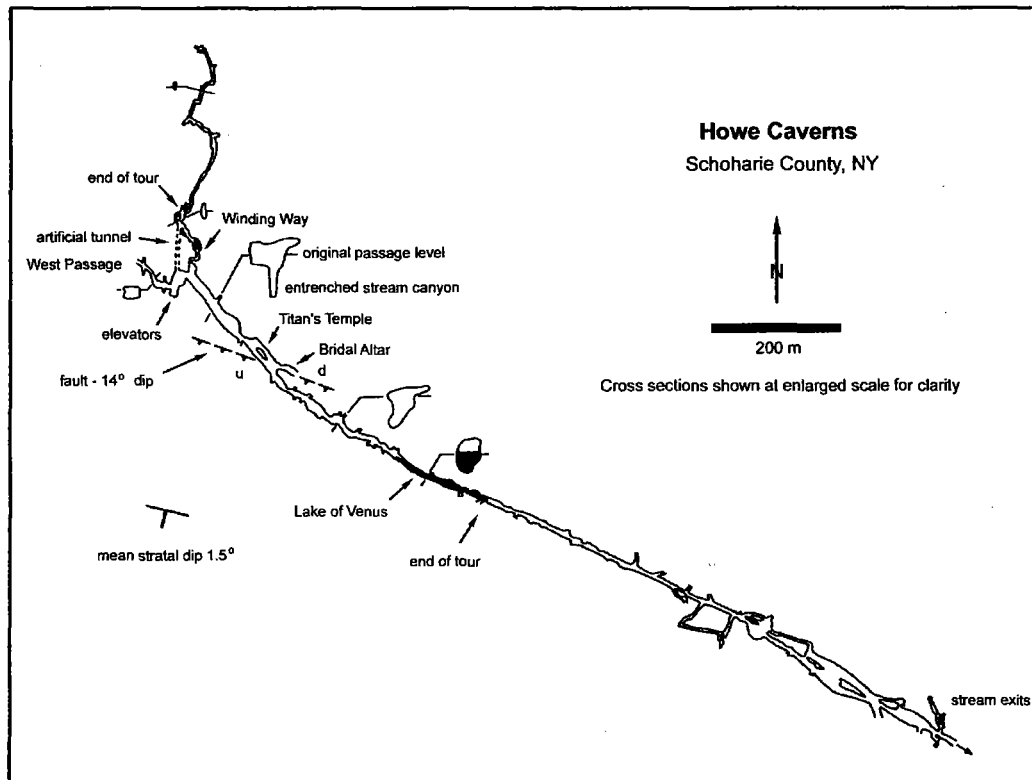
Despite the great amount of surface runoff, the water entering the cave is not particularly aggressive. In fact, during most of the year the cave water is slightly supersaturated with calcite

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

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

## **STOP 5 – HOWE CAVERNS**

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



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

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

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

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

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



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

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

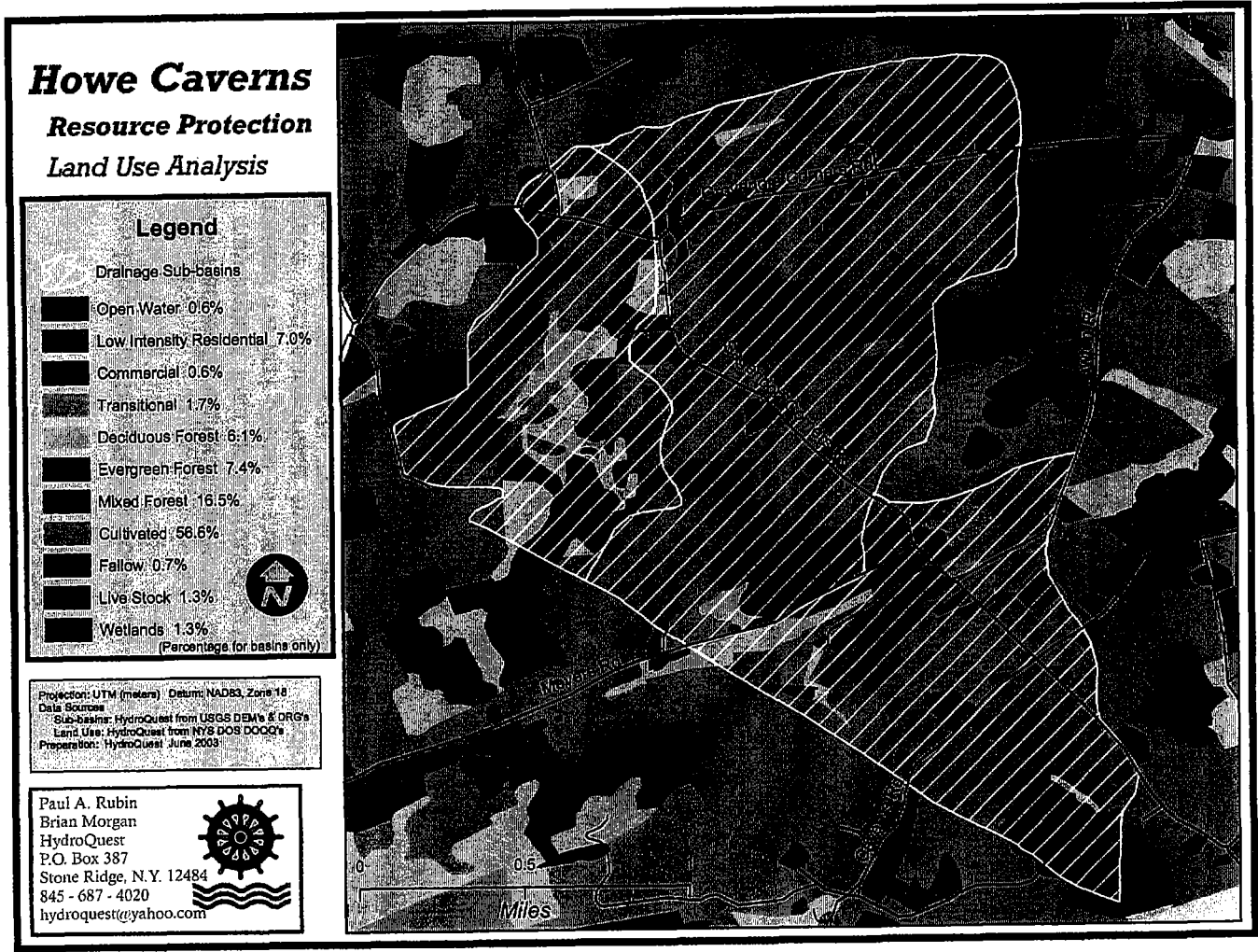
The tours exit the cave through an artificial tunnel between the Winding Way and the elevators. Much of this tunnel is excavated through a fault zone marked by small slickensided exposures and related calcite mineralization.

## **Howe Caverns Karst Hydrology and Cave Resource Protection**

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

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

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



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

## STOP 6 – HOWE CAVE QUARRY

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

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

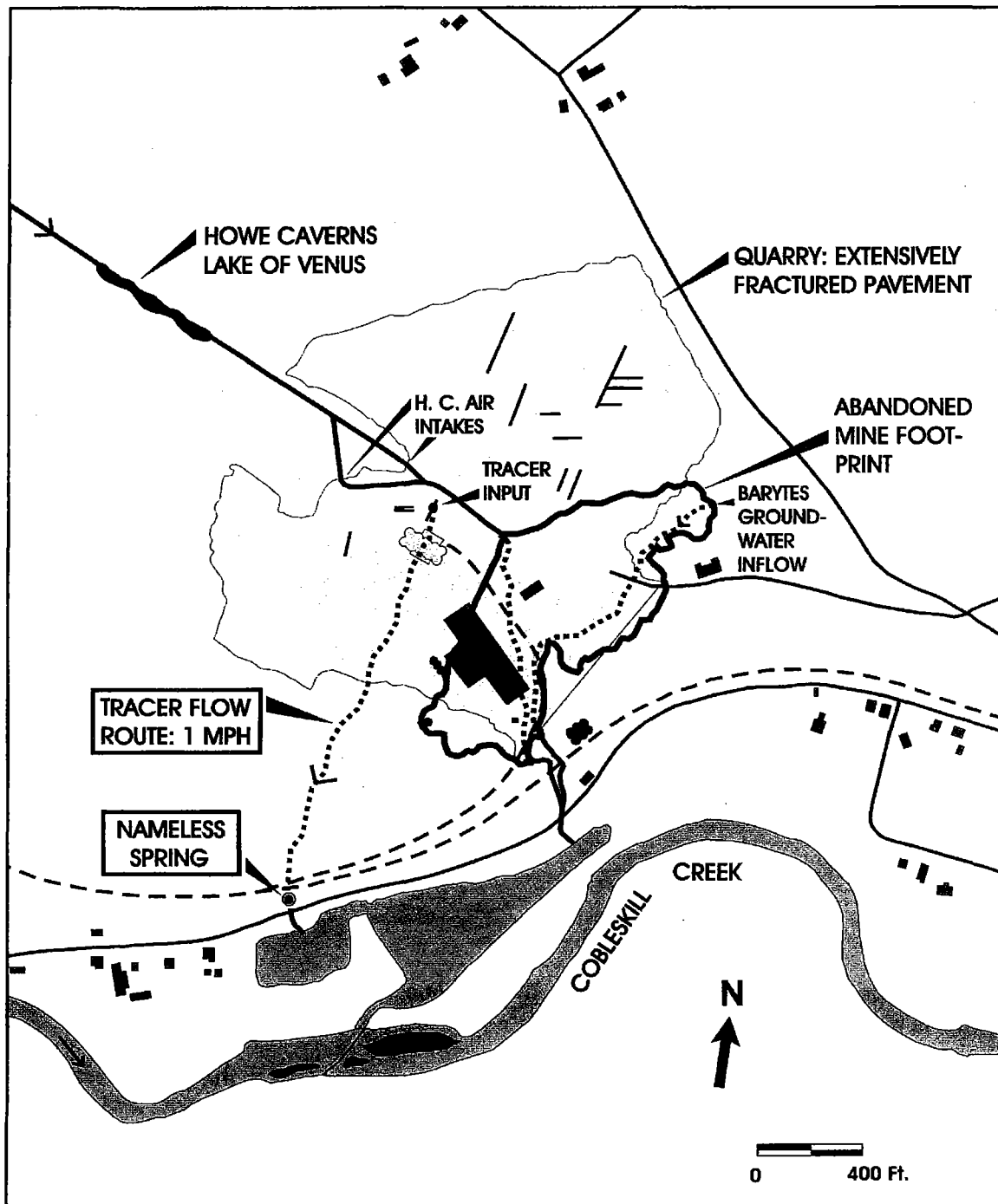


**Figure 13:** Howe Cave Quarry is located in the Manlius and Coeymans Limestones. A small thrust fault is visible in the wall.

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

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

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



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



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

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

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

## **STOP 7 – TERRACE MOUNTAIN**

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

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

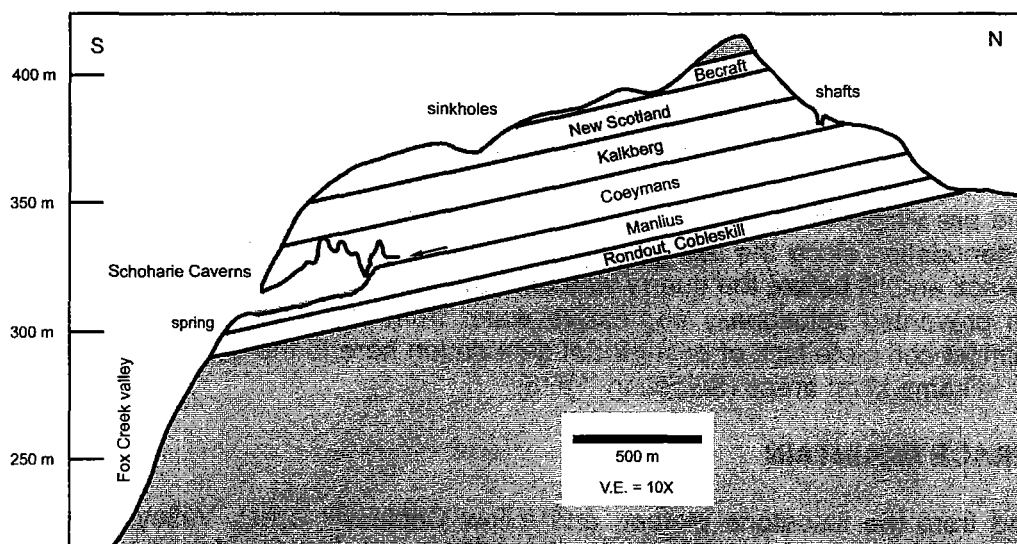
Although Terrace Mountain might seem an ideal location for karst features, today it does not have a catchment area large enough to supply recharge to more than a few shafts, sinkholes, and narrow abrasive caves, all of them small. The largest cave is recently discovered Ain't No Catchment Cave, named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

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

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

### Barton Hill

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



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

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

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

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

### **STOP 8 – SCHOHARIE CAVERNS**

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

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

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

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

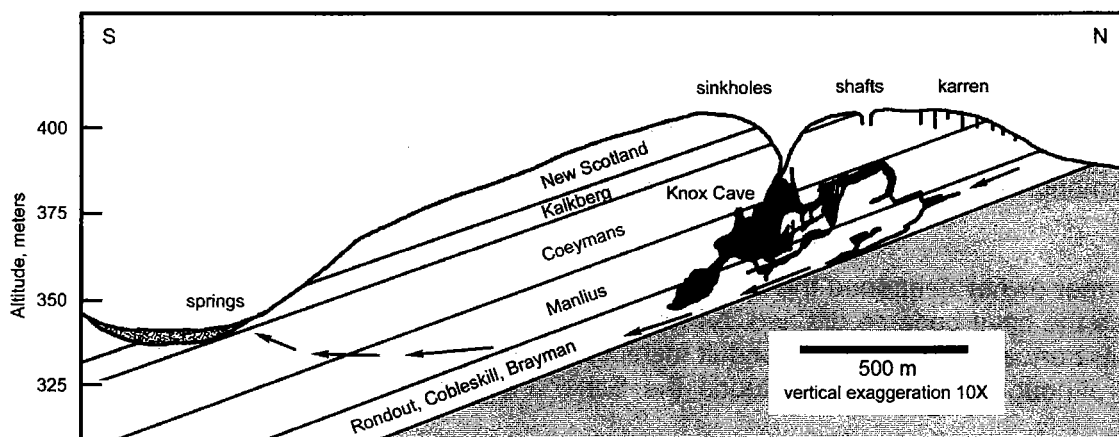
Most or all of the water that feeds Schoharie Caverns has also passed through the soil, but the high degree of aeration in the cave causes calcite to precipitate inside as speleothems. The water still continues to lose carbon dioxide where it exits the cave, maintaining slight supersaturation, but not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

Another much smaller spring lies a few hundred meters to the southeast in the Rondout or Cobleskill Formation. It has deposited a large sheet of tufa on the hillside below the cave entrance. Evidently the water that feeds the cave has passed through soil rich in carbon dioxide and has reached equilibrium with dissolved limestone at these high CO<sub>2</sub> levels. The groundwater was not aerated through cave entrances, and so carbon dioxide rapidly degasses from the water only where it comes in contact with the low carbon-dioxide levels of the outside air. The water rapidly becomes highly supersaturated and deposits some of its dissolved load.

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

### Karst Systems of the Knox Area

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



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

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

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

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

## STOP 9 – LIMESTONE RISE

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



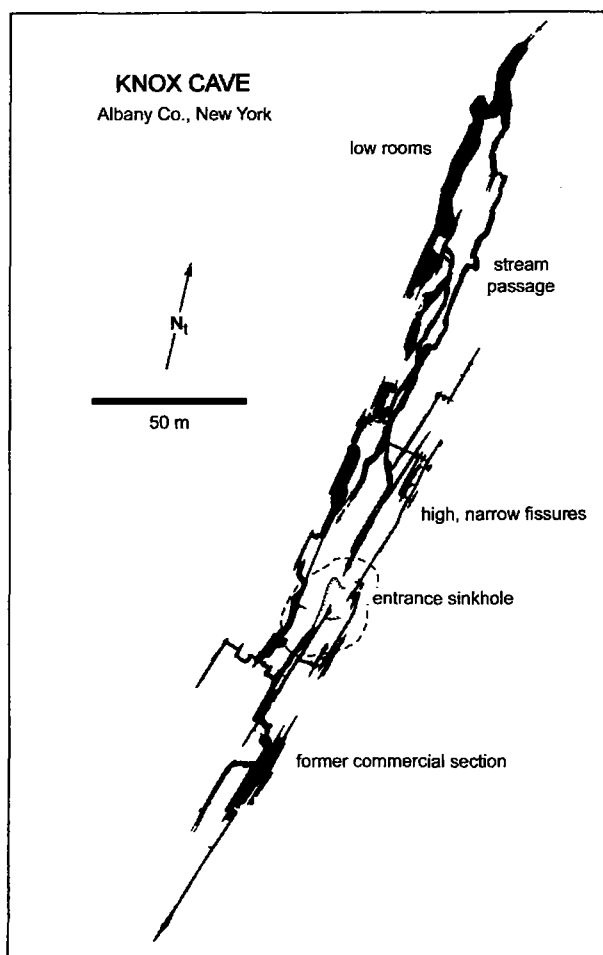
**Figure 17:** Solutionally enlarged joints in the Coeymans Limestone at Limestone Rise.

## STOP 10 – ENTRANCE SINKHOLE OF KNOX CAVE

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

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

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



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



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



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

### **STOP 11 – SPRINGS SOUTH OF KNOX CAVE**

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

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

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

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

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## ROAD LOG

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

Miles – Cumulative / (since last stop)

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

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

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

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

## TRIP A-5

# GEOLOGY AND GEOCHRONOLOGY OF THE SOUTHERN ADIRONDACKS

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Colgate Univ. Emeritus

### INTRODUCTION

The location of the Adirondacks within the larger Grenville Province is shown in Fig. 1. Topographically, the Adirondacks are divided into Highland (H, Fig 2) and Lowland (L, Fig. 2) sectors. The former is underlain principally by orthogneiss, the latter by paragneiss rich in marble, and the Carthage-Colton Mylonite Zone separates the two regions (Fig 2a,b). The region has experienced multiple metamorphic and intrusive events and large-scale ductile structures are common (McLelland, 1984; McLelland *et al.*, 1996). U-Pb zircon geochronology (Table 2, Fig. 3) indicates that the oldest igneous rocks exposed are ca 1350-1300 Ma tonalitic, arc-related plutons (Fig.2a) intrusive into older Highland paragneisses of uncertain age (McLelland *et al.*, 1996). The oldest igneous rocks in the Lowlands consist are ca. 1200 Ma granodiorites (Fig.2a) intrusive into older paragneisses of uncertain age (Wasteneys *et al.*, 1999). In both the Highlands and Lowlands metapelitic migmatites >1200-1300 Ma contain anatectic material of ~1170 Ma age. Following intrusion ca. 1207 Ma granodiorites in the Lowlands, leucogranitic and tonalitic rocks were emplaced at ca 1172 Ma (Fig.2a,b) and were accompanied by deformation and metamorphism (Wasteneys *et al.*, 1999) assigned to the latest, culminating phase (ca. 1220-1160 Ma) of the Elzevirian Orogeny of Moore and Thompson (1980).

From ca 1160-1150 Ma the entire Adirondack-Frontenac region was intruded by anorthosite-charnockite-mangerite-granite (AMCG, 1155 Ma, Fig. 2b) magmas that are associated with four anorthosite massifs (Marcy, Oregon, Snowy, Carthage) recently dated *directly* at  $1155 \pm 10$  Ma by SHRIMP II zircon techniques. Early age determinations of the anorthosite (McLelland and Chiarenzelli, 1990; Silver, 1969) were based on multigrain dating of MCG granitoids that exhibit mutually crosscutting relationships with the Marcy Anorthosite Massif (MM, Fig. 2b) and are interpreted as coeval with it. The absence of direct dating of the anorthosite was due to the sparse igneous zircon populations in rocks of this composition, a condition that posed a serious obstacle to the study of anorthosites prior to the advent of single grain TIMS and SHRIMP II methods. Current SHRIMP II direct dating of the Adirondack anorthosite massifs demonstrates that both they and their associated ferrodiorites and granitoids were emplaced at  $1155 \pm 10$  Ma and that the entire complex represents a classic AMCG suite.

The final major events of Adirondack evolution comprise: 1) the emplacement of the Hawkeye granite suite at ca 1095 Ma (Fig 2b) followed almost immediately by 2) high-grade metamorphism (Storm and Spear, Appendix to this article; Spear and Markussen, 1997; Bohlen *et al.* 1985; Valley *et al.*, 1990) resulting in vapor-absent, peak granulite facies conditions in the Highlands ( $T \sim 750^{\circ}$ - $800^{\circ}$  C,  $P \sim 6$ - $8$  kbar) and associated with widespread recumbent, isoclinal folding and the development of intense penetrative fabrics. This granulite facies metamorphism and deformation are assigned to the collisional Ottawa Orogeny of Moore and Thompson (1980), evidence of which occurs throughout the Grenville Province (cf. Rivers, 1997; McLelland *et al.*, 2001b). Toward the end of this major orogenic event much of the Adirondack region was intruded by late- to post-tectonic leucogranites (ca. 1055 Ma, Fig 2a) belonging to the Lyon Mt. Granite (LMG) and thought to be related to delamination and extensional collapse of the orogen (McLelland *et al.*, 2001b). The Elzevirian and Ottawa Orogenies, taken together, comprise the Grenville Orogenic Cycle (ca. 1350-950 Ma) of Moore and Thompson (1980).

## GENERAL GEOLOGY

The southern Adirondacks are underlain by a package of older rocks that is only sparsely represented north of the Piseco anticline. These same lithologies are present in the eastern Adirondacks east of the Northway (Rt 87). The package is characterized by a significant thickness of migmatitic metapelites (Stop 1), some very thick orthoquartzites (Stop 2), tonalitic and granodioritic plutons (Stops 3 and 4), and various members of the AMCG suite (Stops 5 and 8). Marbles are present (Stop 7) and deformation is profound (Stops 6 and 9). Taken as a whole, the southern and eastern Adirondacks appear to have been derived from one or more magmatic arcs (Fig. 4) similar to those elsewhere in the Central Metasedimentary Belt (cf., Rivers, 1997; Carr *et al.*, 2000). The growth and amalgamation of these arcs spans the time interval ca 1400-1170 Ma and is referred to as the Elzevirian

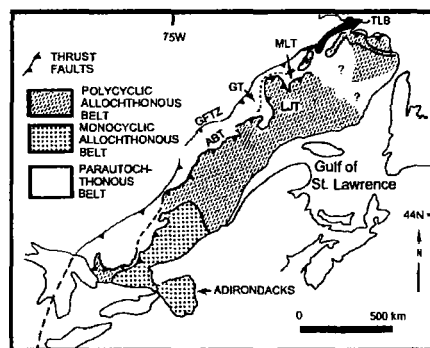


Fig. 1. Location of the Adirondack Mts. within the greater Grenville Province. Tectonic subdivisions after Rivers (1997). GFTZ - Grenville Front Tectonic Zone; ABT- Allochthon Boundary Thrust; GT-Gagnon Terrane; LJT-Lac Jeune Terrane; MLT-Melville Lake Terrane; TLB - Trans-Labrador Batholith

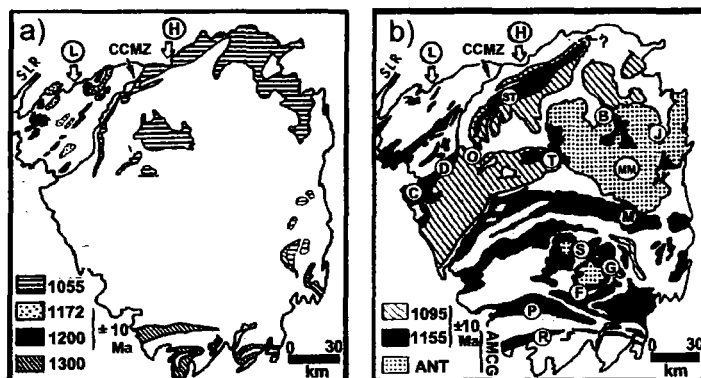


Fig. 2- Generalized geologic-chronologic maps of the Adirondacks broken into two panels for ease of viewing. The major metaigneous units are shown by patterns and their ages are given in the legend. C-Carthage, D-Diana, B-Bloomington; J-Jay; F- Oregon Domic Ferrodiortite; G-Gore Mt; O-Oswegatchie; P-Piseco; R-Rooster Hill; T-Tupper Lake; ST-Stark; MM-Marcy Massif, CCMZ- Carthage Colton Mylonite Zone; H-Highlands; L-Lowlands; SLR-St. Lawrence River.

Orogeny. The culminating Elzevirian Orogeny is thought to have occurred during the interval ca 1210-1170 Ma and resulted from the collision of the Adirondack-Green Mt block with the southeastern margin of Laurentia, which at that time was represented by a magmatic arc of northwest polarity developed on the present day Adirondack Lowlands. This culminating collision resulted in deformation and metamorphism that can be recognized in the southern and central Adirondack Highlands. The tonalites and granodiorites manifest the arc(s) and the migmatitic metapelites represent its apron of flysch. The thick orthoquartzites and thin marbles may represent shelf sequences developed on the passive margin of the Adirondack Highlands-Green Mt block prior to the culminating collision. These events are summarized in Fig. 4.

Following the culminating Elzevirian collision, the overthickened orogen began to delaminate and rebound as buoyancy forces dominated those of contraction. As the orogen rebounded, it underwent structural collapse and exhumation, and regional extensional basins formed and accumulated sediments, e.g., the Flinton Basin and Flinton Group (Fig. 4). At the base of the orogen hot new athenosphere moved in to replace delaminated lithosphere  $\pm$  lower crust, and depressurization led to the production of gabbroic melts that ponded at the crust-mantle interface in response to density inversion. Due to the dominance of buoyancy forces, the crust-mantle environment was relatively stable, and the gabbroic melts were able to undergo quiescent crystallization. Under these conditions olivine and pyroxene sank to the floor of the chambers while plagioclase crystals floated and accumulated into crystal-rich mushes along the chamber roof. Simultaneously, the latent heat of crystallization from this process provided enough thermal energy to cause significant partial melting of the overlying continental crust. The resultant melts were largely anhydrous and ranged from syenitic, to monzonitic, to granitic and, with the crystallization of orthopyroxene, yielded mangerites and charnockites. Ultimately the crust grew weak enough, and the plagioclase mushes (ie gabbroic anorthosite) of sufficiently low density that they began to ascend and were emplaced at upper crustal levels (<10 km, Valley, 1985; Spear and Markussen, 1997) where they crystallized into AMCG complexes. Continued low-pressure fractionation of the anorthositic magmas led to further growth of plagioclase and the consequent evolution of increasingly mafic interstitial, residual liquid. Many of these were filter-pressed into fractures to form dikes and sheets of ferrodiorite; others pooled into plutonic ferrodioritic masses. We speculate that ultimately the evolving liquids became so mafic that they underwent immiscibility to produce magnetite-ilmenite concentrations such as those at Tahawus (Fig. 2) and comb-textured clinopyroxene-plagioclase dikes similar to those seen at Jay on trip C-1. Representative whole rock compositions of the AMCG suite are given in Table 1. U-Pb dating of all of these rock-types documents that the Adirondack AMCG suite was emplaced at ca 1155  $\pm$  10 Ma (Table 2). Recent attempts to assign an age of ~1040-1050 Ma to the emplacement of the Marcy Massif are inconsistent with this hard evidence and are quite simply wrong.

Table 1. Representative whole Rock Analyses of AMCG suite Rocks

|                                | 1*    | 2*     | 3*    | 4*    | 5*    | 6*    | 7*    | 8*     |
|--------------------------------|-------|--------|-------|-------|-------|-------|-------|--------|
| SiO <sub>2</sub>               | 42.8  | 55.88  | 56.89 | 53.65 | 62.12 | 51.63 | 54.54 | 53.54  |
| TiO <sub>2</sub>               | 6.04  | 1.6    | 0.47  | 0.52  | 0.87  | 3.1   | 0.67  | 0.72   |
| Al <sub>2</sub> O <sub>3</sub> | 10.53 | 23.18  | 23.82 | 24.90 | 16.48 | 14.23 | 25.61 | 22.50  |
| Fe <sub>2</sub> O <sub>3</sub> | 21.6  | 2.4    | 1.21  | 0.41  | 1.49  | 2.1   | 1.00  | 1.26   |
| FeO                            | na    | 6.57   | 1.3   | 0.70  | 3.96  | 13.5  | 1.26  | 4.14   |
| MnO                            | 0.02  | 0.09   | 0.02  | 0.02  | 0.09  | 0.16  | 0.02  | 0.07   |
| MgO                            | 5.68  | 2.08   | 0.65  | 1.45  | 1.06  | 2.63  | 1.03  | 2.21   |
| CaO                            | 8.77  | 4.87   | 8.19  | 12.21 | 3.27  | 6.5   | 9.92  | 10.12  |
| Na <sub>2</sub> O              | 2.17  | 4.26   | 5.38  | 3.92  | 4.81  | 2.67  | 4.53  | 3.70   |
| K <sub>2</sub> O               | 0.84  | 2.76   | 1.13  | 1.20  | 5.13  | 2.41  | 1.01  | 1.19   |
| P <sub>2</sub> O <sub>5</sub>  | 0.66  | 0.48   | 0.09  | 0.09  | 0.30  | 0.57  | 0.09  | 0.13   |
| Total                          | 99.34 | 101.47 | 99.57 | 99.9  | 99.9  | 99.5  | 100.1 | 100.00 |

1\*- Oregon Dome Ferrodiorite; 2\*- Gabbroic Anorthosite, Green Mt; 3\*- Anorthosite, Green Mt; 4\*- Anorthosite, Owl's Head; 5\*- Mangerite, Tupper Lake; 6\*- Keene Gneiss, Hull's Falls; 7\*- Marcy Facies Anorthosite; 8\* - Whiteface Facies Anorthosite.

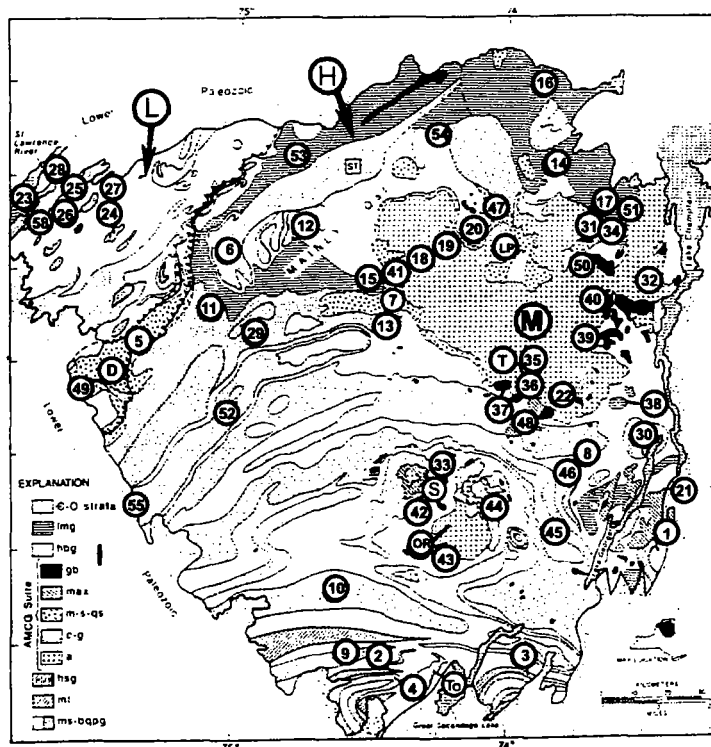


Fig 3. Generalized geologic map of the Adirondack Mtns. Showing locations of samples dated by U-Pb techniques and keyed to Table 2. H-Highlands, L-lowlands, CCMZ-Carthage Colton Mylonite Zone, D-Diana, LP-Lake Placid, OR-Oregon Dome, M-Marcy Massif, T-Tahawus, To-Tomantown pluton, ST-Stark Anticline, lmg-Lyon Mountain Granite, hbg-homblende granite, ga-gabbro, max-mangerite with andesine xenocrysts, m-s-qs-mangerite, syenite, c-g-chamokite, granite, a-anorthosite, hsg-Hyde School Gneiss, mt-metatonalite, ms-bppq-metasediments. Biotite-quartz-plagioclase.



Table 2. Summary of U-Pb Zircon Geochronology For Adirondack Metigneous Rocks

| Map Number                     | Sample Number | Location               | Multigrain TIMS |       | Singlegrain TIMS |     | SHRIMP II Analysis |     |      |
|--------------------------------|---------------|------------------------|-----------------|-------|------------------|-----|--------------------|-----|------|
|                                |               |                        | AGE (Ma)        | ERR   | AGE (Ma)         | ERR | AGE (Ma)           | ERR | TDM  |
| <b>HIGHLANDS</b>               |               |                        |                 |       |                  |     |                    |     |      |
| Tonalite and Granodiorite      |               |                        |                 |       |                  |     |                    |     |      |
| 1                              | AM87-12       | South Bay              | 1329            | 37    |                  |     |                    |     | 1403 |
| 2                              | AM86-12       | Canada Lake            | 1302            | 6     |                  |     |                    |     | 1366 |
| 3                              | LDT           | Lake Desolation        | >1336           |       |                  |     |                    |     | 1380 |
| 4                              | AM87-13       | Canada Lake            | 1253            | 41    |                  |     |                    |     |      |
| Mangerite and Charnockite      |               |                        |                 |       |                  |     |                    |     |      |
| 5                              | AM86-2        | Diana Complex          | 1155            | 4     |                  |     | 1154               | 17  | 1430 |
| 6                              | AM86-15       | Stark Complex          | 1147            | 10    |                  |     |                    |     | 1495 |
| 7                              | AM86-6        | Tupper Lake            | 1134            | 4     |                  |     | 1169               | 11  | 1345 |
| 8                              | 9-23-85-7     | Schroon Lake           | 1125            | 10    |                  |     | ca 1155            |     |      |
| 9                              | AM86-17       | Rooster Hill           | 1156            | 8     |                  |     |                    |     | 1436 |
| 10                             | AM86-9        | Piseco Dome            | 1150            | 5     |                  |     |                    |     | 1346 |
| 11                             | AC85-2        | Oswegatchie            | 1146            | 5     |                  |     |                    |     |      |
| 30                             | Silver, 69    | Ticonderoga            | 1113            | 16    |                  |     | ca 1155            |     |      |
| 42                             | AM86-8        | Snowy Mt               | >1095           |       |                  |     | 1177               | 22  |      |
| 44                             | AM87-3        | Gore Mt                | >1088           |       |                  |     | 1155               | 6   |      |
| 47                             | AC85-10       | Bloomingdale           | 1133            | 51    |                  |     | 1160               | 14  |      |
| 48                             | AM87-10       | Minerva                | >1082           |       |                  |     | 1159               | 12  |      |
| 49                             | AM86-1        | Croghan                | 1155            | 13    |                  |     |                    |     |      |
| 41                             | Granitedike   | Wabeek Quarry          |                 |       |                  |     | ca 1155            |     |      |
| 50                             | AC85-11       | Yard Hill              | 1143            | 33    |                  |     |                    |     |      |
| Anorthosite and Olivine Gabbro |               |                        |                 |       |                  |     |                    |     |      |
| 18                             | AC85-8        | Rt 3, Saranac Lk, ANT  | >1113, 1054     | 22    |                  |     | 1149               | 35  |      |
| 19                             | AC857         | Rt 3, Saranac Lk, ANT  | >1087, 1052     | 20    |                  |     | 1161               | 12  |      |
| 20                             | AC85-9        | Forest Home Rd, ILM    | 996             | 6     |                  |     |                    |     |      |
| 21                             | AM87-11       | Dresden Station Gab.   | 1147            | 7     |                  |     |                    |     | 1331 |
| 22                             | CGAB          | North Hudson Gabbro    | >1109, 1057     | conc. |                  |     | 1150               | 14  |      |
| 31                             | BMH01-4       | Jay, ANT Pegmatite     |                 |       |                  |     | 1160               | 15  |      |
| 34                             | BMH-01-3      | Jay, Cpx-Pgt Dike      |                 |       |                  |     | 1140               | 18  |      |
| 35                             | BMH01-1       | Tahawus ANT            |                 |       |                  |     | ca 1155            |     |      |
| 36                             | BMH01-2       | Blue Ridge ANT         |                 |       |                  |     | 1153               | 11  |      |
| 37                             | BMH01-1       | Blue Ridge Gabbro      |                 |       |                  |     | ca 1155            |     |      |
| 39                             | BMH01-19      | Exit 29 NWY, ANT       |                 |       |                  |     | ca 1155            |     |      |
| 40a                            |               | Woolen Mill Gabbro     |                 |       |                  |     | 1154               | 9   |      |
| 40b                            |               | Woolen Mill ANT2       |                 |       |                  |     | 1151               | 6   |      |
| 43                             | AM87-8        | Oregon Dome Fer'drt    |                 |       |                  |     | 1155               | 6   |      |
| Hawkeye Granite Suite          |               |                        |                 |       |                  |     |                    |     |      |
| 12                             | AM86-3        | Cary Falls             | 1100            | 12    |                  |     |                    |     |      |
| 13                             | AM86-6        | Tupper Lake            | 1098            | 4     |                  |     |                    |     | 1314 |
| 14                             | AM86-13       | Hawkeye                | 1093            | 11    |                  |     |                    |     |      |
| 45                             | Moon Mt.      | Moon Mt                | 1103            | 15    |                  |     |                    |     |      |
| 52                             | NOFO-1        | Stillwater Reservoir   | 1095            | 5     |                  |     |                    |     |      |
| 53                             | AM87-6        | St. Law/Fran. Co. Line | 1090            | 6     |                  |     |                    |     |      |
| 54                             | AM87-7        | Santa Clara            | 1080            | 4     |                  |     |                    |     |      |
| Lyon Mt Granite                |               |                        |                 |       |                  |     |                    |     |      |
| 15                             | AM86-4        | Piercefield            | 1075            | 17    |                  |     | 1058               | 18  | 1576 |
| 16                             | AM86-10       | Dannemora              | 1073            | 6     |                  |     | 1052               | 11  |      |
| 17                             | AM86-14       | Ausable Forks, Qtz-Ab  | 1057            | 10    |                  |     | 1041               | 16  | 1350 |
| 29                             | CLFG          | Wanakona               | 1113            | 10    | 1069             | 10  | 1047               | 10  |      |
| 38                             | 9-23-85-6     | Grasshopper Hill       | >1085           |       | 1049             | 3   |                    |     |      |
| 51                             | AM86-11       | Ausable Fks., Fay GRT  | 1089            | 26    | 1047             | 2   |                    |     |      |
| 55                             | PL-3          | Port Leyden            |                 |       | 1035             | 4   |                    |     |      |
| <b>LOWLANDS</b>                |               |                        |                 |       |                  |     |                    |     |      |
| Hyde School Gneiss             |               |                        |                 |       |                  |     |                    |     |      |
| 23                             | AM86-16       | Wellesley Island       | 1416            |       | 1172             | 5   |                    |     | 1440 |
| 24                             | AC85-4        | Gouverneur             | 1284            |       |                  |     |                    |     | 1525 |
| 25                             | AC87-4        | Fish Creek             | 1236            |       | 1172             | 5   |                    |     | 1210 |
| 26                             | AC85-5        | Hyde School            | 1230            |       | 1172             | 5   |                    |     | 1360 |
| 27                             | AC85-1        | Reservoir Hill         | ca 1172         |       |                  |     |                    |     |      |
| Antwerp Granitoid              |               |                        |                 |       |                  |     |                    |     |      |
| 58                             | ANTG          | Antwerp-Rossie         | 1183            | 7     |                  |     | 1207               | 20  |      |

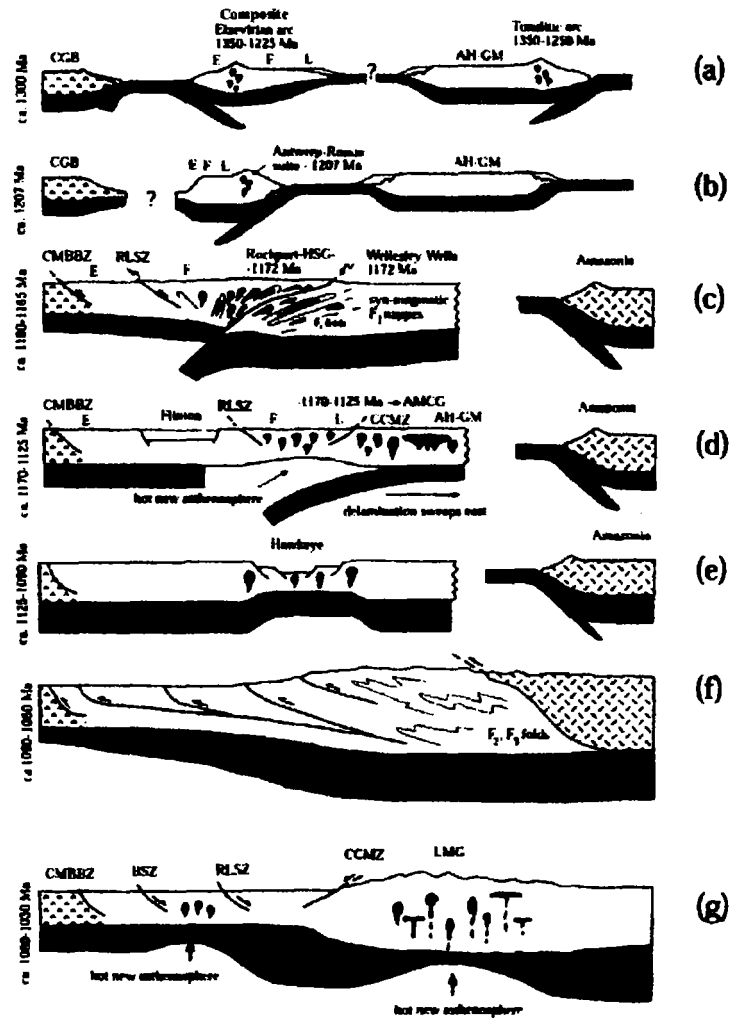


Fig. 4. Tectonic cartoon showing the evolution of the Grenville Province. See text for discussion. Modified after McLelland et al. (1996) and Wateney et al. (1999). Abbreviations as follows: AH-GM, Adirondack Highlands-Green Mountains; BSZ, Baacraft shear zone; CGB, Central Granulite Belt; CCMZ, Carthage Cotton mylonite zone; CMBBZ, Central Meta-sedimentary Belt Boundary Zone; RLSZ, Robertson Lake Shear Zone; E, Elzevir terrane; F, Frontenac terrane; HSG, Hyde School gneiss; L, Adirondack Lowlands; LMG, Lyon Mountain granite.

As pointed out by Buddington (1939), there exist several facies and types of anorthosite and related rocks. Examples are given in Table 1. The coarse Marcy facies appears to be a cumulate, and in its purest form is found in rafts of 10-20cm quasi-euhedral grains with ~10% subophitic pyroxene. These are interpreted as rafts formed by plagioclase flotation at the base of the crust and subsequently transported upward by other, less coarse and commonly more mafic facies of anorthosite. The composition of these large plagioclases ranges from  $An_{45}$ - $An_{52}$ . Fram and Longhi (1994) showed that pressure decreases the An content of plagioclase crystallizing from gabbroic magma at the rate of ~1%An/kbar. Since the rafts are thought to have crystallized from gabbro at ~10-12 Kbar, this would explain their relatively sodic compositions. The presence of giant (10-25cm) aluminous orthopyroxene in the rafts is also consistent with this model. It is common for rafts to be disrupted by the transporting magma and numerous large, blue-gray andesine crystals in finer grained anorthositic rocks are of this origin. The various plagioclase mushes that were emplaced within the crust are thought to have been broadly similar to the Whiteface facies (Buddington, 1939) and evolved towards more pure anorthosite (generally less coarse than rafts) by low-pressure fractionation of plagioclase. During this process ferrodioritic residual magmas were produced and were commonly filter-pressed into dikes and sheets (McLelland et al, 1990). Ultimately, the residual magmas became so mafic that they split into immiscible silicate and Fe, Ti-oxide phases to yield magnetite-ilmenite deposits such as those at Tahawus.

Emplacement of the AMCG suite was followed by ~50 Ma of relative quiescence terminated by emplacement of the Hawkeye granitic suite at 1103-1095 Ma. This interval corresponds almost exactly with the major magmatism in the plume-related Mid-continent Rift, and we attribute Hawkeye magmatism to far-field echoes from the Mid-continent plume (Cannon, 1994). This far-field effect is interpreted to have thinned the crust and lithosphere and led to significant crustal melting to produce the mildly A-type Hawkeye suite. At the same time, the crust underwent heating that continued to be present when the culminating collision with Amazonia (?) took place thus initiating the Ottawa Orogeny at ca 1090 Ma. Note that magmatism in the Mid-continent Rift was shut off at this time by westward-directed thrust faults. Within the Adirondacks, the already heated crust was loaded by thrusting and contraction along ~NW-SE lines of tectonic vergence. This heating followed by loading led to a counterclockwise P,T,t path (see Fig. 12). From ~1090-1060 Ma contraction dominated the region and great nappe structures formed (Figs. 5,6). These are represented today by extremely large isoclinal, recumbent folds such as the Canada Lake isocline and the Little Moose Mt syncline, both of which have ~E-W axial trends and plunge gently about the horizontal (Fig. 6). It is uncertain, but possible, that these structures were initiated as thrusts and then toed-over to form a greatly thinned and attenuated lower limb. Associated with the nappes are pervasive penetrative fabrics imposed upon Hawkeye and older rocks and attesting to the extraordinarily high temperature ductile strains imposed on these rocks (McLelland, 1984).

As discussed by Spear and Markussen, garnet coronas in mafic rocks appear to have formed during late-Ottawan isobaric cooling from ~800-600°C. Associated with these coronitic rocks are small, equant zircons interpreted as metamorphic in origin and yielding ages of ~1050 Ma. As discussed in the text for Stop 8, these are thought to date the corona-forming reaction. This is consistent with the Sm-Nd age of ca 1050 Ma for the large Gore Mt garnets (Mezger et al, 1992). Following isobaric cooling the Ottawa orogen is thought to have undergone delamination and rebound. This was accompanied by emplacement of the distinctive Lyon Mt Granite suite that hosts the great Kiruna-type low-Ti magnetite deposits of the Adirondacks and is exposed across wide tracts of the Highlands. Zircon dating of the Lyon Mt Granite by both single grain TIMS and SHRIMP II methods (McLelland et al, 2002) documents that the thick, and oscillatory zoned, mantles of its zircons grew from melts at  $1050 \pm 10$  Ma. The cores of these zircons are of AMCG and Elzevirian age and whole rock Nd-model ages are consistent with the production of Lyon Mt Granite from melting of these earlier lithologies. An especially interesting member of the Lyon Mt Granite is a quartz-albite ( $Ab_{98}$ ) that is associated with the iron-oxide deposits and is interpreted to be the result of sodic hydrothermal alteration (McLelland et al., 2002).

Further relevant details accompany the descriptions presented in at individual stops.

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McLelland

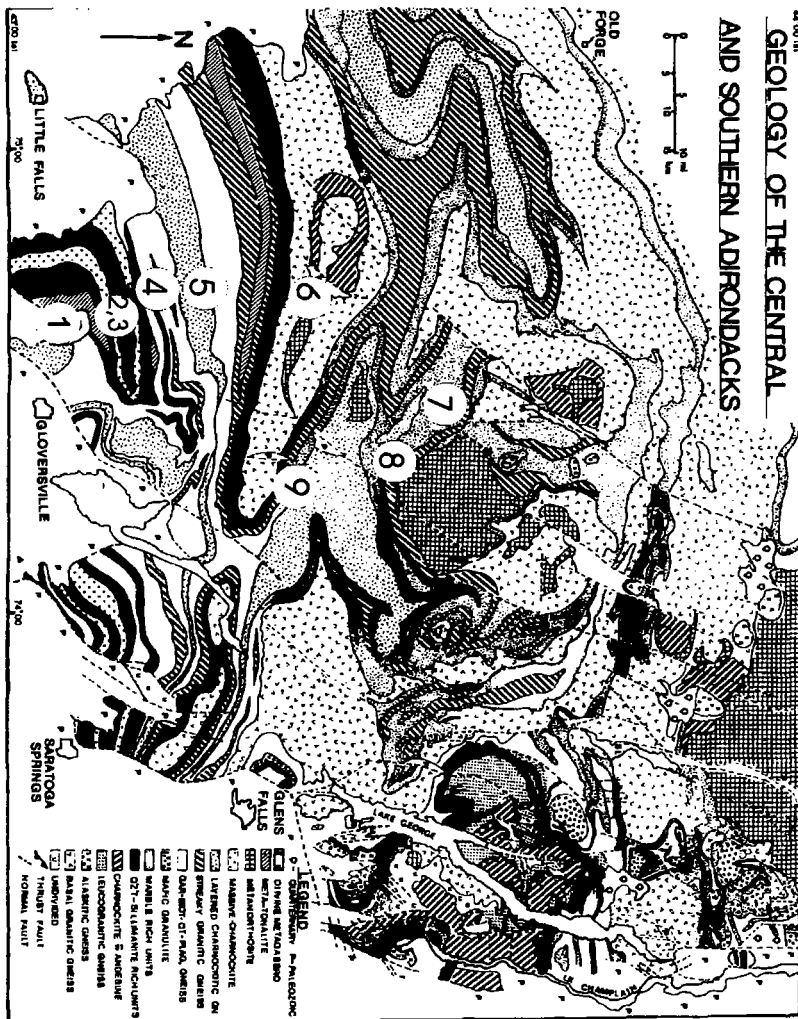
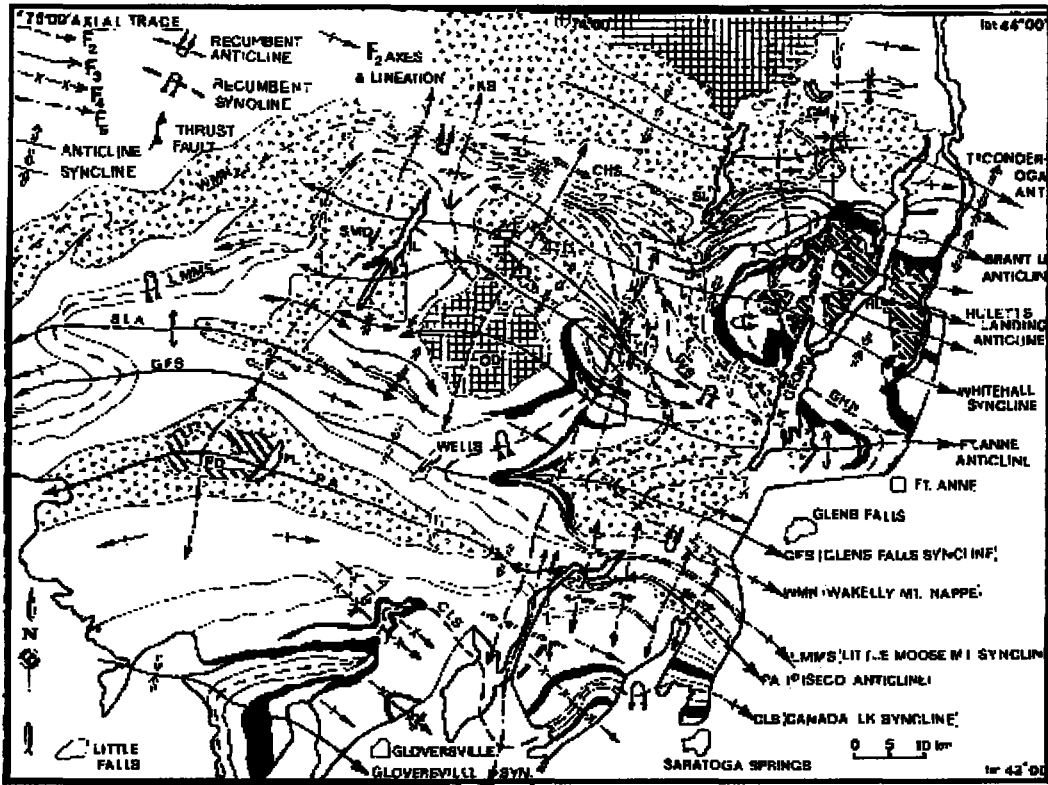


Fig. 5 Generalized geologic map of the southern and central Adirondacks showing field trip stops 1-9.



McLelland

Fig. 6 Fold structures of the southern and central Adirondacks

9

**ROAD LOG  
MILEAGE**

0.0 Caroga Lake Post Office in Caroga Lake, NY.

2.8 Roadcuts of migmatitic metapelite. Park on right (west) shoulder of Rt 29A.

**STOP 1. PECK LAKE MIGMATITIC METAPELITE. (30 MINUTES).** This exposure along Rt 29A just north of Peck Lake is the type locality of the Peck Lake migmatitic metapelites that consist of restitic sillimanite-garnet-biotite-quartz-oligoclase melanosome and quartz and two-feldspar leucosome of approximately minimum melt composition (McLelland and Husain 1986). Small red garnets are common in the leucosome and, in most cases, grow across foliation. The leucosomes occur as irregular, elongate bodies generally parallel to foliation but quite commonly exhibiting crosscutting relationships with respect to the restite and to one another. Based upon these compositional and crosscutting relations, the leucosomes are interpreted as anatectites and a reasonable metapelitic source rock can be prescribed by reintegrating their composition with that of the restite (Table 3) to yield a greywacke-slate precursor. The restriction of these partial melts to within the migmatite is thought to be the result of near-solidus, hydrous melting that would cause ascending melts to intersect the solidus as they began to rise. Close inspection of the leucosomes reveals that, in their most pristine configuration, they consist of coarse granite and pegmatite. In low strain zones elsewhere in the Adirondacks it is manifestly clear that the leucosomes originally formed an anastomosing arrays of veins, dikes, sheets, and pods. Subsequent high strain resulted in rotation into pseudoparallelism and commonly produced disruption that caused separate grains of white feldspar some of which show elongate tails – ie, the rock was on its way to becoming a mylonitic "straight gneiss". At the last stop (Stop 9) of this trip we shall see the end result of this process in a series of platy, stretched, and highly grain size reduced and mylonitic equivalents of the rocks seen here. The anatectic origin of these units is further suggested by the uncommon, but not rare, occurrence of plagioclase- and/or garnet-rimmed hercynitic spinel in the restite. Recently, Bickford and McLelland have run a U/Pb zircon pilot study on the age of the anatectites and have found that they contain 1220-1250 Ma cores, 1020-1050 Ma metamorphic rims, and relatively thick, nicely zoned mantles that fall into the interval 1190-1170 Ma. Some of the rims show zoning, but this is minor compared to the mantling zircon. These results demonstrate that majority of the anatectites were produced by partial melting during the culminating Elzevirian Orogeny dated at ca 1210-1170 Ma (Wasteneys and McLelland, 1999) and are not of Ottawan origin. Currently, we are continuing this research by utilizing both zircon and monazite geochronology.

**TABLE 3. COMPOSITIONS OF AVERAGE LEUCOSOME ,HOST,AND SELECTED CLASTICS**

|                                | Average<br>Leucosome<br>(N = 31) | Average<br>Host<br>(N = 12) | Average Host<br>+ 15% Average<br>Leucosome | Average<br>Greywacke<br>(N = 23) | Average<br>PC Slate<br>(N = 33) | Average<br>Slate<br>(N = 36) |
|--------------------------------|----------------------------------|-----------------------------|--|----------------------------------|---------------------------------|------------------------------|
| SiO <sub>2</sub>               | 74.39                            | 61.75                       | 63.12                                      | 64.70                            | 56.30                           | 60.64                        |
| Al <sub>2</sub> O <sub>3</sub> | 13.85                            | 17.83                       | 17.18                                      | 14.80                            | 17.24                           | 17.32                        |
| TiO <sub>2</sub>               | .05                              | 1.32                        | 1.15                                       | .50                              | .77                             | .73                          |
| Fe <sub>2</sub> O <sub>3</sub> | .89                              | 8.70                        | 7.55                                       | 4.10                             | 7.22                            | 4.81                         |
| MgO                            | .27                              | 2.07                        | 1.83                                       | 2.20                             | 2.54                            | 2.60                         |
| CaO                            | 1.25                             | 2.60                        | 2.40                                       | 3.10                             | 1.00                            | 1.20                         |
| Na <sub>2</sub> O              | 2.77                             | 2.44                        | 2.76                                       | 3.10                             | 1.23                            | 1.20                         |
| K <sub>2</sub> O               | 5.83                             | 3.20                        | 3.44                                       | 1.90                             | 3.79                            | 3.69                         |
| MnO                            | .02                              | .07                         | .06  | .10                              | .10                             |                              |
| P <sub>2</sub> O <sub>5</sub>  | .08                              | .15                         | .12  | .20                              | .14                             |                              |
| LOI                            | .30                              | .42                         | .40  | 2.40                             | 3.70                            | 4.10                         |
| TOTAL                          | 99.70                            | 100.55                      | 100.00                                     | 101.00                           | 98.70                           | 98.00                        |

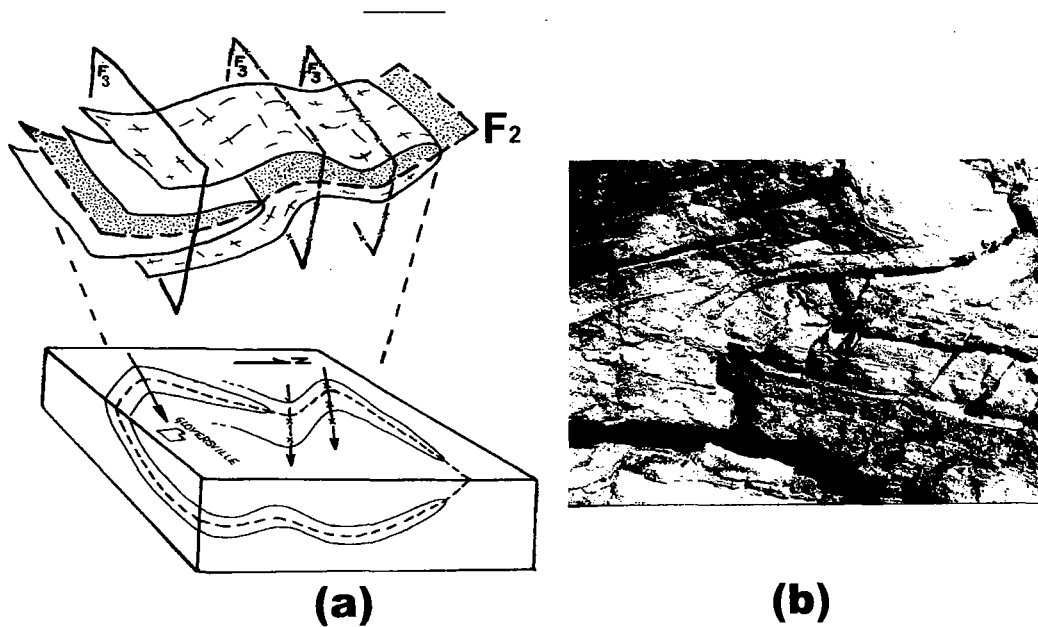


Fig.7 (a) Three dimensional cartoon showing geometry of interference of isoclinal and upright folds in the Canada Lake Isocline ( $F_2$ ) and the resultant outcrop pattern. (b)  $F_2$  minor fold recently blasted from roadcuts at stop 2.

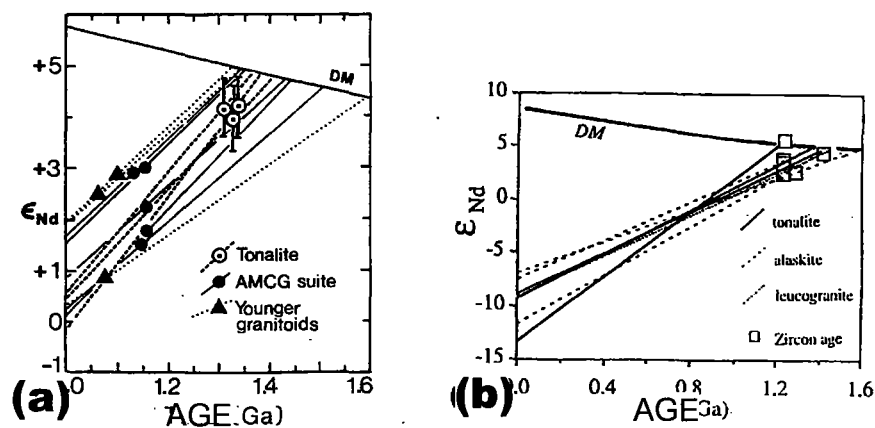


Fig . 8 Epsilon Nd plots for the Adirondack Highlands (a) and Lowlands (b)

This outcrop provides an excellent mesoscale example of Adirondack structure (Fig. 7). The overall strike of foliation is N60-70W and dips vary from north to south and pass through the vertical demonstrating that the roadcut defines a large minor isoclinal, recumbent fold. Smaller, meter-scale isoclinal folds are also well exposed and are aligned parallel to the roadcut-scale fold. All of these are similar in style, and parallel to, the very large, regional Canada Lake isocline or nappe. It is clear that the Canada Lake nappe stage of deformation rotates an earlier foliation and has a moderately dipping axial planar foliation that intersects the earlier foliation at a high angle. The time interval between these foliations remains unspecified, but it is clearly post-anatexis and probably all Ottawa (Ca 1090-1030 Ma). It appears that the Ottawa has successfully obliterated most Elzevirian fabrics. Note that some of the smaller minor folds contain apparently terminated compositional layers that could very well represent Elzevirian isoclinal noses. If so we are looking at isoclinally refolded isoclines.

Migmatitic metapelites of this sort occur throughout the southern and western Adirondacks as well as in the Adirondack Lowlands. We interpret them as flysch sequences of shale and greywacke that were being shed from the Elzevirian magmatic arcs that dominated the region from ca 1400-1300 Ma. Tonalites and granodiorites of the arcs locally crosscut the metapelites and provide a minimum age for them. In a broad sense, they are thought to be coeval. The metapelites of the Adirondack Lowlands are compositionally similar, but are thought to represent a different, younger, and uncorrelative arc environment.

- 3.5 Turn around at Peck Lake and head back north on Rt 29A.
- 5.9 Junction NY Rt. 29A and NY Rt. 10.
- 7.3 Nick Stoner Inn on west side and Nick Stoner Golf Course on east side of Rt 29A-10.
- 7.8 Town of Caroga sand and gravel depository on east (right) side of Rt 29A-10. Pull in and park.

**STOP 2. IRVING POND QUARTZITE. (20 Minutes).** The Irving Pond quartzite unit cores the Canada Lake isocline and is folded back on itself. At map scale it is exposed across strike for over 3000m but its "true?" thickness is on the order of 1000m. Notwithstanding, it represents an enormous volume of orthoquartzites with minor pelitic intercalations. Its minimum age is unknown but is being investigated by zircon geochronology. Similar quartzites along the St Lawrence River contain zircons as young as 1300 Ma, and the same may prove to be true for the Irving Pond. It is suggested that these quartzites may have been deposited along the present eastern margin of the Adirondack-Green Mt block between ca 1250 and 1200 Ma (see Fig. 4). During that interval, this margin was passive as the block moved westward towards the Elzevirian subduction zone that dipped westward beneath Laurentia and its leading margin, ie, the Adirondack Lowlands (Fig. 4b,c). Upon collision, (ca 1200-1170 Ma), the accumulated sandstones were deformed and metamorphosed. Although this scheme is speculative, it is consistent with the little that is known about Elzevirian events.

Within the clearing there are three small, but informative outcrops. The first consists of 30-40cm-scale layers of pure quartzite together with 2-3cm-scale metapelitic layers of. These dip gently to the southeast. The bulk of the Irving Pond quartzite consists of layer upon layer of the pure quartzite with little, if any, intervening metapelite. The second exposure is located a few tens of feet to the west where the rocks form a low ledge running uphill. The southern termination of the ledge that faces the clearing shows steeply dipping layers that are discordant with underlying layers. The discordance is due to some sort of ductile shearing in the outcrop but does not affect the following interpretation. Looking back to the first outcrop, it is clear that the gently dipping layers seen there can be projected up above the ground surface so that they must have been situated overhead at this locality; however, they must also have suddenly dipped steeply and rotated through the vertical in order to have their present configuration, ie, they form recumbent isoclinal folds. The trend of this recumbent, isoclinal fold axis is ~N70W and the plunge is ~15 deg. southwest, ie, it is a minor fold of the Canada Lake isocline family and of Ottawa age. At the base of the southernmost outcropping, and below a small ledge, there is preserved a foot-scale isoclinal nose. Inspection of the geometry makes it clear that the outcrop preserves an isoclinally refolded isocline-perhaps of Elzevirian age.



Farther down the clearing towards the highway a clean outcrop consists of pure, glassy quartzite dip slopes exposed together with dark layer-like bodies of fine-grained pyroxene-plagioclase granulite. Elsewhere, the chemistry of the granulites approximates that of diabase, is unlike calcsilicate, and locally exhibits microscopic diabasic texture (McLelland and Husain, 1986). The igneous nature of the granulites is further manifested by the xenoliths of quartzite present in them. Elsewhere the metadiabases are isoclinally folded and the limbs of these folds clearly truncate 10-15 cm-scale intrafolial isoclinal folds that are rotated by the isoclinal. We interpret this older set of isoclinal minor folds to be Elzevirian in age. The metadiabases have not been dated, but they are similar to some mafic rocks of the ~1150 Ma AMCG suite and are tentatively assigned an AMCG age.

8.0 Large roadcuts of charnockite on both sides of Rt. 29A-10. Park on right hand (east) side just past guardrails at crest of hill.

**STOP 3. CANADA LAKE CHARNOCKITE. (20 MINUTES).** Large roadcuts expose the type outcrops of the Canada Lake charnockite (Figs. 8-11). This highly deformed orthogneiss has a granodioritic composition, and consists of 20-30% quartz, 40-50% mesoperthite, 20-30% oligoclase, and 5-10% mafics including small, sporadic grains of orthopyroxene. The exposures exhibit the drab olive color typical of charnockites around the world. In the woods these rocks tend to weather pink and exhibit a maple sugar brown weathering rind. The unit is ~500m thick and consists throughout of relatively homogeneous granitoid with pegmatites and minor amphibolitic layers. A multigrain U-Pb zircon age of  $1251 \pm 33$  Ma (McLelland and Chiarenzelli, 1990) indicating that this unit belongs with other calcalkaline rocks of broadly Elzevirian age. Mapping along its contact for a total distance of ~300 km, has not revealed any crosscutting features, but xenoliths of country rock have been recognized and substantiate an intrusive origin. The apparent conformity is attributed primarily to extreme and ductile tectonism. In addition, an original conformable, sheet-like form is quite possible. Rocks of similar age and composition are found in the Green Mts. of Vermont (Ratcliffe and Aleinikoff, 19xx).

9.2 Canada Lake Store on the left (south side) of Rt. 29A-10. Turn in to parking lot and park diagonally.

**STOP 4. ROYAL MT. TONALITE (30 MINUTES).** Steep roadcuts exposed across from the Canada Lake Store expose typical tonalitic rocks that are relatively common within the southern and eastern Adirondacks, the Green Mts of Vermont, and the Elzevir terrain of the Central Metasedimentary Belt of the Canadian Grenville. In all of these occurrences, the tonalites manifest the presence of magmatic arcs that existed along the southeastern margin of Laurentia during the interval ca 1400-1200 Ma diagnostic of the Elzevirian. Within the Adirondacks, multi- and single-grain TIMS U-Pb zircon geochronology indicate emplacement of the tonalitic magmas at ca 1350-1300 Ma. The present outcrop has been dated by both TIMS methods and the single grain age is constrained at  $1307 \pm 2$  Ma (Aleinikoff, pers comm., 1991). Well-documented examples of these arc terrains extend to the Llano uplift and Van Horn areas of Texas (Mosher, 1999; Patchett and Ruiz, 1990; Roback, 1996) and attest to a global-scale system that may have been an ancient analogue of the present day East Indies arc. During the Elzevirian, various arcs must have collided and amalgamated, and continental arcs may have come into existence as well. These details remain to be unraveled, but in the meantime we note that the Elzevirian came to a close at ~1210-1170 Ma with the collision of the Adirondack Highlands-Green Mt block with the Andean-style arc then existent along the southeastern edge of Laurentia (Fig 4c). McLelland et al (1991), refer to this collisional event as the "culminating Elzevirian Orogeny" that closed out the Elzevirian interval of arc magmatism and amalgamation in the area.

The whole rock chemistry of the Adirondack calcalkaline rocks are shown in Figs (9,10,11) where their calcalkaline affinities are clearly visible. In addition,  $\epsilon_{Nd}$  characteristics are presented in Fig. 8 and demonstrate that the tonalites represent juvenile additions to the crust from the mantle, and partial melting of the older rocks can produce younger Adirondack granitoids. Neither the Sm-Nd

data, nor any other isotopic data, give any hint of pre-1350 Ma crust in the Adirondacks and suggest that the original arc must have been of ensimatic origin.

Numerous disrupted amphibolitic sheets are present within this outcrop and within Adirondack tonalites in general. The origin of these is enigmatic but they do not appear to have been derived from local country rocks. Their elongate, sheet-like character suggests that they may represent coeval mafic dikes of the sort commonly observed in tonalites. It is also possible that they represent enclaves incorporated from an amphibolitic source region. This interesting problem needs some isotopic investigations to be applied to it.

In a few places the tonalite crosscuts the migmatitic metapelites seen at Stop 1. This fixes a minimum age for the latter at ca 1300 Ma. As indicated in the discussion at Stop 1, the tonalites and metapelitic rocks may represent an original, essentially coeval, magmatic arc-flysch system.

- 11.0 Pine Lake. Junction of Rts 29A and 10. Turn right (north) on Rt 10 towards Speculator.
- 16.7 North end of East Stoner Lake. Pull off into parking area just north of Town of Arietta sign.

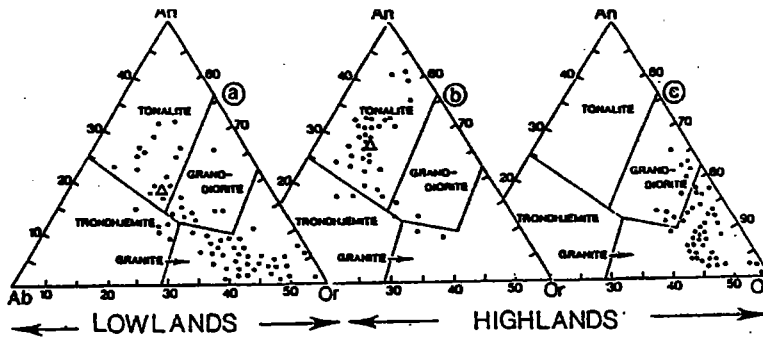


Fig. 9. Plots of normative Ab-An-Or for (a) Hyde School Gneiss, (b) Highlands tonalites and (c) Tomantown pluton. Open symbols give average value for tonalites. Fields after Barker (1979).

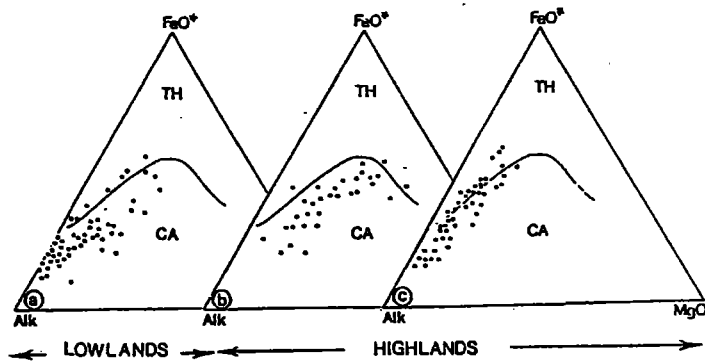


Fig. 10. AFM plots for (a) Hyde School Gneiss, (b) Highland tonalites, and (c) Tomantown pluton.

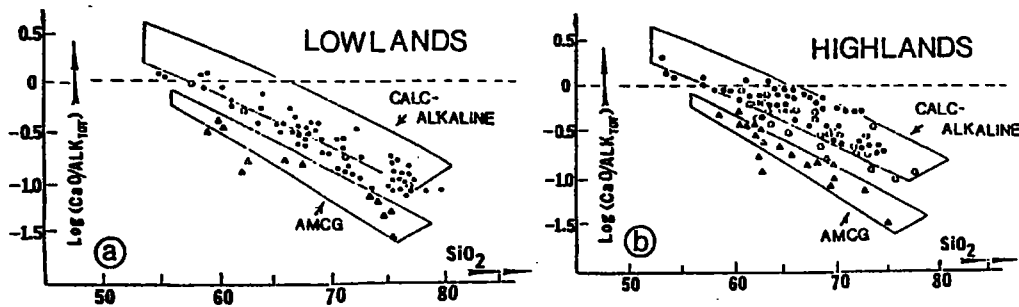


Fig. 11. Calcalkali ratio versus  $SiO_2$  for (a) the Adirondack Lowlands and (b) the Adirondack Highlands. In (a) open circles are average values for 1172 Ma Hyde School Gneiss, closed circles for typical Hyde School, and open triangles for ca 1155 Ma AMCG rocks. In (b) open circles are for the ca 1250 Ma Tomantown pluton, closed circles for the older (ca 1300 Ma) calcalkaline rocks, and open triangles for AMCG rocks. (After Brown, 1982)

**STOP 5. ROOSTER HILL MEGACRYSTIC CHARNOCKITE. (20 MINUTES).** This deformed charnockite is characterized by the presence of 20-40% megacrysts (2-4cm) of alkali feldspar set in a groundmass of quartz, oligoclase, biotite, hornblende, garnet, and sporadic orthopyroxene. In general, these have undergone dynamic recrystallization during high temperature shear strain and have developed asymmetrical tails, although flattening reduces the degree of asymmetry in most

cases. In cases where tail asymmetry permits, a southeast side up and to the northwest (N70W, 10-15SE) is clearly the dominant displacement sense. As the degree of shear strain intensifies, both feldspars and quartz become elongated in the direction of tectonic transport and ribbon or pencil gneisses result. The orientation of these fabric-forming elements is parallel to the regional isoclinal fold axes. In Fig. 8 the  $\epsilon_{Nd}$  growth curve for Rooster Hill charnockite lies on one of the AMCG suite growth lines that pass through the tonalite region indicating that this suite can be derived by partial melting of the ca 1300 Ma arc rocks.

- 19.2 Low roadcut in migmatitic metapelites.
- 20.6 Avery's Hotel on left (west) side of Rt 10.
- 21.7 Roadcut through granite, gabbro, and tonalite.
- 23.2 Pink AMCG granitic rocks, gabbro, and small exposure of fine-grained anorthosite. A large inclusion of calcisilicate in pink granite is exposed and interpreted as a xenolith.
- 23.5 Roadcut of metasedimentary quartzites and pelitic rocks together with anatectites.
- 29.2 Fault breccias in charnockite.
- 29.7 Junction of Rt 10 (ends) and Rt 8. Turn right (east) on Rt 8.
- 30.2 Long roadcut of mylonitic ribbon gneiss in pink granitic rocks of the Piseco anticline. Ribbons trend N70W and plunge 10-15 SE parallel to both recumbent  $F_1$  isoclines and upright  $F_2$  such as the Piseco anticline. A multigrain U-Pb zircon date from this outcrop gives an age of  $1155 \pm 10$  that is interpreted as the age of magmatic emplacement.
- 32.5 Pull off and park on LEFT (NW) shoulder of Rt 8.

#### **STOP 6. RIBBON GNEISS IN THE CORE OF THE PISECO ANTICLINE. (30 MINUTES).**

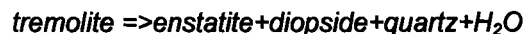
A long, low roadcut on the left (northwest) side of Rt 8 exposes outstanding examples of ribbon gneiss developed in a megacrystic facies of the Piseco core rocks. The overall composition here is similar to that of the Rooster Hill megacrystic charnockite examined at Stop 5, but here the original feldspars and quartz have been extended into ribbons on the order of 60cmx.25mmx.05mm (McLelland, 1984). Assuming interstitial quartz aggregates of ~1cm diameter, the current dimensions indicate an extension of 6000%, ie, if this strain were to be equally distributed throughout a 1km thick block of crust that block would be 60km long.25 km wide, and .005 km thick. Of course, such high ductile strain cannot be integrated through the entire crust, but the numbers provide some flavor for the strain involved.

The ribbon gneisses exposed here are folded into a minor anticline that rotates the foliation of these L>>S tectonites. The axis of this minor fold trends N70W and plunges 10-15SE. This orientation is parallel to the axis of the  $F_3$  Piseco anticline, to the  $F_2$  recumbent isoclines of the Adirondacks (Figs. 6,7), and to the axis-parallel elongation lineations associated with these isoclines. The fact that all of these structural elements are parallel to one another indicates that they all shared some common kinematic experience during their tectonic evolution. In order to identify the common experience, we first note that all of the deformation in these ca 1150 Ma rocks must be of Ottawa age. Next we note that minor isoclinal folds in the outcrop lie with their axial planes in the foliation and their axes defining rod-like structures parallel to the N70W extension lineation. The low dip of the lineation suggests that the mechanism responsible for it was likely to have been thrust faulting out of the southeast and towards the northwest. This would have resulted in the stretching that elongated the quartz and feldspar. The thrusting could also have rotated earlier fold axes into the thrust plane and parallel to the direction of tectonic vergence. However, rather than rigid rotation of the axes, it is proposed that the early folds became too ductile to buckle and their axes began to "flow" as passive markers in the direction of tectonic transport, i.e., the recumbent fold axes began to undergo increasing curvature in response to the velocity gradients in the ductile flow field. The ultimate result of this process is sheath folds, and it is suggested that the isoclinal folds with axes parallel to the ribbon lineation represent parts of sheath folds. It is further suggested that the Canada Lake isocline may itself be a large sheath fold whose southern closure is obscured by Paleozoic overburden. The large  $F_3$  folds such as the Piseco anticline and the

Gloversville syncline are interpreted as structures formed by NS constrictional forces that arose in response to the EW elongation associated with the large scale, ductile thrusting of the Ottawa Orogeny. Later NNE upright folds of the  $F_4$  set were superimposed on the  $F_2$  sheath folds and the  $F_3$  corrugations by continued, but waning, NW-SE contraction associated with the NNE suture situated somewhere beneath, or beyond, the present day Coastal Plain. Finally, it should be remarked that the ribbon lineations and upright lineation-parallel corrugations described above are similar to those encountered in core complexes. The problem with this alternative is that the late- to post-tectonic Lyon Mt Granite emplaced at ca 1050 Ma is unaffected by these fabrics. Accordingly, we associate the fabrics with ductile Ottawa thrusting. Orogen collapse similar to core complex kinematics is thought to have taken place at ca 1050-1030 Ma.

- 43.2 Junction of Rt 8 and Rt 30 in Speculator. Turn right (south) on Rt 30.  
46.7 Roadcuts of Marble. Park on right (south) shoulder.

**STOP 7. MARBLE AND CALCSILICATE.** (30 Minutes). Exposed in roadcuts on both sides of the highway are examples of typical Adirondack marbles and their associated lithologies, ie, garnetiferous amphibolite, calcsilicates, and various disrupted blocks of both internal and external members. The disruption and boudinage attests to the extremely ductile behavior of the marbles. Also exposed are vertical, crosscutting, and undeformed veins of tourmaline-quartz symplectite. Besides calcite, the marbles contain diopside, tourmaline, sulfides, and graphite. The graphite has a biogenic carbon signature, and the marbles are thought to have formed inorganically via stromatolite accretion in an evaporitic environment. Some calcsilicate layers consist of almost monomineralic white, Mg-rich diopside crystals up to 10 cm in length. These are probably the result of metasomatism by fluid phases. The presence of the approximately Mg-pure assemblage



allowed Valley *et al* (1983) to calculate a fluid with  $X(\text{H}_2\text{O}) = .11-.14$  for the reaction. Elsewhere in the outcrop localized occurrences of wollastonite reflect the presence of  $\text{H}_2\text{O}$  as an agent lowering  $\text{CO}_2$  activity. In both cases, the metamorphic conditions were  $T=710^\circ\text{C}$  and  $P = 7$  kbar. These cases provide excellent examples of how the dilution of a fluid phase by a second constituent lowers activities and allows reactions to run to the right at P,T conditions below that for the pure fluid.

In the High Peaks region there exist many well-exposed examples of places where marble and calcsilicate xenoliths occur within anorthosite. This relationship documents that the marbles are older than ca 1150 Ma. It is likely that they formed during the same shelf sequence event that was posited for the Irving Pond quartzite precursor sands, ie, ca 1220-1200 Ma.

- 47.2.1 Roadcuts of steeply dipping metasediments.  
47.6.1 Long roadcuts in pink granitic gneiss interlayered with calcsilicates. The layering here is interpreted as tectonic and the granite as intrusive.  
48.7.1 Small, high roadcut on right (southwest) side of Rt 30. Park on right shoulder.

**STOP 8. MASSIF ANORTHOSITE AND FERRODIORITE OF THE OREGON DOME.** (30 MINUTES). This small, but instructive roadcut has outstanding examples of two important facies of anorthosite as well as a typical ferrodiorite dike associated with massif anorthosite. The best vantage point for examining the anorthosite and ferrodiorite is on top of the roadcut. Whole rock analyses of these rocks are given in Table 1.

On climbing to the top of the outcrop at its south end, one immediately sees a distinctive dark dike of ferrodiorite filled with plagioclase grains (white,  $\sim\text{An}_{43}$ ) and andesine xenocrysts (blue-gray,  $\sim\text{An}_{52}$ ); note reaction rims on the blue-gray andesine. The dike exhibits somewhat soft contacts and irregular veinlets shoot off in a fashion suggesting that the anorthosite was not yet wholly solidified

when the dike intruded. Farther up the outcrop two ~30cm-scale xenoliths occur in a ~5-10m-scale ferrodiorite; one of these is fine grained and the other coarse. The eastern contact of the ferrodiorite is situated just to the highway-side of the coarse xenolith. Calcsilicate xenoliths also occur in the ferrodiorite and are characterized by sulfidic staining near road level.

Farther up the outcrop the exposed rock consists entirely of anorthosite. A fine-grained, leucocratic facies forms a matrix to large (5-20cm long) crystals of blue-gray, iridescent andesine that is typical of massif anorthosite. In places these crystals appear broken as if disrupted from a larger mass. An example of a still intact mass is found at the far end of the outcrop where a raft of very coarse grained Marcy-type anorthosite with ophitic to sub-ophitic orthopyroxene sits in a matrix of fine-grained leucoanorthosite that clearly disrupts the raft at its edges. Close inspection of the matrix reveals that its small pyroxenes exhibit subophitic texture; hence the finer grained facies must be magmatic. Workers in Adirondack anorthosite have always noted that grain size reduction produces leucanorthosite similar to this this fine-grained facies, and examples of this may be seen at this stop. However, it is possible to distinguish between the grain size reduced and the fine-grained igneous varieties by noting that plagioclase in the former has the same composition as the coarse plagioclase, whereas the igneous variety is invariably 5-10% less anorthitic than the coarse plagioclase; in this case 52% versus 44% (Boone et al, 1969). Genetic interpretations and further details of the anorthositic suite are given in the main text.

The smooth, upper surface of the outcrop affords excellent opportunities to examine the garnet coronas or "necklaces" that are found throughout the Adirondack anorthosites. These are accounted for by the reaction



Spear and Markussen (1997) have studied this, and other garnet-producing reactions, and determined that they took place during isobaric cooling ( $P \sim 6-7$  kbar) from  $\sim 700^\circ$ - $\sim 630^\circ\text{C}$  during waning stages of the Ottawa Orogeny (1090-1030 Ma). McLelland et al (2001) have suggested that this reaction took place at ca 1050 Ma, which is the age of small, equant metamorphic zircons associated with the coronites. The zirconium was provided by Fe-Ti oxide, which accepts significant Zr into the Ti lattice site.

Recent SHRIMP II U-Pb dating of zircons from the ferrodiorite documents its age as  $1155 \pm 9$  Ma, and this sets a minimum age for the anorthosite. Thirty kilometers to the northeast at Gore Mt, charnockite enveloping, and showing mutually crosscutting relationships with the Oregon Dome anorthosite (Lettney, 1969), yields a SHRIMP II age of  $1155 \pm 6$  Ma. These results document that the anorthosite series of the Oregon Dome was emplaced at ca 1150 Ma. This age is indistinguishable from emplacement ages of the Marcy massif and marks the time of emplacement of the Adirondack AMCG suite.

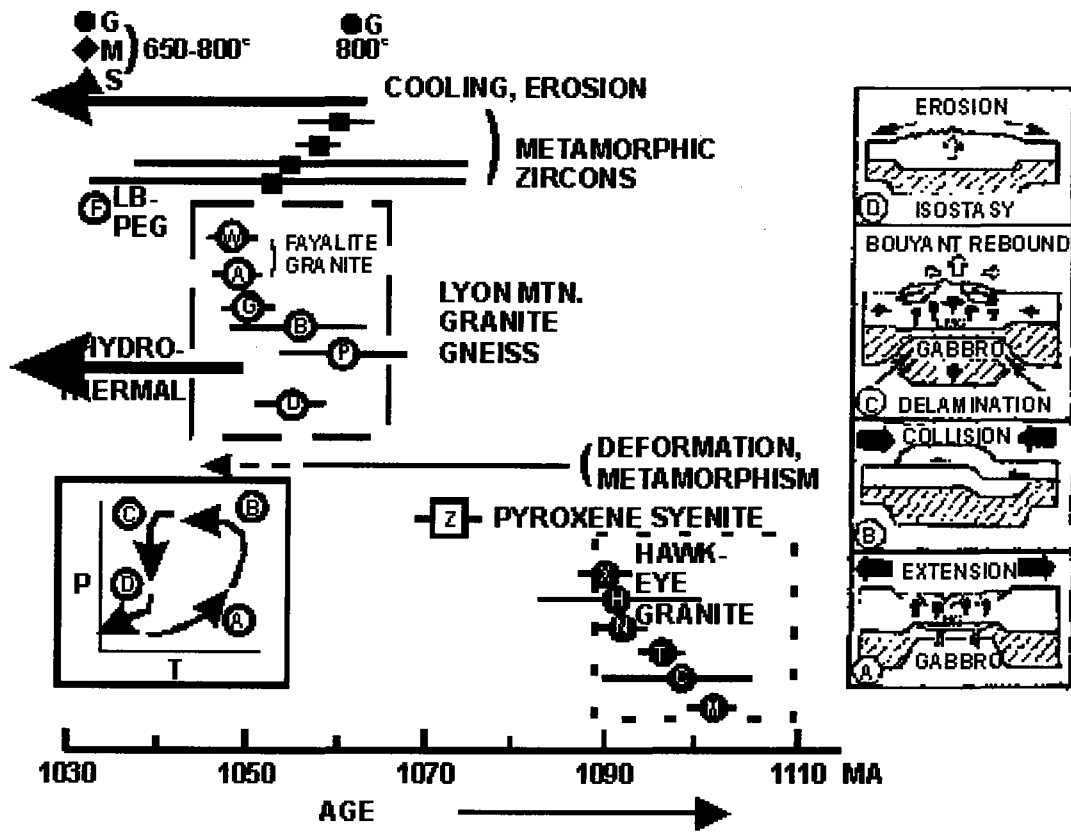
- 50.7 Minor marble, calcsilicate, amphibolite together with meter-scale layers of white quartz-feldspar leucosomes containing sporadic sillimanite and garnet. SHRIMP II ages of zircons from these outcrops indicate that the leucosomes crystallized from melt at 1180-1170 Ma. This interval is interpreted as the time of anatexis associated with the migmatitic metapelites of the region, and falls into the culminating Elzevirian Orogeny.
- 51.7 Junction of Rt 8 and Rt 30. Continue south on Rt 30.
- 52.2 Charnockite on the north limb of the Glens Falls syncline.
- 54.5 Entering the town of Wells that sits on a Paleozoic inlier dropped down at least 700m by a NNE trending graben structure.
- 55.0 Leaving town of Wells.
- 58.7 Pumpkin Hollow. As you round the big bend next to the river look for a large parking area on the right (west) side of Rt 30. Pull into it and park.

**STOP 9. MYLONITIC STRAIGHT GNEISS OF MIGMATITIC METAPELITE. (30 MINUTES).**

Large roadcuts on the east side of Rt 30 expose excellent examples of migmatitic metapelite identical to that seen at Stop 1, but now at a high grade of strain that has resulted in extreme ductile grain size reduction that has produced long tails on the feldspars and remarkable long ribbons of quartz now consisting of annealed subgrains in the process of annealing further into smaller grain size. The high strain has produced a platy mylonite that is a "par-excellence" example of straight gneiss. The alternating light and dark layers represent porcellaneous leucosome together with restitic biotite-garnet-quartz-feldspar  $\pm$  sillimanite. Numerous intrafolial isoclinal minor folds can be seen, especially in the leucosome. All of these observations make it clear that the "layering" seen in the outcrop has nothing to do with any primary or "stratigraphic" features, although some observers still try to make this assertion. The layering is strictly tectonic in origin and provides a fine learning opportunity with which to make the case with students (and others). Note the strong N70W lineation on foliation surfaces. A strong component of flattening has also affected these rocks making it difficult to interpret kinematic indicators.

At the south end of the outcrop the mylonitic migmatite is in contact with homogeneous, strongly foliated granitoids of the Piseco anticline. These granitoids have been dated at ca 1155 Ma, while the structurally overlying migmatites must be at least 1300 Ma of age since they are crosscut by the tonalites. The absence of any crosscutting contact emphasizes how ductile high strain can wipe out angular discordances.

The northerly dip of the mylonitic migmatite sequence is due to the fact that we are located on the southern limb of the Glens Falls syncline or the northern limb of the Piseco anticline. These dipping layers are locally broken across by ductile normal faults with brecciated pegmatite fillings. These features have not been dated, but a ca 1050 Ma age is expected.





**Trip A-7**  
**Paleontology and Stratigraphy of the middle Helderberg Group**  
**at Rickard Hill Road, Schoharie, New York**

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### **Introduction**

Although the Helderberg Group (Upper Silurian and Lower Devonian) has been the subject of study for over 150 years, there are still many questions unanswered, particularly in the areas of regional stratigraphy, sedimentology, paleontology, and paleoecology. Rickard's (1962) stratigraphic synthesis has been the foundation for all modern studies of these rocks. The lower units of the Helderberg Group (Manlius, Coeymans, Kalkberg and New Scotland formations) have served as the "textbook" example of deposition in shallow, epicontinental seas since Laporte (1969). However, even these well-studied units still hold surprises (see Ebert and Matteson, this volume and Ebert and Matteson, 2003). Middle and upper parts of the Helderberg Group have been less well studied (Ebert, 1987). In general, the lower formations may be regarded as recording stepwise, rather than continuous transgression (Ebert and Matteson, 2003). A similar interpretation probably holds for the upper units (Becraft, Alsen and Port Ewen) as well (Ebert, 1987). However, the interval between these transgressive sequences is poorly understood in comparison. It is these middle units which will be examined on this field trip.

### **Stratigraphy of the Rickard Hill Road Outcrops**

The upper portions of the Kalkberg Formation (Lower Devonian) are clearly exposed in the road cuts on Rickard Hill Road, near Schoharie, New York. Kalkberg strata comprise alternations of generally fossiliferous, decimeter-scale, muddy limestones (micrites, wackestones and packstones) and poorly fossiliferous, dark, calcareous shales. In general, the Kalkberg is coarser and more fossiliferous in the uppermost few meters. The Kalkberg Formation is abruptly overlain by the Becraft Formation.

The Kalkberg-Becraft contact is most easily seen on the more weathered northern side of the road (Fig. 1). Here, the contact stands out by the change in relative resistance to weathering and erosion of the two units (Becraft is more resistant.) and by the presence of a thin (1-3 cm)

reentrant that marks the contact. This reentrant is formed by the relatively rapid weathering of a K-bentonite or altered volcanic ash, which is reported for the first time in this paper. Other K-bentonites have been well-documented in the lower Helderberg Group (Rickard, 1962; Smith, Berkheiser and Way; Ebert, Applebaum and Finlayson, 1992; Milunich and Ebert, 1993; Tucker, et. al., 1998). Overlying this K-bentonite, the lowest bed of the Becraft Formation contains rare intraclasts of Kalkberg lithology, including exhumed fossils with adhered matrix. These features suggest that the interval that separates the lower from the upper transgressive sequence is, at least in part, disconformable. Further study of this contact is still in progress.

The Becraft Formation (Lower Devonian) is made up entirely of thick beds of coarse, fossiliferous limestone (grainstones). On many weathered joint surfaces, very well-developed cross-stratification is apparent.



Fig. 1 The disconformable contact between the Kalkberg (below) and Becraft (above) formations on Rickard Hill Road. The prominent reentrant at the contact is a K-bentonite (altered volcanic ash). Four future earth science teachers (SUNY Oneonta students) provide scale.

### Paleoenvironments

Lithologic and paleontologic aspects of the Kalkberg Formation suggest that this unit was deposited on an open marine shelf with nearly ideal conditions for benthic fauna (See also the general interpretations in Isachsen, et. al., 1991, p. 109; and Linsley, 1994). In general, the sea floor was below fair-weather wave base, but shallow enough that the waters were well oxygenated and contained abundant suspended food for the filter-feeding organisms. The

lithologies and sedimentary structures present (rare ripples and cross lamination) indicate that this shelf was periodically affected by storms, which winnowed the bottom and created shell pavements. Ordinarily, the sea floor was blanketed with soft mud as evidenced by the “snowshoe” strategies adopted by much of the fauna.

In contrast, the Becraft Formation was deposited in significantly shallower water, well-above storm and fair-weather wave base. Tidal currents (Ebert, 1987) and waves winnowed finer sediments from the sea floor and produced well-sorted sands and gravels comprised of skeletal material. Sedimentary structures (trough cross stratification and upper stage planar stratification) indicate shallow, energetic conditions. Net accumulation of sediment was probably relatively low as indicated by the well-sorted and abraded nature of the skeletal debris. In the lower parts of the formation, abundant holdfasts of the enigmatic crinoid *Aspidocrinus scutelliformis* (another “snowshoe” adaptation) may have been reworked and concentrated from muddier environments.

### **Paleontology and Paleoecology**

The Kalkberg and New Scotland formations, comprising the upper part of the lower Helderberg Group and the middle part/transition to upper Helderberg, are the most fossiliferous units in the group. Rickard (1962) reported that these formations have yielded over 300 species. Linsley (1994) indicates that there are at least 65 species of brachiopods alone. The fauna is dominated by epifaunal suspension feeders such as brachiopods, bryozoans, echinoderms, sponges, tabulate corals and pelecypods. Most of the community exhibits “snow shoe” types of adaptation to soft substrate. This is best displayed by the broad, flat valves of the strophomenid brachiopods, which are extremely common. Colonial bryozoans, tabulate corals, and solitary crinoids generally availed themselves of localized hard substrates for attachment, such as brachiopod valves or winnowed shell pavements. Motile benthos includes trilobites and gastropods, which probably functioned as scavengers on the Kalkberg sea floor.

The Becraft fauna is markedly more abundant than that of the Kalkberg, but it is considerably less diverse. Disarticulated echinoderm debris predominates, with abundant brachiopods, lesser numbers of bryozoans, and rare stromatoporoids and cephalopods. Again, the fauna is dominated by suspension feeders; however, the Becraft fauna exhibits distinctly different adaptations to the environment. Most of the brachiopods are thick-shelled and relatively equidimensional – a “roly-poly” or “weeble” type of adaptation to a firm, but frequently mobile substrate. This habit is exemplified by several species of uncinulids. It is most easily seen on the characteristic gypidulid, *Gypidula pseudogaleata*, which also possesses an unusual thickening in the umbonal region as a means of remaining upright. Rare, large (up to 6 cm diameter) gastropods (crushed), fragments of trilobites, and conches of orthocone cephalopods are also present. In general, the Becraft community indicates a shallow, well-agitated shoal environment. Sediments were reworked frequently and net rates of deposition were generally low.

In addition to the unusual holdfast of *Aspidocrinus scutelliformis*, which is present in the lower part of the formation, the Becraft at Rickard Hill Road also displays *in situ*, root-like holdfasts of a crinoid, which may be assignable to the genus *Clonocrinus*. These holdfasts are

best seen in the uppermost 1-2 meters of the formation and are particularly well displayed on the exposed bedding plane at the top of the outcrop on the north side of the road. Many of these root-like masses exhibit a preferred orientation, which may reflect an attempt by the crinoids to anchor themselves against a prevailing (probably tidal) current.

### **Fossil Collecting and Teaching Activities**

The abundance and diversity of fossils available at Rickard Hill are such that large collections may be acquired in a short time. Individual fossils, freed from the matrix by weathering, may simply be picked up from the talus at the foot of the outcrop. For some species, it may be possible to collect enough individuals for classroom sets (i.e., 12 samples for a class of 24). Slabs of the Kalkberg Formation are also easily collected from the float. These commonly display a variety of fossils on a single bedding plane. Such samples may be particularly useful because they illustrate the concept of assemblage diversity, as well as ecological interactions. For example, bryozoans may be found encrusting on valves of brachiopods.

Although it may be possible to collect individual fossils from the Becraft, it is difficult owing to the relative resistance of the formation. Therefore, slab samples, commonly with abundant brachiopods on bedding planes, are a better bet. However, the thick-bedded nature of the Becraft generally means fairly large samples.

Crinoidal debris predominates in the Becraft. Because echinoderm plates break with cleavage, fresh surfaces of Becraft samples appear sparkly. Such surfaces may be somewhat misleading for inexperienced students, who may think that the rock has a crystalline rather than bioclastic texture.

Once fossils have been collected, how can they be used in the classroom? Activities using fossils address two of the specific understandings in Standard 4 (Science), Key Idea 1 in the Regents Earth Science Core Curriculum:

|  |   |
|--|---|
| 1.2i The pattern of evolution of life-forms on Earth is at least partially preserved in the rock record.                               |   |
|  | ◆ Fossil evidence indicates that a wide variety of life-forms has existed in the past and that most of these forms have become extinct. |
| 1.2j Geologic history can be reconstructed by observing sequences of rock types and fossils to correlate bedrock at various locations. |   |
|  | ◆ The characteristics of rocks indicate the processes by which they formed and the environments in which these processes took place.    |
|  | ◆ Fossils preserved in rocks provide information about past environmental conditions.   |

Activities addressing Understanding 1.2i might include –

- ◆ Having students recognize the different species in an assemblage
- ◆ Measuring the dimensions of a large number of a single species to identify variations within a population
- ◆ Comparing diversity of two different assemblages
- ◆ Making comparisons between the fossils (extinct) and modern examples of the same types of organism.

Activities related to Understanding 1.2j could include -

- ◆ Having students make inferences regarding environmental conditions based on the ecological needs of the fossils in an assemblage. Comparisons may be made with the ecological requirements of living representatives of the same groups.
- ◆ Environmental information may also be gleaned from the way the fossils are preserved. For example, if fossils are preserved in life positions, this suggests little physical reworking in the environment by waves and currents, therefore relatively quiet, and possibly deep conditions (e.g., some examples from the Kalkberg Formation). Conversely, if fossils are disarticulated and abraded, the environment was likely turbulent and shallow (Becraft Formation).

### Other Geologic Features of Interest

On the top of the outcrop on the north side of the road, excellent examples of glacial striations and polish may be observed. An interesting aspect of these striations is that they comprise two sets which are oriented at nearly  $90^\circ$  to each other (Fig. 2).

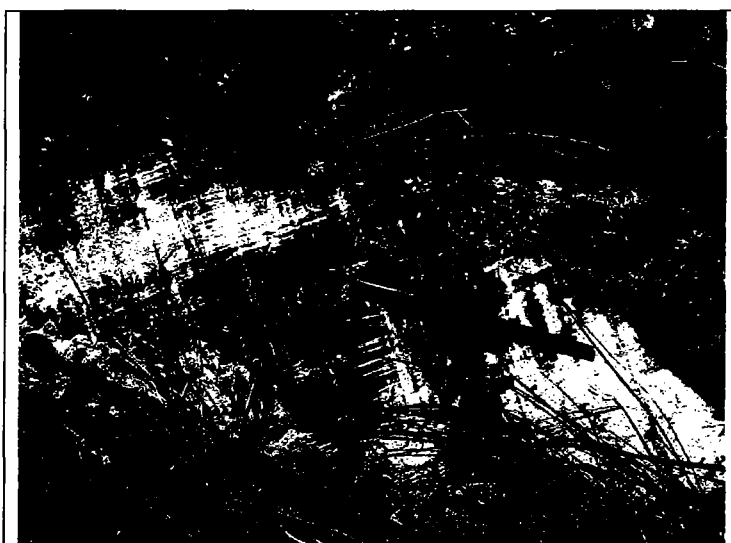


Fig. 2 Two sets of glacial striations at nearly  $90^\circ$  on the top of the road cut on Rickard Hill Road. Pen provides scale.

The northern outcrop overlooks the large, active quarry in Schoharie. Visits to the Rickard Hill road cuts may be interrupted on occasion by blasting within the quarry. At these times, employees of the quarry will ask visitors to leave the outcrop.

### **Conclusion**

Collections of fossils for classroom use are acquired fairly easily at Rickard Hill Road. Samples from the Kalkberg and Becraft formations may be used to design a variety of activities in which students utilize multiple science process skills to explore a snapshot of the history of life on Earth. The abundance and diversity of fossils, along with a variety of other geologic features, make the outcrops on Rickard Hill Road well suited for class field trips also. Ample parking is available on wide shoulders and there is a safe distance between the face of the exposure and the road.

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### **Directions to Rickard Hill Road Outcrops**

Because this trip comprises a single stop, directions to the outcrop are provided in lieu of a road log. Directions are given from the closest exit on Interstate 88.

- ◆ Take Exit 23 for Schoharie and Central Bridge, Rts. 7, 30 and 30A.
- ◆ At the end of the ramp, turn onto Rt. 30A south toward Schoharie.
- ◆ When Rt. 30A and 30 split, stay on Rt. 30 to the village of Schoharie.
- ◆ In Schoharie, you will pass a school on the left, which is set back slightly from the road and uphill. Immediately after the school, there is a blinking traffic light.
- ◆ Turn left at the light.
- ◆ Almost immediately after turning off Rt. 30 the road will split. The road on the right goes uphill and there is a sign that indicates that this is Rickard Hill Road (Schoharie County Rt. 1a).
- ◆ After a few tenths of a mile, you will encounter outcrops on both sides of the road. These are exposures of the Coeymans Formation. The upper surfaces of these outcrops display clear glacial striations and are overlain by glacial sediments.
- ◆ Continue uphill to the very large outcrops on both sides of road. There is ample parking on both sides of the road. Description of the outcrops is in the narrative for this field trip.

If time permits, we may also visit the site of the Cave House Museum of Mining and Geology, a geoscience education facility which is currently under development. This facility will comprise parts of the Howe Cave Quarry, the old Lester Howe Hotel and the original entrance to

Howe Caverns. Strata of the lower Helderberg Group, various cave and karst features, a thrust fault, as well as spectacular fields of ripple marks and glacial striations are some of the features that will be available for study and exhibition. See Ebert and Matteson (this volume) for directions to the site and for contact information regarding the Cave House Museum.



**SUNDAY SEPTEMBER 14, 2003**

**TRIP B-1**

**A VISIT TO THE LAND OF GILBOA**

Robert Titus  
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**Introduction**

Hartwick College has been built upon Oyaron Hill. The slopes here offer a number of outcrops which intermittently expose almost 200 m of strata. These record a complete facies pattern spanning environments that range from offshore marine, to coastal, to well inland on the Catskill Delta complex. All sedimentary environments are well exposed and so Oyaron Hill offers a compact opportunity to visit nearly all facies of the famed Catskill Delta, herein called the "Land of Gilboa."

***Gilboa, Back Then\****

The 23<sup>rd</sup> of Caligulary, 365,841,916 BC, high summer. There are more than 400 days in the year in this time and 14 months. One of those extra months is Caligulary. We are the mind's eye, and we are drifting high above what will someday be the town of Morris. Below us is the dark blue of the Catskill Sea. Ahead of us, the curvature of the Earth's surface is bringing a rising eastern horizon into view. First to appear are the rugged white summits of a very tall mountain range, the Acadians. As we continue eastward, we see that white extending down to the bottom of the snow line, and then steel blue slopes appear beneath.

Ahead the blue slopes grade downward into purple and then brick red foothills. Below those, as the horizon unrolls, we see light, sandy lower slopes. Then, finally, the greens of the delta foliage appear. It is the hottest part of the summer, here in the Southern Hemisphere, and ahead of us, it is the monsoon season. Ahead of us, great masses of terribly hot air rise as plumes above the delta. This natural chimney creates a draft, and behind us storms are being drawn from off of the sea and on toward the interior of the delta. Back there is the first of three very powerful lines of thunderstorms. On modern weather radar images these would be blinking masses of yellow, and orange, surrounded by dark green. What we actually see on the western horizon are great billowing dark clouds, periodically illuminated by flashes of lightning. It is a frightening sight and it would be scarier if we could see the two lines of storms that follow the first; they are worse.

Now we can see that we are approaching the shores of the Catskill Delta. Below us, the Catskill Sea is shallowing and the dark blue is becoming paler. We are the mind's eye; we can go

as fast as we want and as high as we want<sup>1</sup>. We drop down closer to sea level and speed forward. Soon we cross the delta coastline, extending to the north and south. The beaches are not like the brilliant white sands we know today on the East Coast; these are of a dull brown color.

Behind the beaches is the delta interior. A complex of interfingering distributary streams flows lazily toward the coast. We pick out one of the biggest and finest among them and drop down to just above its mouth and then we fly up the river; it will be our avenue into the delta interior. It is immediately obvious that this landscape has been suffering from a serious drought. The water is low and the riverbanks are composed of parched bleached sands.

The river flows lazily down narrow stretches, interrupted, here and there, by wider and deeper pools. It is in the pools that we observe primitive fish nervously swimming in tight circles; the drought conditions seem to have them on edge. The fish are heavily armored with dermal bone and are often hard to recognize as even being fish. Only the fins give them away. Some stretches of streambed are densely littered with colonies of clams. Their shells gape widely as they struggle to glean food from the slow moving currents.

Beyond the banks of the stream, the land of Gilboa is nearly silent. The foliage is made up of tall but primitive looking trees. In the ground cover below are numerous millipedes, centipedes and spiders. These are the only familiar looking elements of the Gilboa Forest.

We are the mind's eye, we can go anywhere and do anything; we accelerate and rush along our stream channel, heading up the river. The banks zoom by in a blur as we go 50, then 60 and then 75 mph. We rise up high into the air and look down upon the vast expanse of dry flatness below. All around, the trees of the delta are pale and yellow, suffering from the drought. All around, there are light colored shrunken pools. It is a desolate, hot and dry delta; it won't be for long.

Behind us, those lines of thunderstorms have continued their approach, covering ground nearly as fast as we have. The flashing of lightning illuminates the darkness of the clouds. But we are the mind's eye and lighting cannot harm us; we drop down again and soar forward even faster. The banks fly by at a giddy pace. But soon the greenery is replaced by a blur of yellow, red and tan and the banks steepen and rise above the river. We are approaching the lower slopes of the Acadian Mountains and these are heaps of gravel that have washed down out of those mountains.

We slow down and gradually enter into a large steep-walled canyon. It takes us up the slopes and into the mountains. There is no green up here; it has instead abruptly become a dead landscape. A brick red earth is soon replaced by a purple color. Then the canyon walls are gray and even black. It is a spooky place, a dark and shadowed natural maze of gray ravines, and it would be very easy to get lost in this dark and alien landscape.

But we are the mind's eye and we cannot get lost. We continue our ascent, always following the widest and steepest ravines as we climb the middle slopes of the Acadians. There is no life up

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<sup>1</sup> This section adapted from Titus (in press).

here at all; Nature has not yet solved the evolutionary problems of living at such elevations. We ascend; we rise to 5,000 feet, then 10,000 feet and then even 20,000 feet. Pockets of snow bring the contrast of gleaming white to the surrounding dark. Then there is some ice, and then it is all ice and snow.

We are the mind's eye, we continue upwards until we reach a summit at 27,653 feet. This is not quite a Mt. Everest, but there are taller mountains to the north and to the south. This, however, is as far as we shall go. We turn around and, below us, is spread out all of the Catskill Delta, and beyond that the distant Catskill Sea. It is a grand view, to say the least, but it is marred by those three lines of thunderstorms. They have all crossed the coast and are making a mess of the delta ecology down below. There is no sound of thunder up here, but there is the continuous flashing of lightning below. The first line of storms is lapping angrily at the base of the Acadians, frustrated in its inability to ascend those high slopes.

It is time to retrace our path, and we begin our descent. We swoop down the slopes at nearly 100 miles per hour. Those mysterious canyons fly by faster than we can see. Soon we have almost reached the bottom and there we enter into that first line of storms. It is a most dreadfully furious tempest. Torrents of horizontal rain whip back and forth in the intensity of the wind. The pounding rain bounces off of those heaps of gravel. The previously dry sediments thirstily soak up the water.

We, the mind's eye, continue westward. We leave the first line of storms and find ourselves out on the delta. The place is a mess. Trees have been blown down in all directions. Shattered trunks and broken bits of foliage have been thrown about randomly. Pools of dirty water have formed where, just an hour earlier, the land was parched. Ahead of us is the second line of storms. It towers as a great black mass high above the delta green. The ominous clouds have a dangerous swirling appearance; there is great power there. The inevitable flashes of lightning are worse than before.

We press forward and the results are to be expected. Once again, terrible winds whip back and forth, driving sheets of horizontal rain, first one way, and then abruptly another. If we were not figments of the imagination, these drops would sting our faces. Those trees that were left standing from the first line of storms are now swinging back and forth violently. Many are breaking off at the base of their trunks. Broken shards of trees are careening across the flat delta. Pools that had just filled up with water are now themselves filling up with broken greenery.

This nightmarish scene ends suddenly as the line of storms passes. The sun soon comes out and nature is calm once again. But the broken delta, all around us, is gurgling and hissing with the weight of the water that has fallen into it. We find ourselves standing next to one of the main streams of the Catskill Delta and its waters are flowing gray, then brown and then red, depending on what masses of mud are washing by. What's worse is that the dirty waters are quickly rising up the banks. Soon, the growing power of this flood is manifest. The water swirls and then foams and then white caps appear. Now the water seems to hum like electricity as it rushes by.

Beneath the surface, fish are quickly losing their struggles for life. These clumsy, heavy animals are no match for the currents; their gills cannot function with all the mud. Death is a slide, not a fall; they quietly give up their hold on life, drift upwards, and soon just float like corks upon the swirling flows. The clams fair better, they are being swept along, tumbling in the currents. They stubbornly keep their shells closed tight and that keeps them alive, at least for now.

Now the snarling waters cut into the banks of the streams and hungrily devour great masses of sediment. The flows turn red and rise above the banks, and then the dense waters flow across the whole breadth of the flat Catskill Delta. Curiously, it is in this, their apparent moment of triumph, when the flood currents seem to give up much of their power. As they cross the stream banks, the foaming channel swirls grade seamlessly into the almost languid flows of the flooded delta. Red and brown waters peacefully rise up the broken trunks of the storm-tossed Gilboa Forest.

That's when the third line of storms breaks. The rain of the first line soaked into the dry delta sands; the second line glutted the streams with dirty water, and began the flooding. This, the third line, is the worst. The power of these storms would be bad enough, but they have generated a new complication. Broken tree trunks have concentrated in the stream channels, and there are so many of them that they began to "knit" into log jams. These catch on the edges of the channels and form log dams. Then the water behind those dams rises even higher. Soon the Catskill Delta is a great, fresh water inland sea.

This, at last, is the worst of it. But, even as the waters begin to subside, there are more dangers. Those poor clams, which have been tumbling along in the turbid flows, now come to rest. For them, this is the climax as they are soon buried under meter thick masses of fresh mud. They face a stark choice: dig or die.

Late in the night, near midnight, the skies above the Catskill Delta are clear and a wine colored full Moon shines down upon a land that has been devastated. The day's horrible flooding is over and the waters are actually beginning to subside. At the mouths of the various streams large plumes of dirty water are spreading out into the Catskill Sea. This dirty fresh water floats upon the salt water. Its surface is littered with broken plants and dead fish.

Meanwhile, back inland, the streams are draining off of the delta. The floodwaters are leaving a thick gooey layer of mud behind. Broken and splintered tree trunks rise above that mud. There should be blinking red lights all across the delta; emergency crews should be rescuing those still in peril. But this is the Devonian and the Catskill Delta is a soul-less land. There are no rescuers, nor are there any rescued. Delta creatures are tending to themselves as best as they can. None of them understands what has happened today. In fact, very few of them can even remember the day's terrible events. But all is not lost. In many of the stream channels, the buried clams have turned out to be very able burrowers and virtually all of them have begun to dig themselves out. They open up their shells and hungrily feed upon the currents that flow by. On this evening, the waters are unusually rich in food. Life goes on.

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### **Appendix: TRIP LOG FOR STRATA OF OYARON HILL**

Field trip assembles at parking lot behind Miller Hall. We hike down to Clinton Street and from there to the bottom of the road where it intersects with Chestnut St.

#### **STOP 1. BAGNARDIS SHOE REPAIR**

The outcrop exposed here is of the Unadilla Formation. The sequence is composed of moderately well bedded sandstones and some shales. There are two levels which represent shell hashes. These are interpreted as lag deposits, dating back to times of unusual current activity. There is at least one wave ripple marked horizon. The beds have been thoroughly searched for fossils these past 30 years and not many remain. In the past a brachiopod and clam assemblage dominated. Nautiloids, snails, trilobites, rare bryozoa, trace fossils were also found. The sequence was first described by Prosser in 1899. He listed 13 species, mostly brachiopods and clams. These strata are interpreted as representing a shallow water marine setting lying some distance offshore of the prograding delta sequence.

#### **STOP 2. HARTWICK MAINTENANCE BUILDING**

When this outcrop was exposed, about ten years ago, it was hoped that the site would yield numerous fossils. Sadly, this was not the case. Most of the sparingly exposed strata between here and Miller Hall are sparsely fossiliferous at best. These facies may represent a "dead zone" lying just offshore of the delta river mouths.

#### **STOP 3 THE MILLER HALL OUTCROP**

The Miller Hall outcrop is situated immediately above a thick sequence of mostly unfossiliferous black shales. These were exposed when Miller Hall was recently expanded. These are apparently the very first non-marine deposits in the local Catskill sequence and they thus mark the moment in time when progradation of the Catskill Delta complex had reached today's Oneonta. The facies here are therefore most likely to be the most nearshore that is possible. This is of some importance as this level thus marks the lowest level of the Oneonta Formation. This part of the sequence was not exposed when workers such as Vanuxem (1840) or Prosser (1899) studied here.

This fine exposure reveals several distinctly different nearshore lithologies. First there is the massive sandstone unit that thins out to the right. Then there are the black shales that underlie the entire outcrop. Between the two lithologies is a thick bedded, rusty sandstone with bedding gently inclined toward the massive sandstones.

The massive sandstone appears to be a stream channel, possibly a sand filled abandoned channel. The base of the deposit rises to the right, apparently up the slopes to the bank. Beyond and

farther to the right is what may be an overbank crevasse splay deposit. In the past the very bottom of the exposure has displayed drag marks which are east to west oriented. These are currently buried.

The black shales are sparsely fossiliferous. A few large horizontal trace fossils have been found along with the brackish water clam *Paleoneilo sp.* There are several horizons of deep scouring in the otherwise very thinly laminated sequence.

The third lithology here is composed of rusty colored sandstone beds that are inclined toward the stream channel. These strata are richly fossiliferous with fragments of land plants. This deposit may represent a natural levee or bar finger deposit.

All in all, the Miller Hall outcrop is related to settings on the lower Mississippi River delta. An analogous location is found on the West Delta 15" Quadrangle. There the Tom Loar Pass distributary flows past Bob Taylors Pond and Zinzin Bay. A channel deposit along with natural levee deposits can be expected to overlay lagoonal black shales.

#### Stop 4          Upper Oyaron Hill

The remainder of the slopes of Oyaron Hill rise from elevations of about 435 to 525 meters. Throughout this section are a number of sandstone ledges providing good exposures of the Oneonta Formation. In between the sandstones are less well-exposed sequences of red shales and siltstones. The Oneonta was named by Conrad in 1841. Vanuxem (1842) described this location.

The sequence seems to represent the distal reaches of the Catskill delta sequence. The sandstone ledges are likely to be the deposits of meandering distributary streams while the fine-grained deposits are most likely overbank deposits. The sandstones have occasionally been observed lying with erosional truncation of underlying overbank deposits.

The sequence here displays a number of fining upwards sedimentary cycles of the sorts that have been described by Bridge and his colleagues among others (see for example Bridge, 2000, Bridge and Gordon, 1985, Gordon and Bridge, 1987, Willis and Bridge, 1988 and Bridge and Willis, 1994).

Within the sandstones are abundant trough cross beds, especially near the bottoms of the ledges. Also found are planar cross beds and horizontally laminated sandstones. Near the tops of the ledges there are occasional ripple marked horizons.

Of special interest is the presence of vertical clam escape burrows (probably *Archanodon catskillensis*). These are found at the "Table Rock" ledge near the top of the hill. The clams seem to have been buried by a meter of sand at the time of a waning flood (see Thoms and Berg, 1985; Bridge et al., 1986 and Gordon, 1988). They produced well-defined meniscus patterns as they worked their ways up through the fresh deposits.

The best overbank deposits are found at the top of the hill, along the sides of the soccer field. Here red shales and fined grained red sandstones are seen in abundance. The deposits are rich in root casts and also display rare *Isopodichnus* trace fossils. There are ripple marks and possible soil horizons. At the very top is a sequence of alternating shales and sandstones which seem to record a sequence of occasional flood events. One, exposed on the surface of an enormous displaced boulder, has several fossil logs on it. These were apparently displaced during a flood event.



**TRIP B2**  
**Grabau's "Transition Beds" – Key Elements**  
**in a Radical Revision of Helderberg Stratigraphy**

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*... transitional beds from the Manlius to the Coeymans extend to the foot of the Coeymans ledge...  
 The fauna of the transition beds... represents oscillating conditions between the Manlius and the  
 Coeymans.*

Amadeus W. Grabau (1906)

*Recognition of an erosional interval between the Cayugan and Helderbergian has been tardy ... as  
 late as 1906... Ulrich and others were talking about AManlius transition beds≅ in east-central  
 New York and the Helderberg region.*

G. H. Chadwick (1944)

*The lateral gradation which exists between the Coeymans limestone at Cherry Valley and the  
 Olney-Jamesville members of the Manlius formation constitutes nearly indisputable evidence for  
 the contemporaneity of at least parts of the Manlius and Coeymans formations.*

L.V. Rickard (1962)

*The Manlius-Coeymans cryptic unconformity in eastern New York is particularly well documented  
 by the erosional loss of as much as 4 metres of section encompassing more than three PACs...*

P.W. Goodwin and E.J. Anderson (1988)

## INTRODUCTION AND PREVIOUS WORK

The nature of the contact between the Manlius and Coeymans formations is a recurring theme in the literature of the Helderberg Group. In the early twentieth century, Grabau (1906) interpreted the contact as gradational, with a distinctive zone of Atransition beds≅ present between true Manlius and true Coeymans. Grabau=s transition beds are best developed in eastern New York in and near the Schoharie Valley and appear to be absent from other areas (Fig. 1).

In succeeding decades, the contact was regarded as a disconformity, with views varying as to the magnitude of hiatus. Goldring (1935, 1943) indicated that as early as 1927, Chadwick saw evidence for some hiatus (see also Chadwick 1944). In central New York, Smith (1929) described the contact of the Jamesville Member of the Manlius Formation with the overlying Coeymans as unconformable. Logie (1933) felt that the erosion was significant with the Elmwood, Clark Reservation and Jamesville members of central New York absent in eastern parts of the state via erosion. Chadwick (1944, p. 152) reported the contact as Airregular, undulating≅ and concurred with Logie that the absence of the higher members of the Manlius in eastern New York was the result of significant erosion.

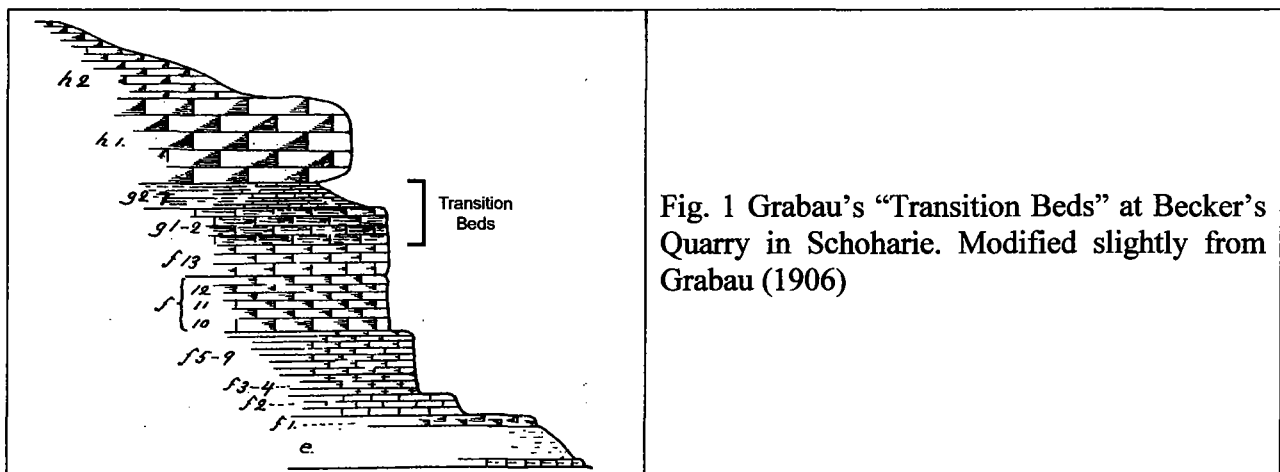


Fig. 1 Grabau's "Transition Beds" at Becker's Quarry in Schoharie. Modified slightly from Grabau (1906)

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Rickard (1962) reinterpreted the Manlius-Coeymans contact as sharp, but conformable in eastern New York. In central New York, he saw the contact as complex with portions of the Coeymans Formation (e.g., Dayville Member) and higher parts of the Manlius (Elmwood, Clark Reservation and Jamesville members) and other members of the Coeymans (Deansboro/Ravena members) intertonguing (Fig. 2). Rickard's stratigraphy became the foundation for nearly all succeeding work on the Helderberg Group. The most notable example is Laporte's (1969) reconstruction of depositional environments.

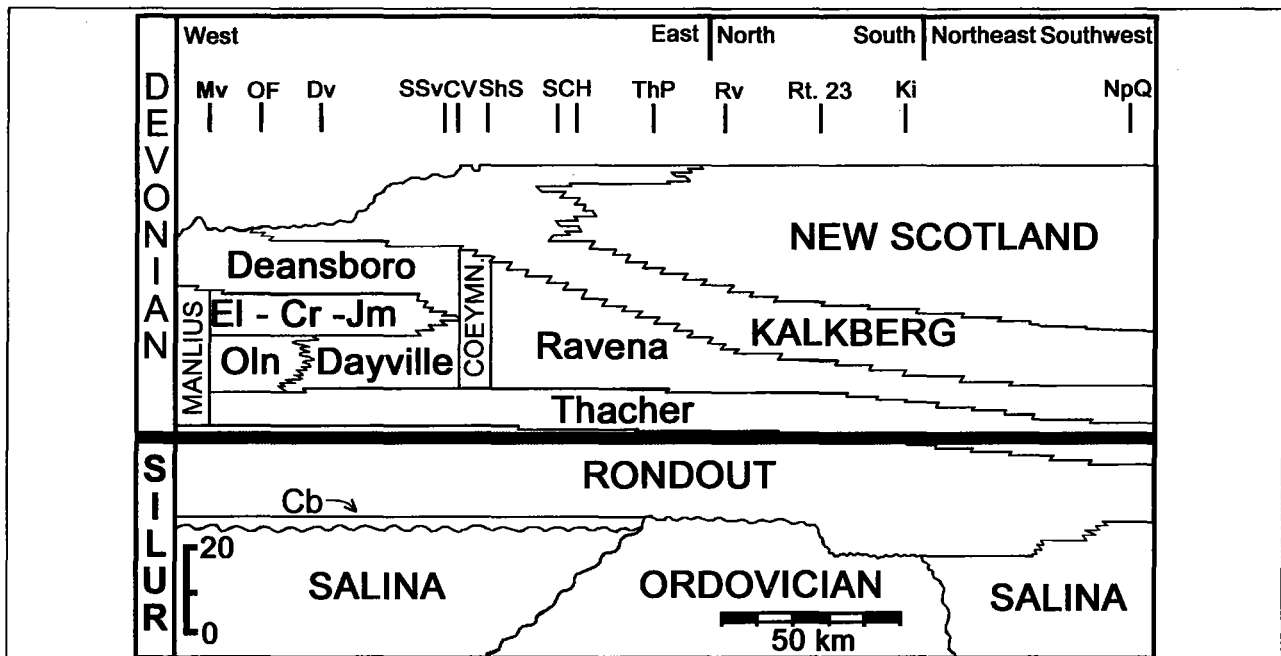


Fig. 2 Rickard's (1962) synthesis of lower Helderberg stratigraphy. Figure is from Ebert and Matteson (2003), modified from Barnett (1977). Formation and group names are shown in upper case. Member names are in lower case. Coeymn. = Coeymans Formation. Members of the upper portion of the Manlius Formation, in ascending order, are Oln = Olney Member, El = Elmwood Member, CR = Clark Reservation Member, Jm = Jamesville Member. Cb = Cobleskill Limestone. Locations are abbreviated as Mv = Munnsville, OF = Oriskany Falls, Dv = Dayville, SSv = Salt Springville, CV = Cherry Valley, ShS = Sharon Springs, SCH = Schoharie outcrops - Howe Cave Quarry and I-88, ThP = Thacher Park, Rv = Ravena, Rt. 23 = Catskill, Ki = Kingston, NpQ = Nearpass Quarry, New Jersey. Vertical scale in meters is approximate. Silurian - Devonian boundary is based on biometric trends in *Ozarkodina remscheidensis* (Barnett 1977).

Rickard (1962, p. 105) acknowledged the possibility that a *Apost-Jamesville* (sub-Coeymans) disconformity in central New York might extend into the middle of the Coeymans in east-central New York and to either the top or base of the Coeymans in the Hudson Valley. However, he discounted this interpretation based on the absence of an obvious faunal or physical break and the presence of the transition beds at Cherry Valley. Later, Rickard and Zenger (1964, p. 51) reported that Aa layer of pebbles and abundant *Favocites* sp. occupies the position of the Jamesville with Dayville and Deansboro crinoidal limestones with *Gypidula coeymanensis* below and above in portions of the Richfield Springs 15' Quadrangle. Whereas this may be viewed as evidence for erosional truncation of the Jamesville below the Coeymans, Rickard and Zenger (1964) retained the interpretation of Rickard (1962) of gradational facies relationships, as did Laporte (1969).

In the 1980's, Goodwin and Anderson published a series of papers outlining and applying their hypothesis of Punctuated Aggradational Cycles (PACs - a type of small scale allostratigraphic cycles) to the Helderberg Group in New York and elsewhere in the Appalachian

Basin. Where the Coeymans Formation overlies the Thatcher Member of the Manlius Formation, Goodwin and Anderson (1988) interpreted the contact as a minor, cryptic unconformity. They documented the progressive cutting out of more of their PACs from central New York eastward. However, they continued to support coeval relationships between higher parts of the Manlius and the Coeymans formations by correlating PACs through these units.

## **SIGNIFICANCE OF THE PROBLEM**

Are these conflicting interpretations of unconformity or gradational contact mutually exclusive, or does the truth lie somewhere between the extremes? Clearly, early workers like Smith, Logie, Chadwick and later Goodwin and Anderson saw physical evidence for unconformity. In contrast, Rickard presented faunal evidence in a coherent stratigraphic framework which supported gradational relationships. Laporte's (1969) sedimentologic study lent additional support. Rickard's interpretation was strongly influenced by the presence of Grabau's transition beds between the two formations. Those that described the Manlius - Coeymans contact as unconformable either dismissed the transition beds with derision or did not mention them at all. It seems apparent that the existence, nature and distribution of these enigmatic transition beds may hold the key to understanding the stratigraphic relationships between the Manlius and Coeymans formations.

One might argue that understanding the nature of the contact between these two formations is a rather trivial matter. However, there are implications beyond these competing interpretations. The Siluro-Devonian sequence of New York State is the Appalachian Standard Succession (Johnson and Murphy 1969). Within this Standard Succession lies the Silurian - Devonian systemic boundary. Rickard (1962; 1975) and Barnett (1971; 1977) placed the boundary within the Helderberg Group in the vicinity of the Rondout - Manlius contact. However, this placement is not based so much on index fossil biostratigraphy as it is on an attempt to be consistent with Rickard's stratigraphic framework of coeval facies. Earlier workers (e.g., Clark 1889; 1900; Schuchert 1900) placed the boundary considerably higher, namely at the Manlius - Coeymans contact. So, there is more at stake than geologica esoterica. The accurate identification of a major biostratigraphic boundary and its attendant global correlation hang in the balance. Without accurate placement of the boundary, correlations between the Standard Succession in New York and other areas within the Appalachian Basin become problematic and comparisons to global sea level curves become increasingly suspect.

## **TRADITIONAL STRATIGRAPHY**

For over 150 years, the formations of the Helderberg Group have been the subjects of stratigraphic, paleontologic and sedimentologic investigation. Comprised of limestones, shaly limestones and dolostones, the Helderberg Group crops out in central and eastern New York State (Fig. 2). Rickard (1962) recognized eight formations, which he regarded as diachronous lithofacies. This correlation made it possible for Laporte (1969) to give a detailed reconstruction of paleoenvironments along a transect from nearshore, peritidal environments (Manlius Formation), through subtidal shoals (Coeymans Formation) to shallow and deeper shelf settings

(Kalkberg and New Scotland formations). This interpretation literally became the textbook example of sedimentation in clear-water (i.e., carbonate) epeiric seas (e.g., Prothero, 1990).

Although Laporte (1969) did not address the sedimentology of the transition beds, their presence was a key factor in enabling Rickard to synthesize facies relationships, which then allowed Laporte to interpret paleoenvironments. So, what exactly are the transition beds?

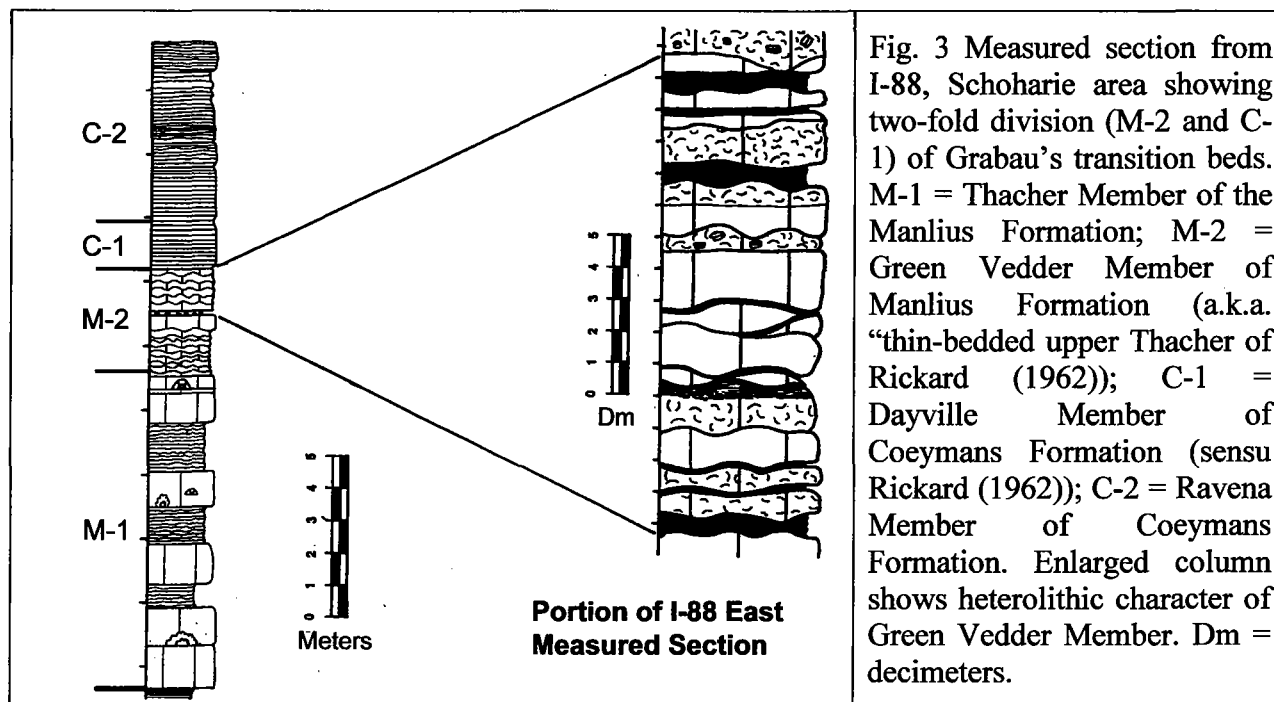
### GRABAU'S TRANSITION BEDS AND THE MANLIUS - COEYMANS CONTACT

Grabau (1906, p. 247) described the transition beds as thin bedded lime sandrocks and lime mudrocks, with shaly argillaceous beds. Grabau also reported lenses of lime mudrocks embedded in some of the shales. Faunally, the transition beds are dominated by the brachiopod *Stropheodonta varistriata* and echinoderm debris. Grabau (1906) measured the total thickness of the transition beds ranging from 3.66 meters to 4.01 meters (12 ft. to 13 ft. 2 in.) in the vicinity of Schoharie (Grabau 1906).

Reexamination of these beds in the vicinity of Schoharie, New York has provided several key insights. First, the transition beds are indeed a recognizable unit, distinctly different from the typical Manlius below and Coeymans above. Rather than an alternation of Manlius and Coeymans lithologies (sensu Grabau 1906), the transition beds actually comprise two units which differ from both the typical Manlius of eastern New York and from the Ravena Member of the Coeymans (Fig 3).

#### Subdivision of the transition beds

The lower portion of the transition beds comprise skeletal packstones, wackestones and mudstones (6-18 cm beds) with a macrofauna dominated by brachiopods, especially *Stropheodonta varistriata* and *Howellella vanuxemi*, pelmatozoans, such as *Lasiocrinus scoparius*, *Conularia* sp., ramose and fenestrate bryozoans, and gastropods. Packstone and wackestone beds display sharp bases and have planar-laminated to undulose tops. Most beds are normally graded, but extensive bioturbation obscures other internal structures. Infiltration fabrics are common in the coarser packstones. These sedimentologic features and the shallow-water aspect of the fauna suggest a tempestite origin. Muddy tempestites in the lower transition beds are interbedded with dark, carbonaceous shales, which have yielded an abundant, carbonized biota comprising scolecodonts, poorly preserved annelid soft tissues and *Medusaegraptus*, a non-calcified, aspondyl, dasycladacean alga (Matteson, Natel and Ebert, 1996). The lower transition beds are separated from unquestionable beds of the Thacher Member of the Manlius Formation by a sharp, non-depositional discontinuity. The lower transition beds are separated from the upper transition beds by an erosional disconformity (see below).



Limestones of the upper transition beds are crinoid-brachiopod grainstones and packstones, which commonly rest with sharp, erosive contacts on interbeds of finer-grained lithologies (variably argillaceous, dolomitic, ostracod - peloid grainstones and packstones to wackestones). Some bedding planes are littered with long, articulated stems of the crinoid *Ctenocrinus pachyactylus* (see Grabau 1906). Imbricate shells, crudely graded bedding and obrution assemblages imply a tempestite origin. Because the coarse packstones and grainstones are similar to the facies that comprise the Coeymans Formation, their presence reinforced the appearance of a gradual transition between the Manlius and Coeymans formations. However, the top of the transition interval is marked by a third, sharp erosional discontinuity.

From the description above, it seems clear that the Manlius - Coeymans contact in the Schoharie area is not a simple disconformity, yet it is not a gradational interfingering either. Rather, the "transition beds" comprise two distinct units that are not typical Manlius or Coeymans. Therefore, the questions are: 1) what are these units and 2) can they be correlated outside of the Schoharie area? Further, how significant are the discontinuities that underlie the lower transition beds and cap the lower and upper transition beds and can they be traced laterally?

#### **Clockville Discontinuity, Terrace Mountain Unconformity and Correlation of the Lower Transition Beds**

The surface that separates the lower transition beds from the more typical, thick-bedded Thacher Member of the Manlius Formation is a newly recognized discontinuity produced by sediment starvation and possible sediment bypass. For ease of discussion, we refer to this surface as the Clockville Discontinuity, a name derived from the excellent road cut at Clockville, NY.

The Clockville Discontinuity is sharp and locally erosional. At the type section, isolated pockets (up to 6 cm deep and 12 cm wide) are scoured into the top of the typical Thacher and filled with coarse skeletal grainstone, a lithology which is atypical for this part of the section and different from those that comprise the typical Thacher below and the lower transition beds above. Skeletal debris includes resistant echinoderm ossicles, fragments of thick-shelled brachiopods and rare branching bryozoans. The Clockville Discontinuity is abruptly overlain by beds referable to the lower transition beds, which appear to represent a significantly deeper environment of deposition than the typical Thacher below. Therefore, it seems likely that the Clockville Discontinuity is a flooding or transgressive surface.

Goodwin and Anderson (1988) recognized a cryptic unconformity at the top of the Thacher Member of the Manlius Formation in central and eastern New York. For ease of discussion, we have christened this surface the Terrace Mountain Unconformity (Ebert and Matteson 2001a; 2001b; 2003), after the excellent exposure on I-88 on the flank of Terrace Mountain near Schoharie. From Cherry Valley westward, the Dayville Member of the Coeymans Formation and its western equivalent, the Olney Member of the Manlius Formation overlie this unconformity (Goodwin and Anderson (1988)). We have correlated the Terrace Mountain Unconformity from the Syracuse area through Cherry Valley to Schoharie, where we recognize it as the surface that marks the top of the lower transition beds. The Terrace Mountain Unconformity has a gently angular nature on a regional scale (see Goodwin and Anderson 1988). At individual outcrops, the contact displays minor erosional relief, ranging from a few millimeters (Fig. 4) up to nearly 10 centimeters. Intraclasts of Thacher-type lithology occur rarely in the bed immediately overlying the Terrace Mountain Unconformity.

If our correlation and that of Goodwin and Anderson (1988) is correct, then the beds between the Clockville Discontinuity and the Terrace Mountain Unconformity should have some affinity with the Thacher Member of the Manlius Formation by their position in the sequence. If this is the case, then why have they been recognized separately as part of the transition beds? The answer lies in their distinctive lithology, paleocommunity and taphonomy, which differ from usual characterization of the Manlius as comprising thick-bedded peritidal lithologies (Laporte 1969).

The lithologies of the lower transition beds comprise an association of poorly skeletal limestones in decimeter beds alternating, in part, with dark, organic-rich, calcareous shales. Moreover, the fauna and flora of these interbeds and their distinctive taphonomy are unique in this part of the section. Matteson, Natel and Ebert (1996) originally described this paleocommunity and its preservation from the exposure along I-88 on Terrace Mountain and correlated it some 140 km westward as far as Chittenango Falls. The interval thickens westward from Schoharie and the beds bearing the carbonized biota clearly occur within the Athin bedded upper Thacher= that Rickard (1962) recognized between Manlius, near Syracuse and Oriskany Falls in central New York. Limestones within this portion of the Thacher are also dominated by *Stropheodonta varistriata* and *Howellella vanuxemi* (Rickard, 1962, p. 54) and include *Lasiocrinus scoparius*. The carbonized flora and fauna in the shaly interbeds is distinctive and consistent across the outcrop belt. Thus, the lower portion of the Grabau=s transition beds is a

traceable unit within the Thacher Member of the Manlius Formation. We refer to this unit as the Green Vedder Member (informal) of the Manlius Formation, after the excellent exposure along Green Vedder Road, outside the large active quarry at Oriskany Falls. Although Rickard did not trace this unit eastward from Oriskany Falls, some of Rickard's descriptions (e.g., R-117, Cullen) indicate that he may have recognized the unit farther to the east. In fact, the Green Vedder Member, with its signature biota and taphonomy, is present at Cherry Valley, Sharon Springs, and Schoharie. East of Schoharie, the lower transition beds/Green Vedder Member are removed by the Terrace Mountain Unconformity. However, a small outlier of the Green Vedder Member has been identified at Ravena.

### **The Howe Cave Unconformity and Correlation of the Upper Transition Beds**

By virtue of their position above the Terrace Mountain Unconformity, it is reasonable that the upper transition beds should have some relationship to the Olney/Dayville interval which overlies the unconformity in central New York. However, owing to some general similarities in lithology, it is also possible that the upper transition beds are related to the Deansboro or Ravena members of the Coeymans Formation. This correlation is less likely because in the Schoharie Valley, the upper transition beds are separated from the superjacent Ravena Member of the Coeymans Formation by an abrupt, erosional discontinuity. We have dubbed this surface the Howe Cave Unconformity, a name derived from the inactive Howe Cave Quarry, where the surface was first observed truncating the upper transition beds. At the type locality and all locations to the east, the Howe Cave Unconformity is overlain by the Ravena Member of the Coeymans Formation. Westward from the type locality, correlation becomes more complex as the Howe Cave Unconformity rises stratigraphically, the upper transition beds thicken below it and additional, higher units are preserved.

From its type section, the Howe Cave Unconformity descends eastward, thinning the upper transition beds to disappearance somewhere between Gallupville and John Boyd Thacher Park. As a result of this descent, the Howe Cave Unconformity merges with the previously described Terrace Mountain Unconformity and, at many localities, with the Clockville Discontinuity. Thus, the contact between the Manlius and Coeymans formations from Thacher Park eastward and south through the Hudson Valley is a composite of the Clockville Discontinuity, the Terrace Mountain Unconformity and the Howe Cave Unconformity. It was this composite unconformity that Chadwick (1944) described from the Hudson Valley. It was also from this composite unconformity that Goodwin and Anderson (1988) reported the loss of some 4.5 meters of section. Because the merging of these two unconformities was probably accomplished by truncation of the Clockville Discontinuity and the Terrace Mountain Unconformity beneath the Howe Cave Unconformity, we continue the use of the designation Howe Cave Unconformity for this surface throughout eastern New York, regardless of the presence or absence of the upper transition beds below the unconformity (Ebert and Matteson 2001a; 2001b; 2003).

Lithologically, the upper transition beds bear a striking resemblance to the Dayville Member of the Coeymans Formation. Indeed, they show remarkable similarities in fauna and taphonomy as well, particularly in the presence and preservation of articulated stems of the



crinoid *Ctenocrinus pachydactylus*, which occurs at the Dayville type section and in the upper transition beds at Schoharie. These similarities strongly suggest that the upper transition beds are a previously unrecognized eastward extension of the Dayville Member. If this is the case, then the Dayville should exist between Schoharie and Cherry Valley, where Rickard acknowledged the presence of the Dayville Member.

From the Gallupville west to Sharon, New York, the upper transition beds vary between one or two meters in thickness down to a few decimeters locally owing to relief on the Howe Cave Unconformity. However, the Dayville/upper transition zone is nearly three meters thick at Sharon Springs and expands rapidly to approximately six meters at Cherry Valley, where Rickard indicates that it comprises the lower half of the Coeymans Formation. West of Cherry Valley, this interval is referred to as the Dayville Member of the Coeymans Formation (Rickard, 1962). Tracing of recognizable marker beds from the type section of the Dayville to Cherry Valley confirms assignment of the lower Coeymans at Cherry Valley to the Dayville Member. From these correlations, it is now apparent that the upper portion of Grabau's transition beds in the Schoharie area is an erosionally thinned, eastward continuation of the Dayville Member, which Rickard (1962) had previously restricted to the area west of Cherry Valley. It is also clear that from Cherry Valley to Gallupville, the Howe Cave Unconformity occurs within the Coeymans Formation (between the Dayville and Ravena members) as defined by Rickard (1962).

At Cherry Valley intraclasts of Manlius aspect and favocitid corals, which encrust a sharp surface, mark the position of the Howe Cave Unconformity between the Dayville Member and the Deansboro/Ravena Member, both parts of the Coeymans Formation. West of Cherry Valley, the Howe Cave Unconformity rises stratigraphically and truncates units that have been regarded as parts of the Manlius Formation (Elmwood, Clark Reservation and Jamesville members). Intraclasts or lithoclasts of the Elmwood or Clark Reservation (?) have been observed in a discontinuous outcrop at Salt Springville, near Cherry Valley. Rickard and Zenger (1964) also reported such clasts in this area. Exposures where the Howe Cave Unconformity can be directly observed cutting the Elmwood and then the Clark Reservation members have not been found. We attribute this to the general paucity of outcrops in the area between Salt Springville and Jordanville and to the extremely limited thickness of these units (approximately 3 meters maximum in this area). From Jordanville westward, the Howe Cave Unconformity overlies the Jamesville Member of the Manlius Formation. Indeed, it is this unconformity that was reported by Smith (1929) and Logie (1933). In some exposures, such as the active quarry at Oriskany Falls, the Howe Cave Unconformity exhibits well-developed paleokarstic features (Fig. 4). Steep-sided solutional pits penetrate up to five centimeters into the top of the Jamesville and rounded lithoclasts of pebble to small cobble size litter the unconformity. At some locations, such as Jordanville, *in situ* colonies of stromatoporoids are truncated (Fig. 4). In eastern New York (e.g., Catskill vicinity), the Howe Cave Unconformity is sharp, undulating and, displays *Trypanites* (?) borings (Fig. 4; see also Laporte 1969).

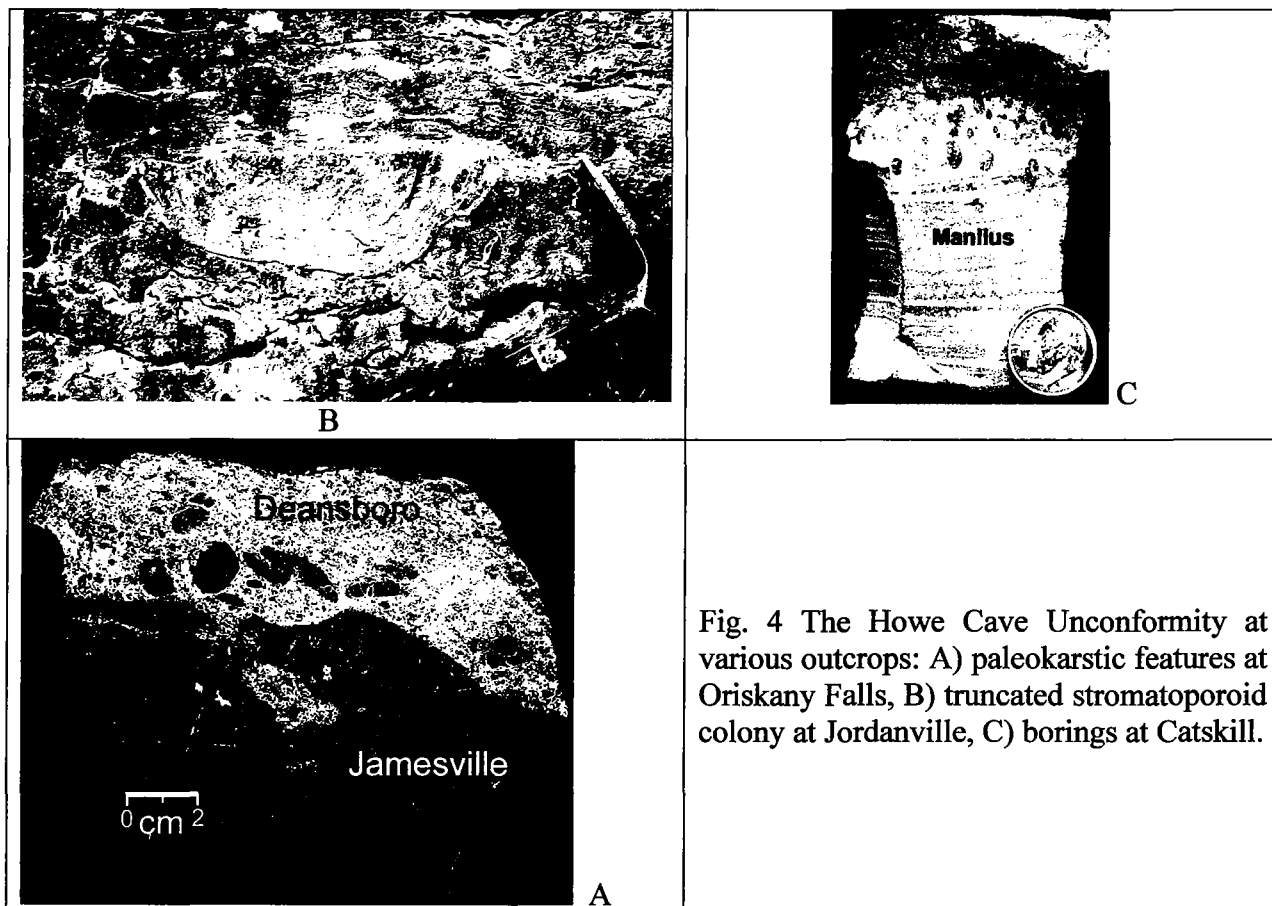


Fig. 4 The Howe Cave Unconformity at various outcrops: A) paleokarstic features at Oriskany Falls, B) truncated stromatoporoid colony at Jordanville, C) borings at Catskill.

The regionally angular nature of the Howe Cave Unconformity is much more pronounced than that which is displayed by the older Terrace Mountain Unconformity (Fig. 5). The Howe Cave Unconformity has greater stratigraphic relief and represents a much greater erosional vacuity. In central New York, the Howe Cave Unconformity overlies the Jamesville Member of the Manlius Formation. The subjacent Clark Reservation and Elmwood members of the Manlius are removed in a relatively narrow region around Salt Springville. From Cherry Valley to Gallupville, the unconformity truncates the Dayville Member. East of Gallupville, the Howe Cave Unconformity progressively bevels the Thacher Member of the Manlius Formation. Therefore, the stratigraphic relief on the Howe Cave Unconformity is at least 24 meters across 150 kilometers of outcrop.

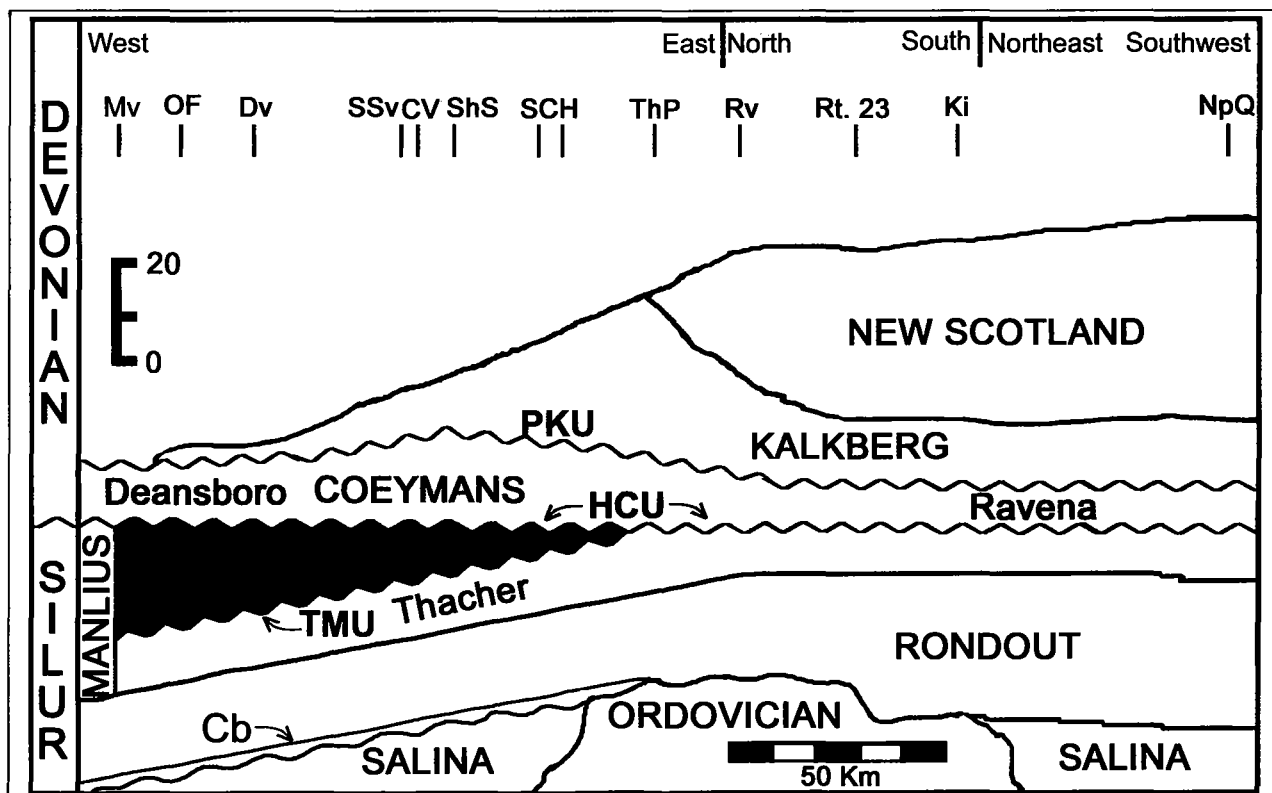


Fig. 5 Revised stratigraphy of the lower Helderberg Group, showing the Terrace Mountain and Howe Cave unconformities (after Ebert and Matteson 2003).

### Stratigraphic Significance of the Terrace Mountain and Howe Cave Unconformities

The combined presence of the Terrace Mountain Unconformity in central New York and the eastern portions of the Howe Cave Unconformity precludes any portion of the Thacher Member of the Manlius Formation from having been coeval with other members of the Manlius Formation or any part of the Coeymans Formation (Fig. 5). Although Rickard (1962) acknowledged this possibility, he did not give it much credence. More significantly, the western portions of the Howe Cave Unconformity contradict the previously assumed lateral equivalence of the upper members of the Manlius Formation with portions of the Coeymans, Kalkberg and New Scotland formations (e.g., Rickard 1962; Laporte 1969). Owing to the temporal separation necessitated by the Howe Cave Unconformity, the widely cited paleoenvironmental spectrum interpreted by Laporte (1969) could not have existed. Goodwin and Anderson (1988) began the process of dismantling Rickard's stratigraphy and Laporte's paleoenvironmental reconstruction with their documentation of erosional loss beneath the surface that we term the Terrace Mountain Unconformity. Recognition of the Howe Cave Unconformity completes the disassembly of the Helderberg epeiric sea model (Ebert and Matteson 2001a; 2001b; 2003) and invalidates correlations of the small scale allocycles (PACs) of Goodwin and Anderson (1988) in central and eastern New York.

Recognition of the Terrace Mountain and Howe Cave unconformities carries additional implications for locating the Silurian - Devonian boundary in New York State. The existence of these unconformities changes the relative age relationships between the Coeymans Formation and the underlying Manlius and Rondout formations. Since these units can not be laterally equivalent to the acknowledged Lower Devonian Coeymans Formation, the question arises as to their age and the position of the Silurian - Devonian boundary. Previous projections of the boundary from eastern New York into portions of the Manlius and Rondout formations in the central and western areas are no longer tenable because they cross the Howe Cave and Terrace Mountain unconformities. The age of the Rondout and Manlius Formations and the position of the Silurian - Devonian boundary are considered below.

### **Silurian-Devonian Boundary and the Age of Helderberg Units**

The nature of the Manlius-Coeymans contact is important in establishing the position of the Silurian-Devonian boundary in the Appalachian Standard Succession of New York. In the nineteenth century, the entire Helderberg Group was regarded as Silurian. Clark (1889, 1900) and Schuchert (1900) transferred most formations of the Helderberg Group to the Devonian. However, the Manlius and older units were retained in the Silurian. These designations were widely used until Rickard (1962) reinterpreted the lithostratigraphy.

Rickard's (1962) interpretation of the Manlius-Coeymans contact as a facies change provided the rationale for reassigning the Manlius and portions of the underlying Rondout Formation to the Lower Devonian. The Ravena and Deansboro members of the Coeymans Formation bear the terebratulids *Cyrtina*, *Podolella* and *Nanothyris*, and the index conodont *Icriodus woschmidti* and are therefore unquestionably Lower Devonian (Barnett 1971; Rickard 1975). Since parts of the Manlius and Rondout formations were viewed as laterally equivalent to and therefore coeval with the Coeymans Formation, these units were also regarded as Lower Devonian (Rickard, 1962; 1975) despite the lack of any diagnostic paleontologic criteria. In Rickard's reconstruction, the oldest part of the Coeymans Formation (Ravena Member) is in eastern New York. Therefore, the first occurrence of *I. woschmidti* at the base of the Coeymans in the Hudson Valley could be projected into the Manlius and Rondout formations in central New York. The absence of *I. woschmidti* from these formations could be explained as a consequence of facies preference and exclusion from unfavorable environments (e.g., Barnett 1977; Johannessen, et. al., 1997).

Lacking *I. woschmidti* in the Manlius and Rondout formations, Barnett (1971; 1972; 1977) utilized biometric trends in the more common and abundant elements of *Ozarkodina remscheidensis eosteinhornensis* and *O. r. remscheidensis* to locate the Silurian - Devonian boundary. The result was an inferred boundary that fell within the Manlius and Rondout formations, well below the first occurrence of *I. woschmidti* in almost all areas of the state. Because Barnett failed to distinguish between the subspecies *eosteinhornensis* and *remscheidensis*, Klapper (1981) questioned this placement of the boundary.

Milunich and Ebert (1991) discussed a sparse and fragmental conodont fauna from the Rondout and Manlius formations near Schoharie, but were unable to locate the boundary.

Johannessen, Natel, and Ebert (1997) indicated that the boundary might be considerably higher than the Rondout position projected by Barnett (e.g., 1977), for example, as high as 4 m above the base of the Coeymans at Cherry Valley and near the top of the transition beds in the Schoharie Valley, positions that are near the Howe Cave Unconformity.

Recognition of the Howe Cave Unconformity dictates that no part of the Manlius was coeval with the Coeymans. Indeed, the Coeymans Formation (Ravena and Deansboro members) is entirely younger than all members of the Manlius Formation. To test this reconstruction and to determine the age of the Manlius Formation, units above and below the unconformity were sampled for conodonts. Additionally, Rickard's conodont collection, on loan from the New York State Museum was also studied.

Our samples and the Rickard collection clearly demonstrate that the first occurrence of *Icriodus woschmidti* is in the Ravena or Deansboro members of the Coeymans Formation, above the Howe Cave Unconformity. This is consistent with the findings of Barnett (1971) and Johannessen, Natel, and Ebert (1997) who attributed this to facies selectivity. However, sub-unconformity units (members of the Manlius and the Dayville Member), do not bear a more nearshore, coeval fauna.

If our correlations are correct, then the stratigraphically highest unit below the Howe Cave Unconformity is the Jamesville Member of the Manlius Formation. The Rickard collection and our samples from the Jamesville have yielded the following conodonts: *Ozarkodina confluens*, *O. excavata*, *O. remscheidensis remscheidensis* and *O. r. eosteinhornensis*. These taxa range from the Ludlovian through the Pridoli stages. The presence of *O. remscheidensis eosteinhornensis* indicates a latest Silurian or Pridolian age for the Jamesville. Thus, the conodont biostratigraphy corroborates the existence of the Howe Cave Unconformity and our interpretation of the temporal separation of the Manlius and Coeymans formations.

At Cherry Valley, portions of the transition beds/Dayville Member have yielded *Ozarkodina r. remscheidensis*, *O. r. eosteinhornensis* and *O. confluens*. The occurrence of *O. confluens* in these strata indicates a mid-Pridoli age (Johannessen, et. al., 1997). Therefore, nearly half of the Pridoli may be missing beneath the Howe Cave Unconformity in this area.

The presence of the Howe Cave Unconformity provides a satisfactory explanation for the stratigraphic distribution of *I. woschmidti*. The first occurrence of this proxy for the Silurian - Devonian boundary at the base of the Coeymans Formation is not the result of the appearance of a favorable environment as indicated by Barnett (1971). Rather, it is attributable to the resumption of deposition following the Howe Cave Unconformity. The Coeymans Formation (minus the Dayville Member) is the oldest Lochkovian unit in New York. Therefore, the Silurian B Devonian boundary occurs within erosional vacuity of Howe Cave Unconformity (Fig. 6), a placement which is consistent with the views of earlier workers (e.g., Logie, 1933; Goldring 1935, 1943; Schuchert 1943 and Chadwick 1944) who viewed the Manlius-Coeymans contact as a systemic unconformity (Rickard 1962, p. 49).

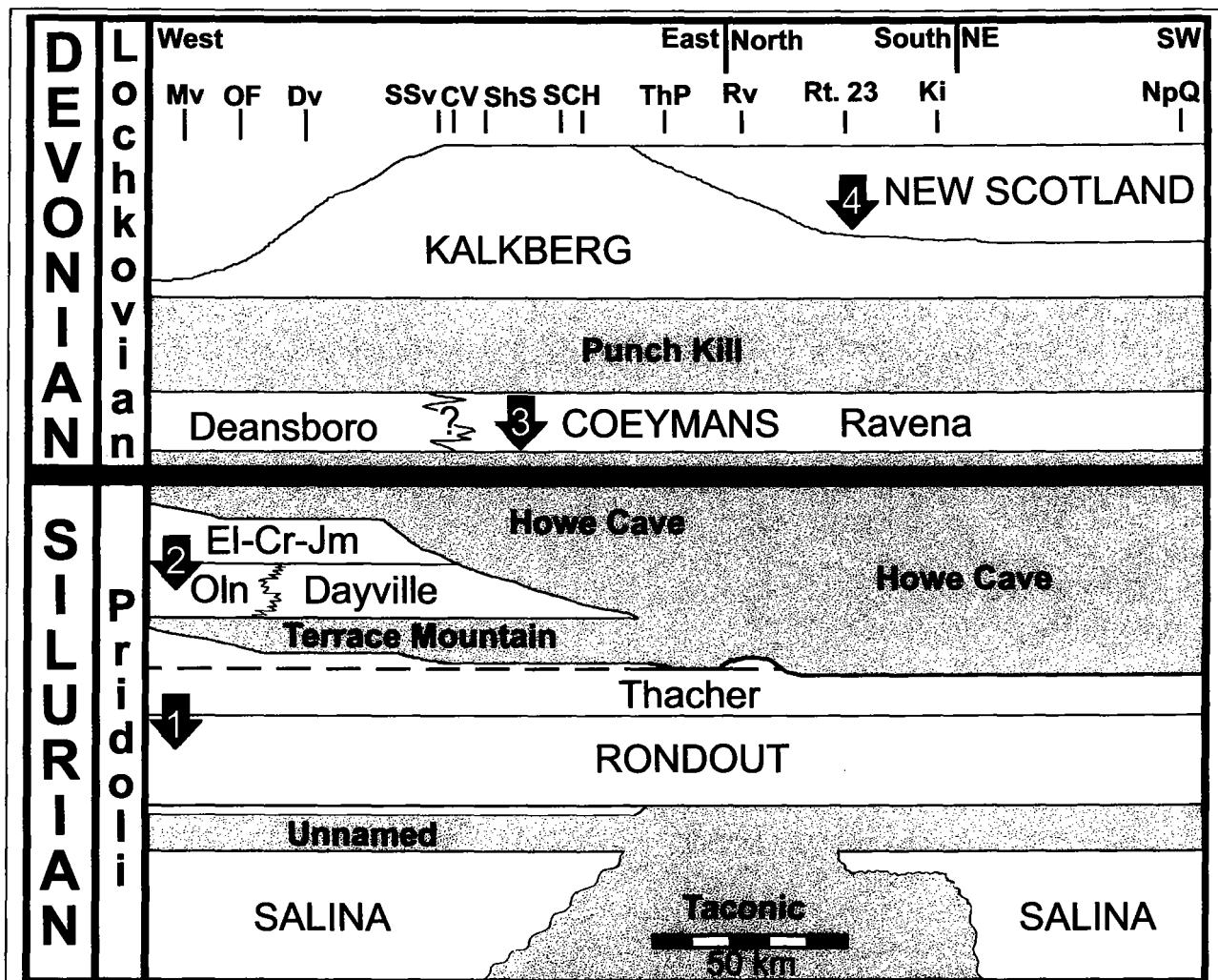


Fig. 6 Time-stratigraphic cross section of the lower Helderberg Group showing erosional vacuities of the Terrace Mountain and Howe Cave unconformities (after Ebert and Matteson 2003).

### Reassignment of Dayville Member

The Dayville Member was defined as a unit within the Coeymans Formation by Rickard (1962), owing to general lithologic similarity. Faunal similarities are less pronounced, but Rickard felt that overall this unit was much more closely allied with the Coeymans despite its lateral equivalence with Olney Member of the Manlius Formation (Rickard, 1962, p. 72).

Throughout its extent, the Dayville rests with unconformity (Terrace Mountain Unconformity) on the Thacher Member of the Manlius Formation. It is overlain by higher members of the Manlius (Elmwood/Clark Reservation) from its western transition with the Olney to Salt Springville, near its previously presumed eastern limit. The Ravena Member of the Coeymans Formation overlies the Dayville from Cherry Valley eastward only where these units

are brought into juxtaposition by the Howe Cave Unconformity. These relationships suggest that the Dayville is much more closely allied spatially and temporally with the Manlius than with the more lithologically similar Coeymans Formation. For these reasons, we suggest that the Dayville Member should be reassigned to the Manlius Formation.

## SUMMARY AND CONCLUSIONS

Grabau's transition beds have proven to be key elements in a new understanding of stratigraphic relationships within the Helderberg Group (Fig. 5). The transition beds are underlain by the Clockville Discontinuity and comprise two units; each capped by a regionally extensive unconformity. The lower unit is an eastward extension of the Green Vedder Member (thin-bedded upper Thacher of Rickard (1962)) of the Manlius Formation, which is thinned below the Terrace Mountain Unconformity. The upper transition unit is an eastward extension of the Dayville Member, which should be reassigned from the Coeymans to the Manlius Formation. The Dayville/upper transition beds are truncated by the Howe Cave Unconformity such that they thicken westward from the Schoharie Valley and are overlain by higher units which are progressively preserved beneath the rising Howe Cave Unconformity. Conodonts indicate that this unconformity marks the boundary between the Silurian and Devonian periods, exactly as envisioned by Grabau in 1906 and earlier workers in the nineteenth century.

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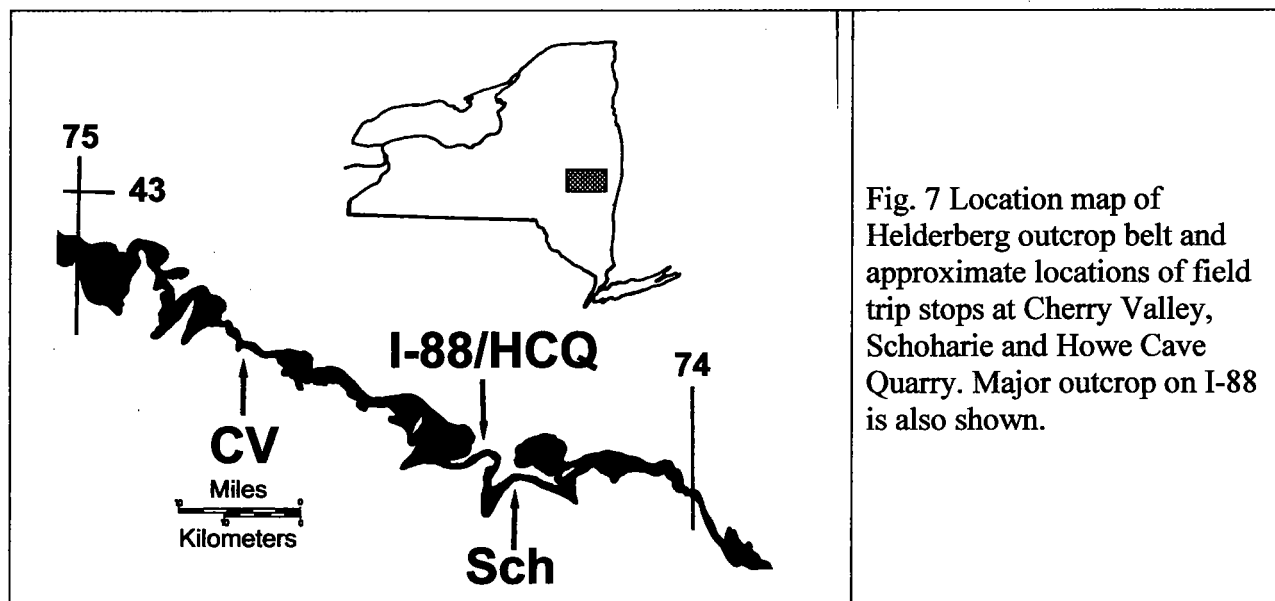


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## FIELD TRIP ROAD LOG AND DESCRIPTION OF STOPS



| Cumulative Mileage | Mileage between points | Description  |
|--------------------|------------------------|--|
| 0.0                | 0.0                    | Road log begins at the intersection of Ravine Parkway and West Street – the main entrance to the SUNY Oneonta campus   |
| 0.4                | 0.4                    | Turn left (east) at “graffiti wall” onto Center Street   |
| 0.8                | 0.4                    | Traffic light. Turn right (south) onto Maple Street  |
| 1.0                | 0.2                    | Intersection of Maple Street with Main Street. Continue straight (south) on James Lettis Highway   |
| 1.6                | 0.6                    | After underpass, traffic light for I-88 eastbound onramp. Turn left to get on highway.   |
| 6.2                | 4.6                    | Exit 17 for Rts. 7 and 28 to Colliersville and Cooperstown. Exit from highway.   |
| 6.55               | 0.35                   | End of off ramp. Turn left (north) on Rt. 28   |
| 6.9                | 0.35                   | Schenevus Creek. Note the well developed cut banks on large meanders.  |
| 8.0                | 1.1                    | Intersection with connector to Rt. 7. Continue north on Rt. 28   |
| 8.3                | 0.3                    | Outcrop of Gilboa Formation with well-developed ball and pillow structures (see Hutchinson, 1977). Opposite is NYSEG power station and dam on Susquehanna River to form Goodyear Lake. |
| 16.3               | 7.7                    | Traffic light in village of Milford. Intersection with Rt. 166. Turn right (north) onto Rt. 166.   |
| 35.4               | 19.1                   | Traffic light in village of Cherry Valley. Turn right and continue   |

|      |     |   |
|------|-----|---|
|      |     | north on Rt. 166.   |
| 36.8 | 1.4 | Outcrops of Onondaga Formation.   |
| 37.4 | 0.6 | Discontinuous outcrops of Tristates Group   |
| 37.5 | 0.1 | Outcrop of Kalkberg Formation   |
| 37.6 | 0.1 | Underpass for U.S. Rt. 20. Rt. 166 has ended and road is now Sprout Brook Road (Otsego County Rt. 32).  |
| 37.9 | 0.3 | Outcrop of Coeymans Formation in contact with Kalkberg Formation. Contact is marked by a silt-rich bed and the Punch Kill Unconformity of Ebert and Matteson (2003). See also description below for <b>STOP 1</b> . |
| 38.1 | 0.2 | North end of large Helderberg outcrop. This is <b>STOP 1</b> .  |

### **STOP 1: Cherry Valley – Sprout Brook Road, Otsego County Rt. 32**

This road cut is the stratigraphically lowest of a series on Sprout Brook Road (Cty. Rt. 32), Rt. 166 and, U.S. Rt. 20 that expose the entire Lower Devonian section that is present in New York State (See Brett and Ver Straeten, 1997). This outcrop corresponds approximately to Rickard's (1962) section number 94, which was measured in nearby Judd's Falls.

The section begins at the north end of the outcrop with approximately 1 meter of the Rondout Formation. The Rondout is abruptly overlain by the Thacher Member of the Manlius Formation. Near the top of the Manlius Formation, the style of bedding becomes thinner (decimeter scale) and intercalations of dark gray to black shale appear. These beds (total thickness = 0.82 m) constitute an erosionally thinned portion of the Green Vedder Member ("thin-bedded upper Thacher" of Rickard (1962)). These dark shale interbeds have yielded the carbonized biota discussed above and in Matteson, Natel and Ebert (1996).

The contact with the overlying Coeymans Formation has traditionally been placed at the appearance of the first, coarse crinoidal bed. This bed overlies the Terrace Mountain Unconformity (13.27 m from base of section). The lower Coeymans (6.03 m thick) comprises interbedded coarse and fine beds which Rickard equated with the Dayville Member. At the top of the Dayville Member, there is a change in the style of bedding to thicker and more irregular beds. The contact is marked by an erosional surface with several centimeters of relief. Sporadic, *in situ* favositid corals encrust the contact. Rare, fine grained intraclasts indicate that this surface marks the position where the Elmwood and Clark Reservation members of the Manlius (which overlie the Dayville to the west) have been removed. This is the Howe Cave Unconformity (19.3m from base of section).

Overlying the Howe Cave Unconformity, thick, irregular to vaguely nodular beds of the upper Coeymans here have been referred to both the Deansboro and the Ravena members of the Coeymans Formation (see Rickard, 1962). This unit continues to the top of the outcrop (26.6 m). Using distinct horizons of chert and beds bearing holdfasts of the cystoid *Lepocrinites gebhardi*, we have correlated beds from the top of this outcrop with the base of the small outcrop 0.2 miles to the south on Sprout Brook Road. At the latter outcrop, 4.7 m of the Ravena Member is exposed. Thus, the maximum thickness of the Coeymans (Dayville and Ravena combined) is

slightly over 19 meters. Previous accounts of the stratigraphy in this area (e.g., Rickard, 1962, 1981; Gurney and Friedman, 1986; Liebe and Grasso, 1990) reported a covered interval between the main outcrop and this smaller exposure and a total thickness of 30 meters for the Coeymans Formation (Rickard, 1981). However, it appears that these authors did not take into account the regional dip, which is approximately two degrees in this area. When relative elevations, the regional dip and the distance between these exposures are utilized, the top of the large outcrop and the base of the smaller exposure coincide. Thus, it appears that previously published measured sections for Cherry Valley have exaggerated the thickness of the Coeymans Formation by as much as 50%. Although we will not visit the smaller exposure on this trip, the Punch Kill Unconformity (Ebert and Matteson, 2003) is well exposed here, where it separates the Coeymans Formation from approximately 2 m of the Kalkberg Formation.

Taken together, the Green Vedder beds and the Dayville Member appear to constitute a gradational transition between the Manlius (proper) and Coeymans formations (Rickard, 1962, 1975, 1981). However, such a gradation cannot exist owing to the presence of the regionally traceable Terrace Mountain and Howe Cave unconformities which cap each unit respectively. Furthermore, the upper Thacher (Green Vedder Member) and Dayville Member exposed on Sprout Brook Road correlate to the west with units bearing the same names. They are distinct units and not an ambiguous zone of transition, assignable to neither formation. When correlated westward, they are separated by other members of the Manlius Formation. When traced eastward, these units thin and become the zone that Grabau (1906) designated as transition beds in the Schoharie area. Grabau's (1906) transition beds (*sensu strictu*) will be seen at STOP 2.

| <b>Cumulative Mileage</b> | <b>Mileage between points</b> | <b>Description</b>  |
|---------------------------|-------------------------------|---|
| 38.6                      | 0.5                           | After visiting <b>STOP 1</b> , continue north on Sprout Brook Road for approximately 0.5 miles (0.3 miles from north end of outcrop). At this point, the shoulder is wide on both sides of the road. Execute a U-turn and return south on Spout Brook Road. |
| 39.4                      | 0.8                           | Cross under Rt. 20  |
| 39.5                      | 0.1                           | Turn right onto ramp to enter U.S. Rt. 20 eastbound.  |
| 39.7                      | 0.2                           | Kalkberg outcrops   |
| 40.0                      | 0.3                           | Railroad overpass. Kalkberg outcrops with well-developed K-bentonites (see Smith, Berkheiser and Way, 1986; Ebert, Applebaum and Finlayson, 1992; Tucker, et.al., 1998). Kalkberg is disconformably overlain by the Oriskany Sandstone.                     |
| 40.1                      | 0.1                           | Outcrops of Tristates Group which also bears K-bentonites (see Ver Straeten and Brett, 1997).   |
| 40.4                      | 0.3                           | Begin outcrop of Onondaga Formation. Edgecliff, Nedrow and Moorehouse Members are well exposed (see also Brett and Ver Straeten, 1997).   |
| 41.5                      | 1.1                           | Small outcrop of Seneca Member of the Onondaga Formation. The Tioga-B (Onondaga Indian Nation) K-bentonite is exposed at the base of the outcrop. This outcrop has provided conodonts which are   |

|      |      |   |
|------|------|---|
|      |      | the subject of direct radiometric dating studies (Research is in progress, but see Elrick, et. al. (2002) for a preliminary report.)  |
| 42.0 | 0.5  | Outcrop of the Marcellus Formation, including Union Springs, Cherry Valley Limestone and Chittenango members, exposed on Chestnut Street, sub parallel to Rt. 20 (see Griffing and Ver Straeten, 1991; also Brett and Ver Straeten 1997). |
| 44.5 | 2.5  | Outcrops of the Kalkberg Formation at the edge of Leesville. Exposed section is similar to the outcrop at Cherry Valley (40.0 in this road log).  |
| 45.9 | 1.4  | Outcrops of Kalkberg at edge of Sharon Springs. Section is similar to Leesville and Cherry Valley.  |
| 46.2 | 0.3  | Traffic light at intersection with Rt. 10. Continue east on U.S. 20.  |
| 49.4 | 3.2  | Outcrop of Onondaga Formation.  |
| 62.1 | 12.7 | Turn right (south) onto Rt. 30A at traffic light in Sloansville.  |
| 63.6 | 1.5  | Outcrop of Ordovician flysch of the Schenectady/Frankfort Fm.   |
| 66.0 | 2.4  | Junction with Rt. 7. Turn left to continue south on Rt. 30A   |
| 67.2 | 1.2  | Turn right at blinking light to continue south on Rt. 30A. Cross over I-88. Note view of Terrace Mountain to the right.   |
| 67.8 | 0.6  | Dunkin' Donuts – a convenient rest stop in this area.   |
| 68.3 | 0.5  | Junction with Rt. 30. End of 30A. Continue straight on Rt. 30 south.  |
| 69.8 | 1.5  | Junction with Rt. 443. Continue on Rt. 30.  |
| 71.6 | 1.8  | Turn right onto Bridge Street.  |
| 72.2 | 0.6  | Cross bridge  |
| 72.5 | 0.3  | Turn right onto Terrace Mountain Road   |
| 72.6 | 0.1  | Pull over on right – note that shoulder is not overly wide! Outcrop on left is <b>STOP 2</b> .  |

### **STOP 2: Terrace Mountain Road, Schoharie**

In the vicinity of Schoharie, there are numerous outcrops of Grabau's (1906) transition beds. This exposure on Terrace Mountain Road probably postdates Grabau's investigations, however it is one of the most accessible outcrops for examining these beds. Therefore, we selected it for this trip, rather than one of Grabau's "classic" outcrops, many of which still exist.

In this exposure, the twofold character of Grabau's transition beds is readily apparent. The lowest beds in the outcrop comprise a thinned eastward extension of the Green Vedder Member (Rickard's "thin-bedded upper Thacher). The Terrace Mountain Unconformity is marked by the abrupt appearance of coarse, crinoidal grainstones and packstones of the Dayville Member of the Coeymans Formation. Note that this is an extension of the term Dayville. In Rickard's (1962) stratigraphy, the Dayville did not extend eastward beyond Cherry Valley. Several of the Dayville beds display well-developed structures associated with tempestites: vertical grading, undulating to hummocky cross stratification and rare symmetrical ripple caps.

The Dayville has been substantially thinned from 6 m at Cherry Valley to approximately 1 m in the Schoharie area. This thinning is a result of the eastward descent of the Howe Cave

Unconformity, which separates the Dayville from the overlying Ravena Member of the Coeymans Formation. Minor local relief on the Howe Cave Unconformity is present in the Schoharie region, where the thickness of the Dayville remnant ranges from 0.7 to nearly 1.25 m at various outcrops.

Overlying the Howe Cave Unconformity, thick beds of the Ravena Member of the Coeymans Formation comprise the majority of this outcrop and they extend to the top of the exposed section. Upper portions of the Ravena display abundant club-like holdfasts of the cystoid *Lepocrinites gebhardi*. We are currently investigating the lateral extent of this subdivision of the Ravena Member (See also Matteson and Ebert, 2001).

| Cumulative Mileage | Mileage between points | Description   |
|--------------------|------------------------|---|
| 72.7               | 0.1                    | Continue on Terrace Mountain Road to driveway for turn around. Return downhill on Terrace Mountain Road.  |
| 73.1               | 0.4                    | Turn left onto Bridge Street  |
| 73.3               | 0.2                    | Cross bridge  |
| 73.9               | 0.6                    | Turn left onto South Main Street (Rt. 30)   |
| 75.6               | 1.7                    | Junction with Rt. 443   |
| 77.0               | 1.4                    | Junction with Rt. 30A. Continue straight on Rt. 30A as Rt. 30 splits away to the right.   |
| 78.0               | 1.0                    | Cross over I-88   |
| 78.3               | 0.3                    | Blinking light at intersection with Rt. 7. Turn left to continue on Rt. 30A north. Rt. 7 west runs concurrently with Rt. 30A.                               |
| 79.4               | 1.1                    | Intersection where Rt. 30A and Rt. 7 split. Continue west on Rt. 7.   |
| 80.5               | 1.1                    | To the left and uphill, a large outcrop of the Helderberg Group is visible on I-88. This outcrop exposes the Kalkberg, New Scotland and Becraft formations. |
| 82.0               | 0.5                    | Traffic light. Turn right onto Howe's Cave Road (Schoharie County Rt. 8)  |
| 82.5               | 0.5                    | Cross railroad tracks.  |
| 82.6               | 0.1                    | Turn left (opposite Enders Avenue) onto Industrial Drive (private). This road forks almost immediately, keep to the right.                                  |
| 82.8               | 0.2                    | The original Lester Howe Hotel and future Cave House Museum. Also, original entrance to Howe Caverns.   |
| 82.9               | 0.1                    | Enter main Howe Cave Quarry – <b>STOP 3</b> .   |

### STOP 3: Howe Cave Quarry

The inactive Howe Cave Quarry is the type section of the Howe Cave Unconformity. Clear truncation of the upper transition beds (eastern extension of the Dayville Member) in this quarry enabled recognition of this unconformity *within* the Coeymans Formation as presently



defined. The truncation of beds in the Dayville is best observed in the weathered joint surface on the high wall at the entrance to the quarry, just beyond the old Lester Howe Hotel. In this face, the Green Vedder Member of the Manlius Formation (a.k.a. "thin-bedded upper Thacher) is also readily apparent. The upper part of the high wall is comprised of the massive Ravena Member, which, in parts, bears abundant holdfasts of *Lepocrinites gebhardi*. Elsewhere in the quarry, the Punch Kill Unconformity is exposed, overlain by approximately one meter of the Kalkberg Formation. This is visible at the top of the high wall in the main part of the quarry, but is only accessible for direct observation in a few places.

Other features of interest in the quarry include a thrust fault (Marshak and Bosworth, 1991; Mylroie and Palmer, 1977), extensive fields of ripples on the floor of the quarry (Ebert, et.al., 2000), numerous glacial striations on the surface above the quarry and, various karst and cave features associated with Howe Caverns and Barytes Cave (Mylroie and Palmer, 1977 and this volume), including the original entrance to Howe Caverns, adjacent to the Lester Howe Hotel.

The Howe Cave Quarry and Lester Howe Hotel will be centerpieces of the future Cave House Museum of Mining and Geology, a scientific and industrial educational facility, which is in the initial stages of development. In addition to various geologic, hydrologic and biologic aspects, the Museum will feature a history of mining and the mining industry, as well as multiple activities in which the public will be able to observe active mining, which is scheduled to resume in parts of the quarry. For additional information on the Cave House Museum, contact the Education Coordinator and member of the Board of Directors, Benson Guenther at the following:

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## *Trip B-3*

### **EXAMPLES OF ALONG-STRIKE CHANGES IN FOLD-THRUST BELT ARCHITECTURE; STRUCTURAL GEOLOGY OF THE ROSENDALE NATURAL CEMENT REGION, ULSTER COUNTY, NEW YORK**

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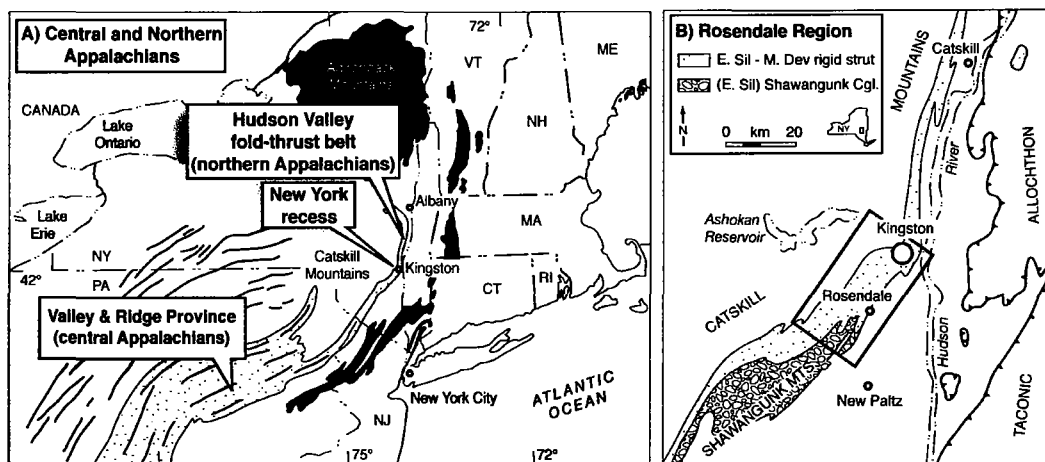
#### **INTRODUCTION**

*It was necessary for the leaders of this trip to specify a set number of vehicles to obtain permission to visit several of the following stops. For this reason, personal cars will not be allowed to follow the field trip caravan.*

Throughout much of the 19<sup>th</sup> century, the Rosendale natural cement region (Figure 1) was recognized worldwide as a center for the production of high-quality hydraulic cement. Although the industry collapsed in the early 1900s, the geology of this region continues to attract attention. The Rosendale area contains the northernmost portion of the central Appalachian foreland fold-thrust belt. This fold-thrust belt segment lies in the southern arm of the New York recess, a convex to the foreland, map-view curvature in the Appalachian orogen. Fold-thrust belt deformation across the New York recess involves a mechanically rigid strut of Late Silurian through Middle Devonian sedimentary strata (Figure 1; Wanless, 1921; Waines and Hoar, 1967; Rodgers, 1971; Marshak, 1986; Epstein and Lyttle, 1987; Marshak and Tabor, 1989; Marshak, 1990). This rigid strut is sandwiched between thick, relatively ductile units of Ordovician shale (below) and Middle Devonian shale (above). Near Rosendale, the Siluro-Devonian strut thins markedly from greater than 1000 m thick in central Pennsylvania to little more than 100 m thick north of Kingston. These contrasts in the pre-deformational stratigraphy of the affected units give rise to dramatic, along-strike transitions in the scale, style, and trend of structures in the fold-thrust belt across the New York recess. Preliminary balanced cross sections based on recent geologic mapping in the vicinity of Rosendale, combined with recently re-discovered historical photographs, are providing new insights into the structural relationships underlying these along-strike transitions in fold-thrust belt architecture. Specifically, the south to north thinning of the Siluro-Devonian rigid strut that takes place near Rosendale appears to trigger a south to north: 1) slight westward rotation of structural trends in the fold-thrust belt; 2) eastward migration of the pin line of deformation resulting in a narrower cross-strike fold-thrust belt width; 3) tightening of fold amplitudes and wavelengths; 4) decrease in the spacing between thrust faults and the development of lateral ramps; 5) northward dying out of thrust faults into fault-propagation folds, and 6) redistribution of regional detachment horizons.

The Rosendale natural cement region is ideally suited for the study of along-strike changes in the architecture of fold-thrust belts. In particular, the unique regional stratigraphy provides a basis for directly examining the role of changing mechanical stratigraphy in the development of along-strike changes in the internal architecture of these tectonic provinces. Further, the fold-thrust belt near Rosendale contains all of the structural complexities of regions like the Valley and Ridge Province of Pennsylvania, yet with a cross-strike width of little more than 2 to 8 km, it is possible to walk transects of the fold-thrust belt in one day. This excursion will focus on the following aspects of the Rosendale region's geology. The Rosendale region is an ideal place to examine: 1) fault mechanisms, including ramp-flat thrust fault geometries, accommodation faulting, and duplex structures; 2) fault-bend, fault-propagation, and detachment folding; 3) strain distribution in thrust sheets; and 4) the controls of rock type and structural relationships on cleavage morphology and intensity. In addition to structural relationships, the strata in

the Rosendale region preserve an abundance of Early to Middle Devonian marine fossils, so we can examine examples of classic shallow-marine carbonate facies successions (Rickard, 1962; LaPorte, 1969), and outcrops of the region provide exposures of the Taconic angular unconformity (Rodgers, 1971; Toots, 1976; Epstein and Lyttle, 1987).



**FIGURE 1:** Regional map (A) showing: central (Valley and Ridge Province) and northern (Hudson Valley fold-thrust belt) segments of Appalachian fold-thrust belt; New York recess; and cities of New York, Kingston, and Albany. Blackened areas delineate exposures of Precambrian, crystalline basement rocks in hinterland of orogen. Location map (B) shows: Rosendale natural cement region (outlined by heavy black box); cities of Kingston and New Paltz; Hudson River; and westernmost Taconic thrust sheet. Patterns show distribution of Late Silurian to Middle Devonian strata involved in fold-thrust belt deformation across New York recess. Modified after Marshak and Tabor (1989).

## HISTORY AND GEOLOGY AT ROSENDALE

From the 1820s through the early 1900s, the Rosendale natural cement region produced the highest-quality hydraulic cement in North America. For nearly a century following its discovery during the construction of the Delaware & Hudson Canal, miners in the Rosendale area quarried the Rosendale and Whiteport Members of the Rondout Formation, persistently following the beds around folds and across faults. The unique chemistry of these dolomitic strata required no further modification during the manufacture of cement, making them an ideal resource. Despite the dangers and arduous labor of cement mining, the abundance and quality of this resource in the Rosendale area sparked an explosion of local industry. So proficient was their pursuit of these rocks that the long-abandoned remains of their efforts are ubiquitous in the region's dark and overgrown corners. Thus, the history Rosendale region is deeply tied to its geology.

Fresh outcrops generated by Rosendale's nascent cement industry began drawing geologists as soon as they were uncovered. Notable early geologists, including Mather (1838, 1843), Darton (1893), Nason (1893), and van Ingen and Clark (1903), produced a series of comprehensive reports for the various publications of the New York State Museum. The natural cement industry collapsed during the early 1900s, leaving behind widespread quarry exposures. Although something of an environmental catastrophe, these abandoned quarries provide unique and otherwise inaccessible glimpses of structural relationships in the fold-thrust belt. For this reason, it is of little surprise that the Rosendale natural cement region has long served as a classroom for many colleges and universities teaching the basics of field geology, structure, stratigraphy, and sedimentology. Data collected by the countless field courses that pass through the Rosendale region represent a considerable and underappreciated legacy. Perhaps the

most significant of these collections consists of a series of reports held by Princeton University. Gilbert van Ingen, a Professor of Geology at Princeton University, taught annual field courses in the Rosendale area between roughly 1915 and 1923. Van Ingen's students painstakingly compiled the results of their fieldwork into a series of impressive senior and master theses (Hamil, 1916; Cairnes, 1920; Wanless, 1920; 1921; Osborne, 1921; Wiggans, 1923). These reports contain: photographs illustrating fresh quarry faces – exposures that were destroyed long ago or are presently overgrown; detailed surveys of exposures in the walls of subsurface mines that are now unsafe to enter; and dozens of carefully measured stratigraphic sections. The field relationships recorded in these old photographs are an invaluable resource for delineating the structural complexities of this region.

## **GEOLOGIC SETTING**

The Rosendale natural cement region lies along the western margin of the Hudson Valley, about 5 km southwest of the city of Kingston (Figure 1). The region encompasses the northernmost Shawangunk Mountains, the southern extent of the Helderberg Plateau, and portions of the Wallkill and Rondout-Esopus River Valleys. The Rosendale area is characterized by a series of northeast trending hills with rarely more than 50 to 100 m relief. This portion of the central Hudson Valley is underlain by a thick and strongly deformed sequence of Ordovician shale, siltstone, and greywacke (McBride, 1962; Waines et al., 1983; Vollmer and Bosworth, 1984; Kalaka and Waines, 1986; 1987). East of the Hudson River, these Ordovician strata form the footwall of the extensive thrust sheets of the Taconic allochthon. These thrust sheets include distal marine sedimentary strata that were thrust westwards over the North American passive margin during the Taconic Orogeny. The Ordovician rocks are separated from overlying strata by the regional Taconic angular unconformity (Rodgers, 1971).

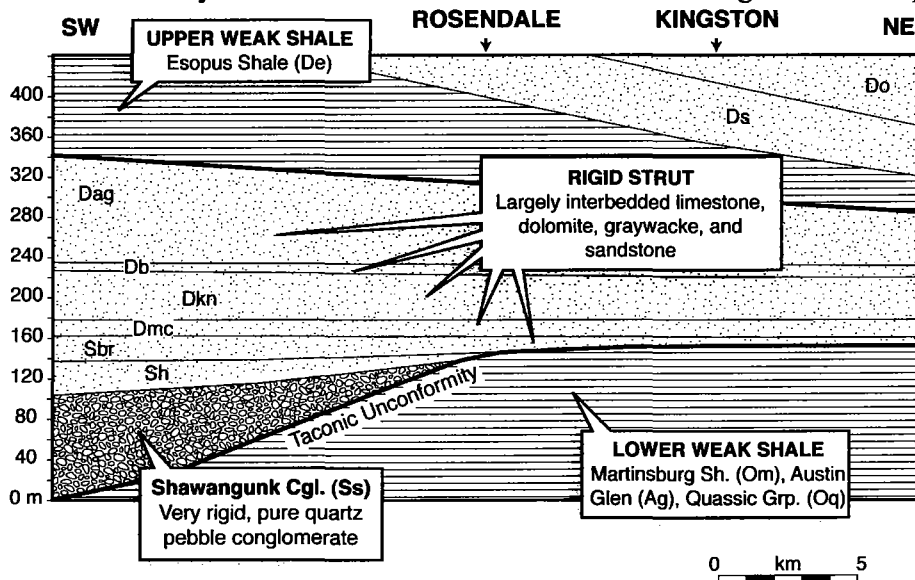
The Rosendale region includes the northernmost portion of the central Appalachian foreland fold-thrust belt. Deformation associated with the fold-thrust belt at Rosendale occurs within a 2 to 8 km-wide, northeast trending zone and involves Ordovician through early Middle Devonian strata. The Helderberg Escarpment, a 20 to 40 m high cliff, forms the eastern mapable extent of the fold-thrust belt at Rosendale. The escarpment comprises the sequence of Late Silurian through Middle Devonian strata that directly overlie the Taconic unconformity. The western margin of the fold-thrust belt lies in the foothills of the Catskill Mountains, west of the Esopus River Valley. The Catskill Mountains contain a thick sequence of Middle Devonian shale, sandstone, and conglomerate shed westward during the Acadian Orogeny. Strata within the Catskill clastic wedge are un-deformed and were either deposited subsequent to fold-thrust belt deformation, or lie beyond the western extent of deformation.

The fold-thrust belt at Rosendale lies within the southern arm of the New York recess (Figure 1), a regional curvature in the Appalachian orogen centered in near Kingston. Southwest of Rosendale, the fold-thrust belt widens into the Valley and Ridge Province of Pennsylvania. These structures continue north of Kingston in the Hudson Valley fold-thrust belt, which can be traced along the western edge of the Hudson Valley to the city of Albany. North of Albany, erosion across the Mohawk Valley has removed the strata involved in the fold-thrust belt.

## **STRATIGRAPHY**

Fold-thrust belt deformation in the central Hudson Valley region involves a sequence of Ordovician through Middle Devonian marine clastic and carbonate sedimentary strata (Figure 2; Waines and Hoar, 1967). Thick Ordovician turbidite sequences of the Austin Glen, Martinsburg, and Quassaic Formations underlie the lowlands of the Hudson Valley (Waines et al, 1983) and are separated from overlying strata by the Taconic angular unconformity (Rodgers, 1971). At Rosendale, the Taconic unconformity is overlain by a sequence of Late Silurian through Middle Devonian strata that form the highlands of the Shawangunk Mountains and the Helderberg Plateau. The base of this Siluro-Devonian sequence comprises a wedge of Late Silurian, near-shore marine strata. The stratigraphically lowest of the Silurian units near Rosendale is the Shawangunk Conglomerate, a massive, silica-cemented quartz pebble

conglomerate, which is overlain by the High Falls Shale. The High Falls Shale contains red and green shale and siltstone with locally interbedded limestone and dolomite. The High Falls Shale passes upwards



**FIGURE 2:** Schematic, along-strike stratigraphic chart showing the simplified relationships of Ordovician through Middle Devonian units in the vicinity of the Rosendale natural cement region. Note dramatic northward thinning of rigid stratigraphic strut of Siluro-Devonian strata. As units progressively pinch out, Late Silurian strata thin from nearly 150 m to roughly 10 m thick. The Shawangunk Conglomerate, a rigid quartz-rich unit, is stratigraphically lowest and mechanically most significant of Late Silurian units. Silurian strata are overlain by a roughly 125 m thick sequence of Early Devonian limestone, dolomite, and sandstone. The rigid strut is sandwiched between relatively ductile Ordovician Martinsburg Shale/Quassaic Group strata (below) and Esopus Shale (above). Abbreviated units are: High Falls Shale (Sh); Rondout and Binnewater formations (Sbr); Manlius and Coeymans formations (Dmc); Kalkberg and New Scotland formations (Dkn); Becraft Limestone (Db); Alsen through Glenerie formations (Dag); Schoharie Formation (Ds); and Onondaga Limestone (Do). Base line for stratigraphic section is top of Rondout Formation. Adapted from Wainess and Hoar (1967).

into the Binnewater Sandstone, a thin to moderately bedded quartz arenite with abundant sedimentary structures. The Rondout Formation, a thickly bedded sequence of dolostone and moderately fossiliferous limestone, overlies the Binnewater Sandstone. The thickness of the Silurian wedge changes dramatically near Rosendale as the lower units progressively thin and pinch out northwards (Wainess and Hoar, 1967). The thickness of the Shawangunk Conglomerate decreases sharply at the latitude of Rosendale, and is no longer present at Bloomington. The High Falls Shale and Binnewater Sandstone pinch out near Wilbur, leaving only the Rondout Formation north of Kingston. The Rondout Formation is present at the latitude of Catskill, where it is only 1 to 2 m thick.

Silurian strata are in turn overlain by the Early to Middle Devonian Helderberg and Tristates Groups, which record a series of transgressions in a shallow sea (Rickard, 1962; LaPorte, 1969; Sanders, 1969). The Helderberg Group contains two transgressive sequences. The first transgressive sequence begins with the near-shore tidal and beach facies of the thinly bedded Manlius Limestone and the wavy bedded, moderately fossiliferous, and locally cherty Coeymans Formation. These units are in turn overlain by the fossiliferous and increasingly argillaceous lime wackestone of the deeper-water facies of the Kalkberg and New Scotland Formations. The Becraft Formation overlies the New Scotland Formation. The Becraft Formation, the first unit in the second transgressive sequence of the Helderberg

Group, is a characteristically fossiliferous, near-shore, pinkish gray, coarse-grained lime grainstone, with locally interbedded green shale lenses. The Becraft Formation grades upwards into the progressively deeper-water argillaceous lime wackestone of the Alsen and Port Ewen Formations.

The Helderberg Group is overlain by the Tristates Group. The lower units in the Tristates Group are the Connelly Sandstone and the cherty limestone of the Glenerie Formation. The thick, black Esopus Shale overlies the Glenerie Formation, and grades upwards into the argillaceous limestone of the Schoharie Formation. The Schoharie Formation passes upwards into the reefal Onondaga Limestone, which contains abundant black nodular chert. The black, laminated shale of the Bakoven Member of the Union Springs Formation, abruptly overlies the Onondaga Limestone. The Union Springs Formation is the basal unit in the Hamilton Group, which is part of the Catskill clastic sequence.

### **STRUCTURAL FRAMEWORK**

Deformation in the New York recess of the Appalachian foreland fold-thrust belt involves a mechanically rigid strut of Late Silurian through Middle Devonian sedimentary strata (Figure 2; Wanless, 1921; Waines and Hoar, 1967; Rodgers, 1971; Marshak, 1986; Epstein and Lyttle, 1987; Marshak and Tabor, 1989; Marshak, 1990). This strut thins rapidly northwards along the strike of the New York recess, from greater than 1000 m thick in central Pennsylvania to little more than 100 m thick in southeastern New York. The strut is sandwiched between thick, relatively ductile units of Ordovician shale (below) and Middle Devonian shale (above). Such contrasts in the pre-deformational, mechanical stratigraphy (stratigraphic sequence defined in terms of rock rheology) of affected rocks are known to give rise to dramatic, along-strike changes in the architecture of fold-thrust belts (McDowell, 1998; Turrini et al., 2001; Soto et al., 2002). Specifically, in the vast Valley-and-Ridge Province of Pennsylvania, folds have kilometer-scale amplitudes and wavelengths, and large subsurface duplex structures occur at depth. At the apex of the Valley-and-Ridge Province in central Pennsylvania, the cross-strike width of the fold-thrust belt exceeds 140 km. The architecture of the Hudson Valley fold-thrust belt in the northern arm of the New York recess is similar to Valley-and-Ridge Province, but contains folds with amplitudes and wavelengths measured in 10s to 100s of meters. As a result, the fold-thrust belt in the northern arm is often as little as 2 km wide – almost two orders of magnitude smaller than in the Valley-and-Ridge Province of Pennsylvania.

Stratigraphic changes along strike in the New York recess appear to cause changes in the architecture of the regional detachment faults within the fold-thrust belt. At least two regional detachments underlie the Siluro-Devonian strut in the Hudson Valley fold-thrust belt, north of the city of Kingston in the New York recess. The uppermost of these horizons, the Rondout detachment, lies within the base of the Siluro-Devonian strut (Marshak, 1986; Marshak and Engelder, 1987; Marshak and Tabor, 1989; Marshak, 1990). The Rondout detachment is folded in outcrops within the Hudson Valley, suggesting at least one additional detachment horizon at depth in the underlying Ordovician Martinsburg Formation (Marshak, 1986; Marshak and Engelder, 1987; Marshak and Tabor, 1989; Marshak, 1990). The exact number of detachment horizons in the Hudson Valley fold-thrust belt, however, is unclear since cross sections constructed in this region are unbalanced. Previously, the Rondout detachment was interpreted as extending south of the New York recess (Marshak, 1990). However, recent work suggests that a regional detachment fault does not occur at this stratigraphic position south of Kingston (Burmeister and Marshak, 2002). Therefore, shortening accommodated by the Rondout detachment north of Kingston is being redistributed to other faults as the Siluro-Devonian strut thickens southward through the New York recess, resulting in a fundamental change in the internal architecture of the fold-thrust belt.

### **ALONG-STRIKE CHANGES IN THE FOLD-THRUST BELT NEAR ROSENDALE**

The first-order structural architecture of the fold-thrust belt near Rosendale comprises a series of north-northeast trending, kilometer-scale anticlinoria and synclinoria that are subsequently faulted and folded by smaller, second-order structures. Subsequent erosion carved the Hudson Valley into the core of one such anticlinoria; exposing the underlying Ordovician strata and separating the Silurian-Devonian

strata of the Helderberg Plateau from outliers (Mount Ida and Becraft Mountain) on the east side of the Hudson River. Strain developed in the various units of the Siluro-Devonian strut by a variety of mechanisms, including folding, thrust faulting, pressure solution, and intracrystalline twinning.

Nowhere are the structural effects of along-strike stratigraphic changes to the rigid, Siluro-Devonian strut in the New York recess more apparent than in exposures near Rosendale. Recent geologic mapping of the fold-thrust belt in the Rosendale vicinity suggests changes in the mechanical stratigraphy profoundly influenced the architecture of the fold-thrust belt (Burmeister and Marshak, 2002). The northward thinning of the Shawangunk Conglomerate corresponds with a south to north: 1) slight westward rotation of structural trends in the fold-thrust belt; 2) eastward migration of the pin line of deformation resulting in a narrower cross-strike fold-thrust belt width; 3) tightening of fold amplitudes and wavelengths; 4) decrease in the spacing between thrust faults and the development of lateral ramps; 5) northward dying out of thrust faults into fault-propagation folds, and 6) redistribution of regional detachment horizons.

Perhaps the most visible change in the structural architecture of the fold-thrust belt occurs along the along the Helderberg Escarpment. The structures exposed in quarries along the Helderberg Escarpment between Bloomington (Stop 4 and Optional Stop B) and Rosendale include a series of left-stepping, northwest dipping thrust faults that ramp out of the cores of tight synclines and laterally ramp up section to the south. In contrast, structures exposed in the escarpment at the latitude of Kingston (Stop 3, Hasbrouck Park) are extremely complex, involving stacks of horses in a large duplex structure (Marshak and Tabor, 1989; Marshak, 1990).

Preliminary balanced cross sections suggest that two regional-scale detachment horizons exist at depth in the Ordovician strata near Rosendale. The upper detachment lies roughly 10 to 20 m below the base of the rigid Siluro-Devonian stratigraphic strut and is deformed by structures developing on the lower detachment. The lower detachment is located a depth approximately 150 m further beneath the upper fault. A large fault roughly matching the estimated depth of the lower detachment crops out in the Ordovician strata at the base of the Helderberg Escarpment along the Rondout Creek at Creeklacks, about 1.0 km east of Rosendale. Preliminary balanced cross sections and field observations suggest that detachments also develop locally within several units in the Siluro-Devonian strut, including the Shawangunk, Rondout, Manlius-Coeymans, and Kalkberg Formations. Hanging-wall and/or footwall flat geometries in the Shawangunk Conglomerate are exposed in outcrop at High Falls, the south end of the Rosendale trestle, and in quarries east of Tillson. Where faulted, the Rondout Formation usually shows evidence of pervasive bedding-parallel slip, flat-on-flat fault geometries, or shallow thrust ramps. Although rarely visibly faulted in outcrop (except for at Stop C), faults generated near the contact between Manlius and Coeymans Formations usually form flats or low angle ramps. Preliminary balanced cross sections also suggest that localized detachments develop in the Kalkberg Formation, particularly within tight folds. Marshak (1990) describes similar faulting in the Kalkberg Formation in road cuts along State Route 23 near Catskill.

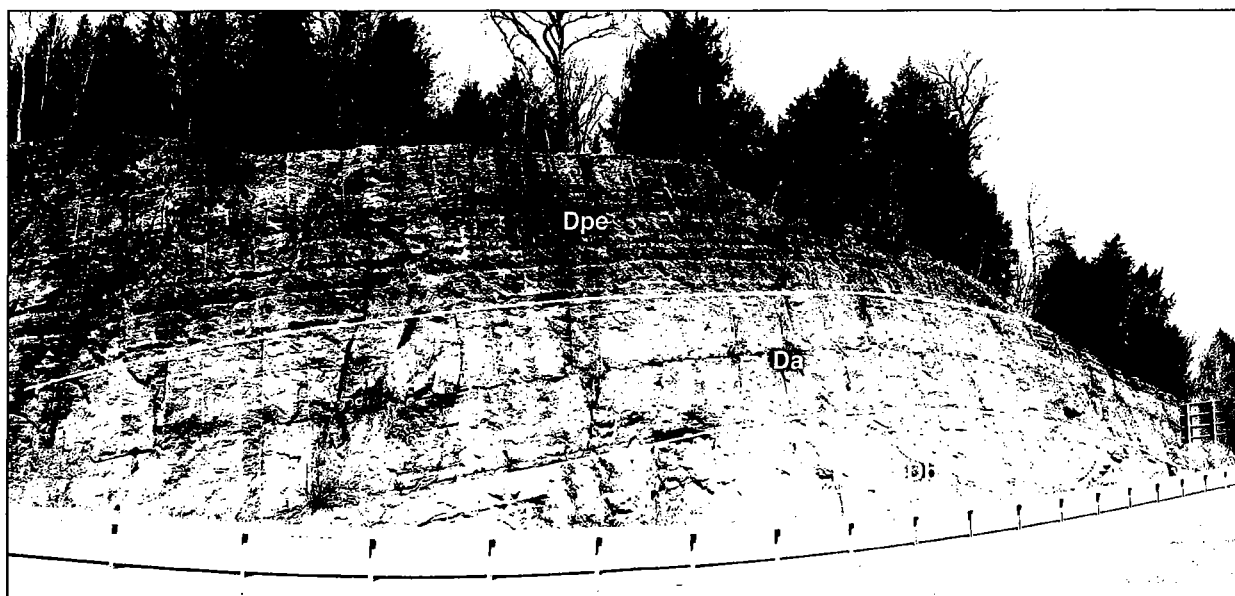
## STOP LOCATIONS

Our excursion will follow an oblique, north to south transect across the fold-thrust belt in the vicinity of the Rosendale natural cement region. This route will allow us to observe several of the along-strike changes in the structural architecture of the fold-thrust belt occurring in the southern arm of the New York recess. Our excursion will begin west of Kingston in the foreland of the fold-thrust belt. We will make our first stop in the vicinity of Kingston, where we will examine structures characteristic of the Hudson Valley fold-thrust belt. Here, we will also observe the structural complexity of the belt along its eastern margin in the Helderberg Escarpment. We will then proceed south into the northernmost Appalachian fold-thrust belt along the Helderberg Escarpment, where we will examine along-strike transitions in the structural architecture of the eastern margin of the belt. Near Rosendale, we will turn southwest towards High Falls, stopping at several locations in a roughly cross-strike transect of the fold-

thrust belt. Finally, we will return to Kingston to examine deformation associated with the upper detachment along the western margin of the fold-thrust belt.

### Stop 1: Folds exposed in road cuts along State Route 209 north of Kingston

During our drive east, we have passed from the foreland of the Appalachian orogen into the core of the Hudson Valley fold-thrust belt north of Kingston. Here, road cuts along State Route 209 expose large, slightly asymmetric open folds with no visible thrust faulting or mesoscopic folding. The geology of this stop was first described by McEachran (1985) and later used as a field trip stop by Marshak (1990). Five major, map-scale fold hinges cross State Route 209 between Routes 9W and 32. The most prominent of the folds at this stop is a large anticline (Figure 3) involving the upper Becraft Limestone, the Alsen Formation, and the Port Ewen Formation. Note the differential development of cleavage in these units. The Becraft Limestone contains little to no cleavage, while the Alsen and Port Ewen Formations clearly possess a strong, southeast-dipping cleavage. Marshak (1990) notes that these folds are characteristic of structural styles north of Kingston, in that they trend roughly  $015^{\circ}$  and lack structural complexity. Return to the vehicles and carefully merge back into traffic. As we proceed a short distance further to the east on State Route 209, notice the west-dipping beds of the Manlius through New Scotland Formations. **End.**

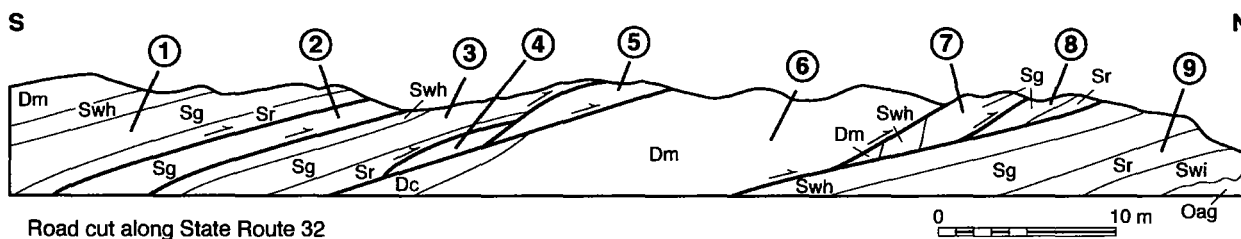


**FIGURE 3 (STOP 1):** Photograph of road cut exposure looking north showing first order anticline involving Becraft Limestone (Db), Alsen Formation (Da), and Port Ewen Formation (Dpe) along the north side of State Route 209/199, just east of intersection with State Route 32.



### Stop 2: Laterally ramping thrust faults in road cut along Route 32 north of Kingston

Four major thrust faults duplicate the Rondout and Manlius Formations in a series of imbricate thrust sheets in the northern half of the long road cut along the west side of State Route 32 (Figure 4). The geology of this stop was first described in detail by Waines and Hoar (1967) and was further interpreted by McEachran (1985) and Marshak (1990). The major thrust faults in this outcrop strike roughly  $040^{\circ}$  and calcite slip fibers on these faults suggest a transport direction of  $060^{\circ}$ - $070^{\circ}$ . In places, these faults have a flat-on-flat geometry (i.e. the thrust faults are parallel bedding in both the hanging wall and footwall), suggesting large displacements. In places (particularly in the upper portion of fault block 6; Figure 4), thrust faults appear to cut down section. However, this is likely an artifact of the obliquity of the road cut face to the structures, where the hanging walls are thrust westward into the outcrop. These faults could not be traced west of the road cut, but may ramp laterally up-section and die out to the north. Return to vehicle and continue south on State Route 32 South. From here, we will proceed southeast towards the Helderberg Escarpment **End**.



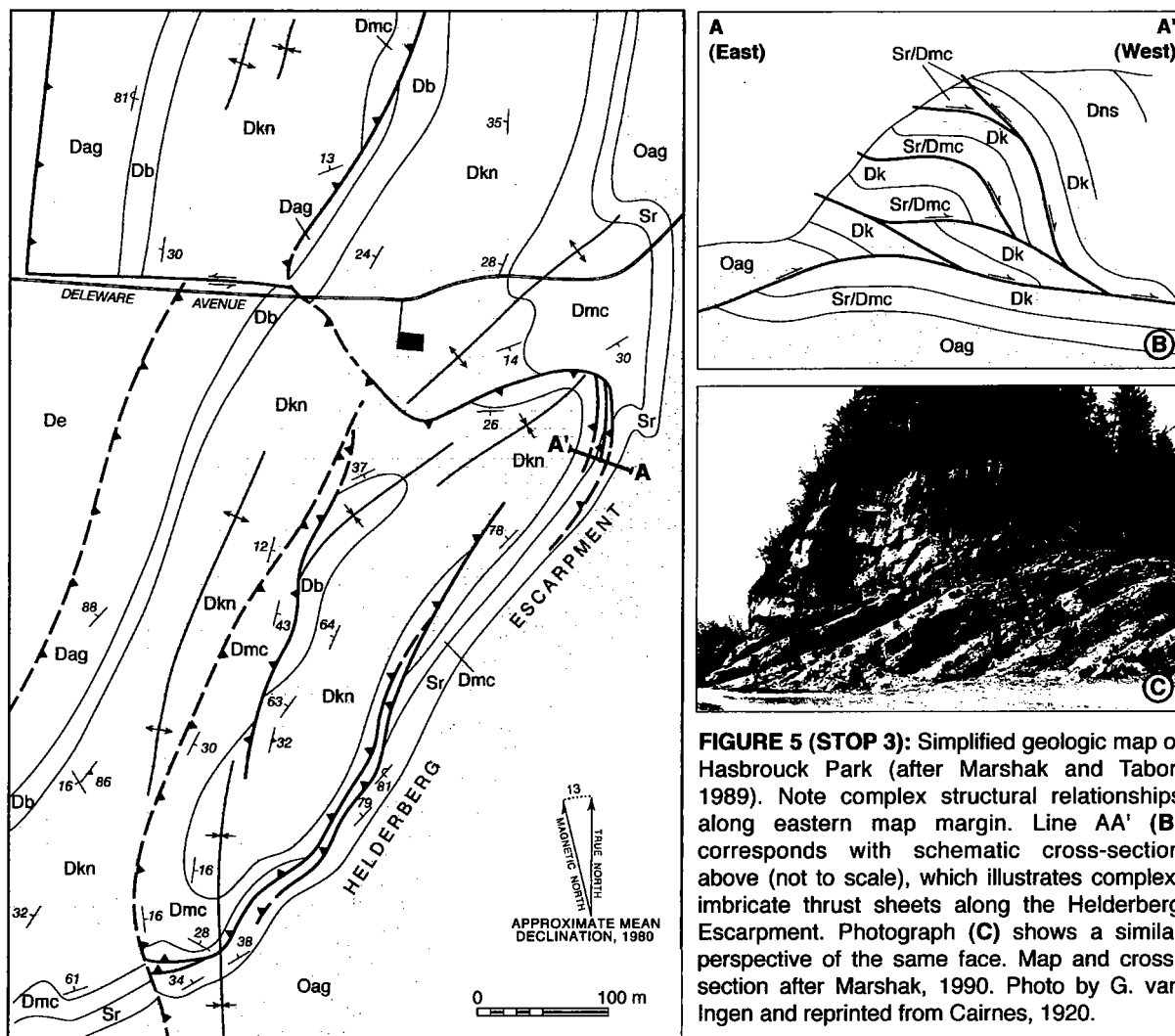
**FIGURE 4 (STOP 2):** Schematic cross section with slight vertical exaggeration of road cut exposure along west side of State Route 32 just south of intersection with State Route 199. Note duplication of Silurian and Devonian strata in series of imbricate thrust sheets and horses. The hanging wall (1) of the southernmost of these four major thrust faults contains Rosendale Member through Manlius Formation, which is thrust over a horse (2) of Glasco Member. These rocks are then thrust over a block (3) containing Rosendale through Whiteport Member rocks. Here, the Rosendale Member is thrust over two small horses (4, 5) of the Glasco Member, and over a block (6) containing rocks of the Manlius through Coeymans Formation. This block is then thrust over two small horses containing (7) Glasco Member through Manlius Formation and (8) Rosendale through Glasco Member, which are in turn thrust over (9) a sequence of Austin Glen Formation through Whiteport Member strata. Austin Glen Formation (Oag), Rondout Formation [Wilbur Member (Swi), Rosendale Member (Sr), Glasco Member (Sg), Whiteport Member (Swh)], Manlius Limestone (Dm), and Coeymans Formation (Dc). Figure redrawn after Marshak (1990).

### Stop 3: Complex structural relationships of Hasbrouck Park, east Kingston

Lock cars and follow the footpath at the east end of the parking lot for roughly 75 m as it descends around the nose of a small hill of Manlius, Coeymans and Kalkberg Formations south of the trail. At the foot of this slope is an abandoned adit cut into an extensively quarried face in the Helderberg Escarpment. The geology of the Helderberg Escarpment at Hasbrouck Park was first described by Marshak and Tabor (1989; Figure 5) and later revisited as a field trip stop by Marshak (1990). This face provides a cross-sectional view through the leading edge of a duplex composed of a stack of horses involving Rondout through Kalkberg Formation strata (Figure 5B). The basal thrust fault in this duplex, the Hasbrouck thrust, cuts laterally up section along the sloping footpath we walked along (Figure 5C), and through the woods. The Hasbrouck thrust fault reappears on the north side of Delaware Avenue, where it places the Manlius Formation over the Alsen Formation, above extensive roof-and-pillar cement mines just east of Corporate Drive. A tear fault with a trace roughly coincident with Delaware Avenue may extend to the west of the lateral ramp. The complexity of these structural relationships amongst the

Silurian and Devonian strata in Hasbrouck Park is characteristic of the geology along the Helderberg Escarpment at the latitude of Kingston.

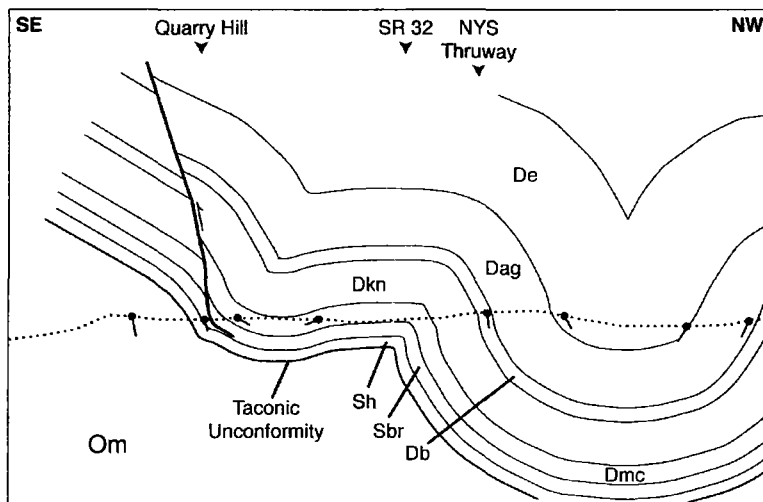
Return to main footpath and follow it further to the south, skirting along the base of the Helderberg Escarpment. Stay to the right at the first fork in the footpath and proceed up narrow ridge. From this locality, it is possible to observe steeply dipping to overturned beds exposed by quarrying of portions of the Rondout Formation. Retrace your path along the ridge back to the main footpath, return to the vehicles, and exit the park. We will now proceed south and examine some of the along-strike changes in structural architecture of the eastern margin of the northern Appalachian fold-thrust belt near Rosendale. **End.**



**FIGURE 5 (STOP 3):** Simplified geologic map of Hasbrouck Park (after Marshak and Tabor, 1989). Note complex structural relationships along eastern map margin. Line AA' (B) corresponds with schematic cross-section above (not to scale), which illustrates complex, imbricate thrust sheets along the Helderberg Escarpment. Photograph (C) shows a similar perspective of the same face. Map and cross-section after Marshak, 1990. Photo by G. van Ingen and reprinted from Cairnes, 1920.



**FIGURE 6 (STOP 4):** Photograph taken of eastern mine entrance at north end of Quarry Hill, looking south along mine cut in Rosendale Member (Sr) of the Rondout Formation. Eastern quarry wall is the gradational contact between Binnewater Sandstone (Sb) and Rondout Formation. Excavation removed all of Rosendale Member save support pillars; a thickness of nearly 4.5 m. Western quarry wall is the contact between the Rosendale Member and Glasco Member (Sg). Quarry affords an exceptional view of steeply dipping eastern limb and tight, west-verging hinge of a syncline.



**FIGURE 7 (STOP 4):** Schematic and un-balanced cross-sectional sketch of structural relationships in Quarry Hill vicinity, Drawn along a line drawn roughly perpendicular to the structural grain and is based upon data obtained during recent geologic mapping. Note tight syncline and associated out-of-the-syncline thrust fault in the Siluro-Devonian strata along the Helderberg Escarpment at Quarry Hill.

#### **Stop 4: North end of Quarry Hill, Bloomington**

This stop is on private property. You must obtain permission to visit from the owners of the house near the garage before proceeding. Walk the short distance to the garage on the edge of the tree line just south of the parking lot. Upon reaching the tree line, you will notice two depressions just through the trees to the south. These are the northern end of a series of extensive abandoned cement quarries spanning the entire length of Quarry Hill. Please proceed with caution.

First, walk into the western quarry pit. Here, a small quarry exposes east-dipping strata of the Rosendale and Glasco Members of the Rondout Formation. Retrace your path back to the edge of the tree line, and circle back around to the eastern quarry opening. Watch your step as you descend into this quarry. The slope is uneven and littered with broken glass and scrap metal. Here, the Rosendale Member of the Rondout Formation was extensively mined (all that remain are support pillars) and affords a fantastic view to the south along the axis and steeply dipping eastern limb of a northwest-verging syncline (Figure 6). The eastern quarry wall contains the gradational contact between the Rosendale Member and the underlying Binnewater Sandstone. The base of the Glasco Member forms the western quarry wall.

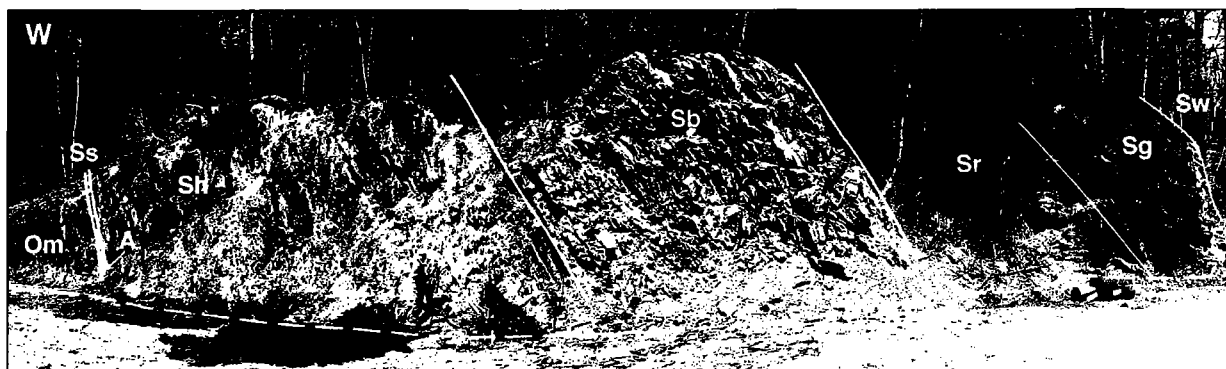
Note the tight hinge of the syncline and recall the relationship observed in the smaller quarry to the west. A northwest dipping, out-of-the-syncline thrust fault that apparently ramps up section out of the Rondout Formation separates the two quarries at this location (Figure 7). Thus, strata exposed in the western quarry are in the hanging wall, and were thrust to the southeast over the strata in the eastern quarry. Consider the contrasts in structural style between this location and the complex duplexes and the stacking of imbricate thrust sheets near Kingston. The quarry exposures at this location afford an excellent view of the structural relationships that are characteristic along the Helderberg Escarpment between Bloomington and Rosendale. Here, the escarpment is defined by a series of left-stepping, thrust faults that cut up section to the east, out of the cores of large synclines. Return to vehicles. Turn left (west) out of Bloomington Fire Company parking lot onto Taylor Road. **End.**

#### **Stop 5.1: Eastern limb of the Hickory Bush anticline, Rosendale Landfill**

Lock vehicles and walk approximately 100 m northeast along Hickory Bush Road to the entrance of the Rosendale Landfill and Recycling Center. Proceed through gate and to the north corner of the facility. Recent excavation along the north wall created fresh exposures of an east-dipping sequence of strata (Figure 8). The north wall of the landfill contains Martinsburg Shale, the Taconic unconformity, Shawangunk Conglomerate, High Falls Shale, Binnewater Sandstone, Rondout Formation, and Manlius Limestone.

From a distance, the Martinsburg Shale appears un-deformed, but closer examination reveals that it is strongly deformed by complex brittle faulting. Above the Taconic unconformity, the Shawangunk Conglomerate is little more than a 5 to 10 cm thick lag deposit of the characteristic milky white quartz pebbles. Overlying the Shawangunk Conglomerate is a complete and apparently continuous sequence of High Falls Shale and Binnewater Sandstone. Both the Rosendale and Whiteport members of the Rondout Formation are quarried at this location. Separating the two quarries is the Glasco Member, which contains beautiful *Halysites* chain corals. At the eastern end of the north wall are exposures of the overlying Manlius Limestone, which are in faulted contact.

The east-dipping strata exposed in the Rosendale Landfill form the eastern limb of the Hickory Bush anticline. The involvement of Martinsburg Shale in core of the fold suggests this structure developed as the result of slip along a detachment at depth in the underlying Ordovician strata. The large scale of the Hickory Bush anticline suggests that the underlying thrust is a master fault in the fold-thrust belt at this latitude and ramps directly from the lower detachment horizon. The thrust fault exposed at the eastern end of the exposures along the north wall is the westernmost fault in a complex imbricate fan of thrusts that ramp out of the Rondout Formation and cut through Hickory Bush Hill (the large hill southeast of the landfill). Return to the vehicles. **End.**



**FIGURE 8 (STOP 5.1):** Photograph of north wall of Rosendale Landfill at Hickory Bush showing southeast dipping Ordovician through Silurian strata in the eastern limb of the Hickory Bush anticline. Martinsburg Shale (Om) is in unconformable contact with Shawangunk Conglomerate (Ss), which is a 6.0 cm lag deposit of white quartz pebbles. High Falls Shale (Sh) and Binnewater Sandstone (Sb) are overlain by the Rondout Formation, out of which the Rosendale (Sr) and Whiteport (Sw) members were quarried leaving the Glasco Member (Sg). Dashed line denotes base of outcrop. Geologist for scale.

#### **Stop 5.2: Western limb of the Hickory Bush anticline, rail cut near Fourth Lake**

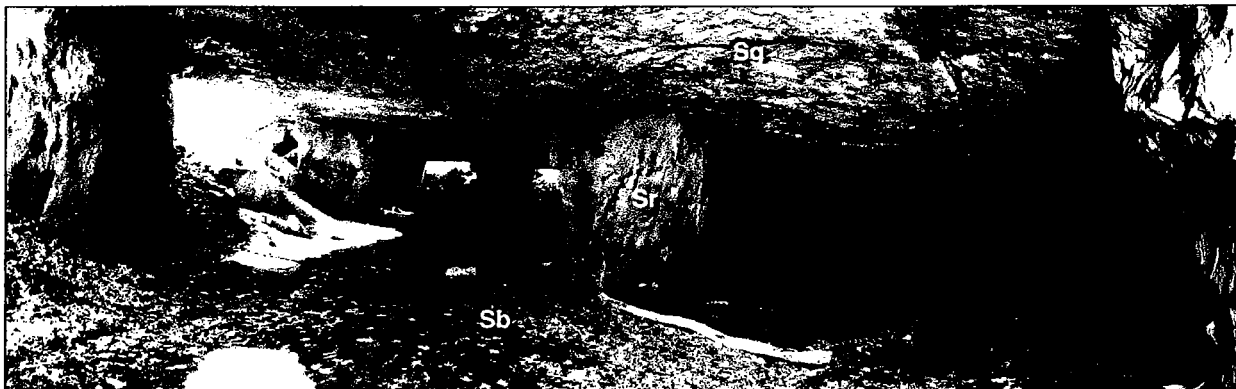
Walk southwest along the rail trail from the parking lot along Hickory Bush Road for roughly 350 m until you encounter a rail cut exposing moderately northwest dipping strata on both sides of the trail. The exposed sequence includes High Falls, Binnewater, Rondout, and Manlius Formations (Figure 9). Along the northwest side of the trail, the High Falls Shale is difficult to distinguish, but it is overlain by a complete thickness of the Binnewater Sandstone. Cross bedding, ripple marks, and graded bedding are clearly visible, as is a characteristic, roughly 10 cm thick shale horizon just below the contact with the overlying Rondout Formation. This shale layer is also present in an outcrop of the Binnewater sandstone at the Snyder Estate, where it is deformed. The Rosendale and Whiteport members of the Rondout Formation are quarried at this location. Notice how air circulating through the abandoned mines keeps this area noticeably cooler during the hot summer months.

Climb onto the embankment on the south side of the rail trail using the small path through the trees located just north of the northernmost quarry opening. The embankment was once a tramway that serviced the cement quarries in this area. Watch your step. Abandoned cement kilns and sunken shed foundations are scattered along this tramway and are often difficult to see. Proceed along tramway to the southeast, skirting along an exposure of the Binnewater Sandstone. When the tramway narrows and prevents further progress, drop down the slope to the next lower abandoned roadbed and continue to the southeast. After hiking approximately 150 m from the rail trail, you will reach an intersection with a dirt road. This road is part of the Perfume Trail on the Williams Lake Resort. Follow the dirt road to the right (southwest) and you will soon see a large mine entrance. Proceed about 10 m into the mine.

This mine was operated by the Lawrence Cement Company (Werner, personal communication, 2003) and is cut into the Rosendale Member of the Rondout Formation. The contact with the underlying Binnewater Sandstone forms the floor of the mine, and the contact with the overlying Glasco Member forms the ceiling (Figure 10). The quarry is in a large, asymmetric, open anticline. Note how the shallow dip of strata in the western limb of the fold gently increases to the southwest. Proceed southwest into the mine along the axis of the fold. Note the relatively rapidly increasing dip of the strata in the eastern limb of the fold. Quarrying of the eastern limb was most likely halted due to the proximity of a fault with a trace that roughly follows the path of the Perfume Trail a few meters further to the east.

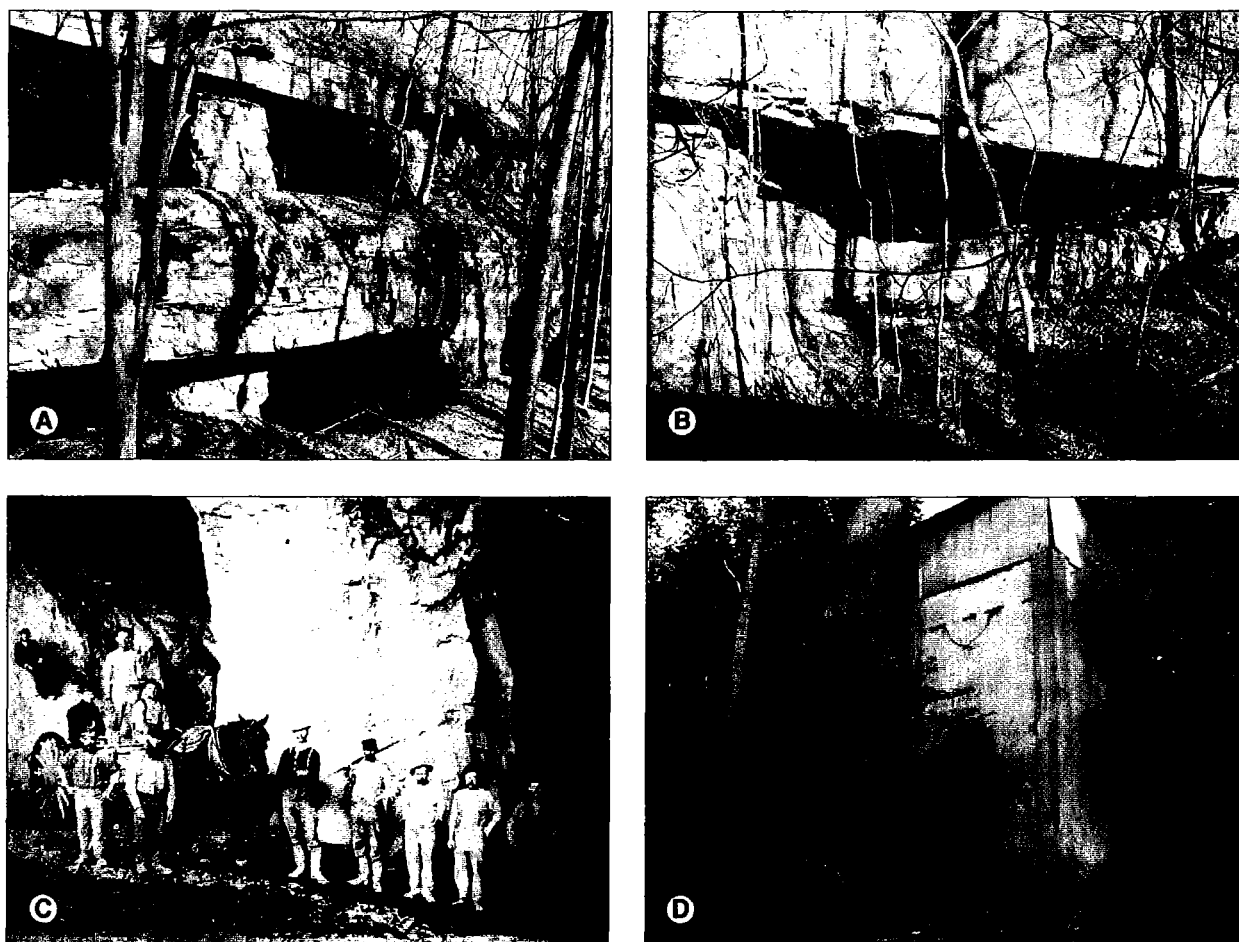


**FIGURE 9 (STOP 5.2):** Photograph looking north at railroad cut (presently a rail trail) between Fourth Lake and Hickory Bush. Cut exposes northwest dipping Silurian and Devonian strata in the western limb of the Hickory Bush anticline: High Falls Shale (Sh), Binnewater Sandstone (Sb), Rondout Formation [Rosendale Member (Sr), Glasco Member (Sg), Whiteport Member (Sw)], and Manlius Limestone (Dm). Photo by G. van Ingen and reprinted from Osborne, 1921.



**FIGURE 10 (STOP 5.2):** Photograph looking south along the shallowly dipping, western limb of anticline exposed in Lawrence Cement Company quarry near Fourth Lake. Binnewater Sandstone (Sb), Rosendale Member (Sr), and Glasco Member (Sg).

Climb up and out of the mine through the opening left by the partial collapse of the roof at the southeast end of the quarry. As you climb out of the mine, pause to note the other portions of the quarried fold around the periphery of the collapsed area. It is unclear whether strata of the Whiteport Member above the quarried fold was removed by erosion or was stripped by quarrying. Follow the footpath path down to the Perfume Trail and continue to the south until you reach a intersection of several trails. Return to the vehicles by walking north along the rail trail (covered in black gravel) from this junction. Along the way, note the various quarries and outcrops as you pass through the Fourth Lake region. The Williams Lake Resort property around Fourth Lake contains some of the best exposures of structural relationships in the region, and has long been used to teach the methods of field geology. **End.**



**FIGURE 11 (STOP 6):** Photograph of the Van Tassel's Quarry (A) on the Snyder Estate. Van Tassel's Quarry is in the footwall of the Century thrust fault (Optional Stop C). Mining of hydraulic cement was conducted using the room-and-pillar method in which a series of individual rooms (B) are quarried and eventually interconnected. The tenacity with which these mines were worked is apparent, and even more impressive considering that most quarries were worked entirely with hand tools and dynamite. Photograph (C) shows workers in front of the Widow Jane Mine, which is also on the Snyder Estate. The quarryman at center is holding star drills, which were driven into rock with sledgehammers to slowly carve out holes for black powder blasting charges. Rock collected in quarries was separated into cobble-sized fragments in screening house (D). Crushed rock was then burnt in cement kilns to remove  $\text{CO}_2$  before being ground into the powdered final product. Photograph C courtesy of the Century House Historical Society.

### Stop 6: The Snyder Estate

This land was originally owned by the Snyder family, who were central players in the local cement industry since its beginning during the construction of the Delaware & Hudson Canal (Werner, personal communication, 2003). The Century House Historical Society now maintains the Snyder property. This non-profit educational organization is dedicated to preserving the unique history of the Rosendale natural cement region. Leave cars and walk towards the battery of cement kilns at the east end of the lot. Continue eastward along the small trail leading down a small slope and through a small stand of trees. Proceed along the base of another battery of cement kilns to the driveway of the small, adjacent house. *This lot is private property. Please obtain permission from the owners before continuing along this route.* Use the flight of small stone steps directly behind the house to climb up to the abandoned tramway atop the embankment. Walk eastward along the abandoned tramway for several meters until you reach a set of small cuts. These cuts expose shallowly east-dipping beds of the uppermost Binnewater Sandstone. Note the characteristic shale horizon (previously seen at Stop 5.2) and examine the well-developed cleavage duplex that suggests some degree of bedding-parallel slip.

Continue eastward on the abandoned tramway to Van Tassels Quarry, a section of extensively mined Rondout Formation in the footwall of the Century thrust fault (Figure 11A). Van Tassels Quarry is an excellent example of the classic Rosendale natural cement quarry. Mines are cut into the Rosendale and Whiteport members, which were historically referred to as the upper and lower cements, respectively. The Glasco Member, historically known as the middle ledge, is left un-quarried between the cement layers. If time permits, follow the tramway north along the front of the quarries. The Rosendale Member is several feet thicker at the Snyder Estate than is to the north and south. Continue on the tramway to the northwest, further up onto the embankment behind the small house. Return to the vehicles in the parking lot of the Snyder Estate from the north, along the top of the embankment. Pause along the way to examine the screening house (Figure 11D) and other remnants of the cement industry, but be careful to stay away from the large openings atop the cement kilns.

As we leave the Snyder Estate, note the Brooklyn Bridge ornaments atop the gateposts. Over 100,000 barrels of Rosendale Natural Cement were used in the construction of the Brooklyn Bridge because of its unparalleled strength. Other famous landmarks built with locally produced cement include the pedestal of the Statue of Liberty and the Wings of the United States Capitol Building (Werner, personal communication, 2003). Turn right (west) onto State Route 213. **End.**

### Stop 7: Bedding-parallel slip in Shawangunk Conglomerate, State Route 213 east of High Falls

Leave the cars and walk east along the south side of State Route 213. Here, steep, 4 to 5 m high cuts in the uppermost Shawangunk Conglomerate line both sides of the road. The contact with the overlying High Falls Shale is roughly at the top of the road cuts. The rocks exposed in this road cut are in the footwall of a thrust fault, which cuts roughly perpendicular to the road at the east end of these cuts. As you proceed eastward, note the next set of road cuts in the Shawangunk Conglomerate. The strata exposed in the cuts further to the east are in the hanging wall of the thrust fault.

When you have nearly reached the eastern end of the lower (westernmost) road cuts, stop and examine the exposure along the north side of the road, west of the sign announcing High Falls District. Note the distinct, bedding-parallel slip surface (Figure 12), then carefully cross the road and examine this horizon up close. The slip surface contains a cleavage duplex composed of powdered quartz that gives a top to the west sense of shear. However, the lack of piercing points prevents a quantification of the amount of slip that has occurred on this surface. For this reason, it is unclear if the slip on this surface was the result of flexural-slip folding (suggesting little lateral displacement between bedding layers) or if it is a segment of a thrust fault exhibiting a flat-on-flat geometry (suggesting large lateral displacement).





**FIGURE 12 (STOP 7):** Photograph of road cut in Shawangunk Conglomerate (Ss) along north side of State Route 213 just east of High Falls. Bedding-parallel slip surface is indicated by solid white line. Approximate contact with overlying High Falls Shale (Sh) indicated with dashed white line. Road sign is roughly 4.0 m tall.



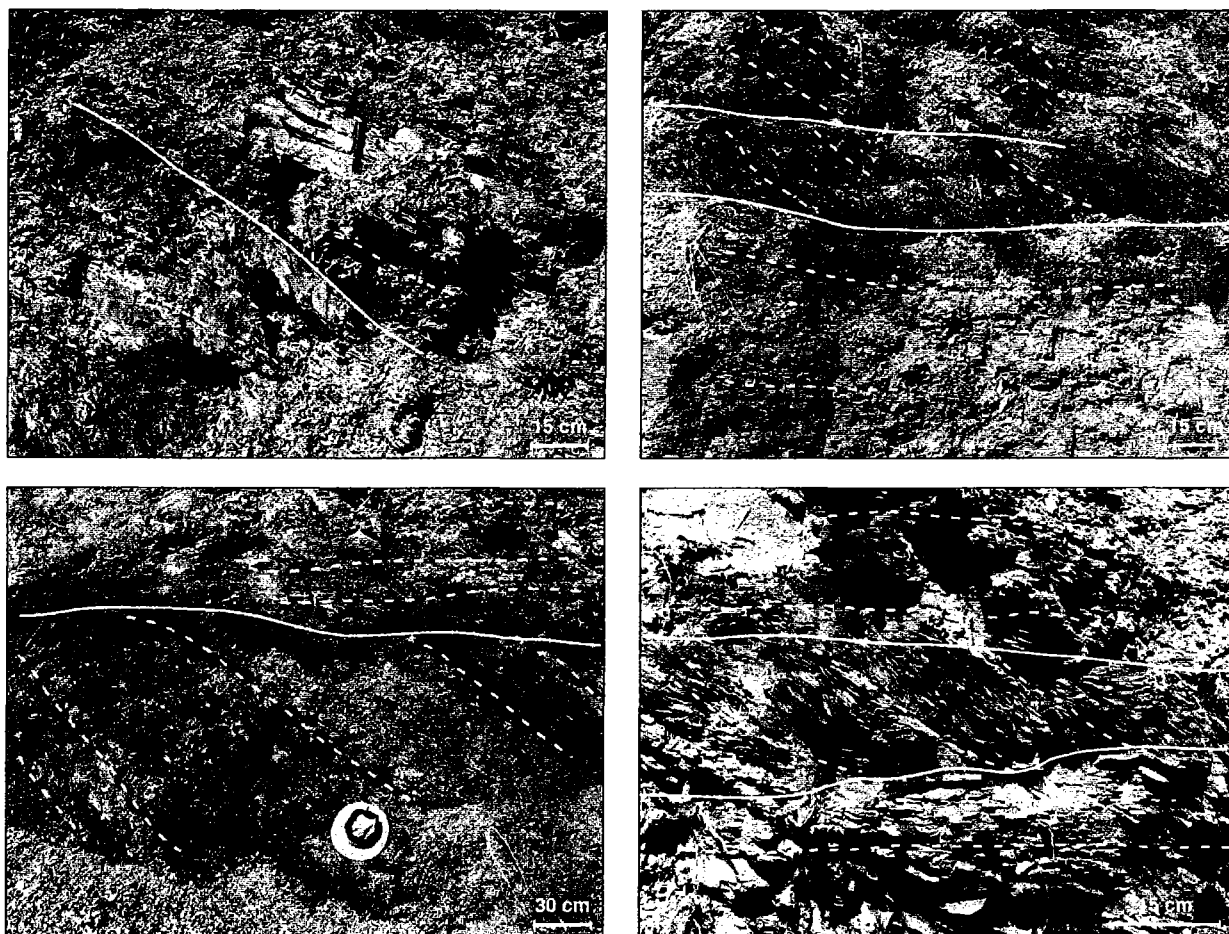
**FIGURE 13 (STOP 8):** Panoramic photograph of outcrop along north bank of Rondout Creek at the High Falls Park/Central Hudson power plant in High Falls, NY. Note large detachment fold in Binnewater Sandstone (Sb) and High Falls Shale (Sh) at center, the westernmost mesoscopic structure in the Rosendale-High Falls vicinity. Strata to east are part of the moderately northwest-dipping west flank of an anticline that roughly parallels Bruceville Road. Outcrop at fold is approximately 30 m high.

### Stop 8: Mesoscopic folding in Silurian strata, along the Rondout Creek at High Falls

Southwest of High Falls, the Rondout Creek flows roughly parallel to strike of the Rondout Formation and Manlius Limestone. At High Falls, the creek bends sharply to the east and begins cutting perpendicular to strike, down-section towards Rosendale. The highest falls are over an outcrop of the Rondout Formation. Downstream, smaller subsequent falls separated by shallow pools occur in the Binnewater Sandstone and High Falls Shale downstream followed by minor rapids caused by debris of the Shawangunk Conglomerate. As you follow the footpath past the High Falls power station northeast, note the shallowly northwest dipping beds of Binnewater Sandstone in the cut on the right. Once past the fenced-in area surrounding the power station, leave the trail and head north across the grassy area towards the wooded area long the creek. Stop to examine the millstones cut from Shawangunk Conglomerate placed as decoration in the grass. Also note the ruins of the stone building behind the cyclone fencing.

This was once the processing plant of the F.O. Norton Cement Company, which operated mines in High Falls and along Binnewater Road near Williams (Fifth) Lake (Werner, personal communication, 2003).

When you reach the waters edge, examine the asymmetric antiform in the Binnewater Sandstone and High Falls shale exposed along the north bank of the creek. This fold is the westernmost mesoscopic structure in the fold-thrust belt near Rosendale. What is unusual, however, is that the axial surface of the fold is dipping to the west, a sense of vergence opposite to the regional trend. The shape and scale of this fold suggest that it is a fault-propagation fold above a blind thrust fault. The fault underlying this structure is most likely a west dipping back thrust associated with the merger of two detachment horizons in the Ordovician strata at depth.



**FIGURE 14 (STOP 9):** Photographs of mesoscopic-scale cleavage duplex structures and shear zones in the Bakoven Member of the Union Springs Formation as described by Bosworth (1984) and Nickelson (1986). Duplex structures range between rough 10 cm and 1.0 m thick, and suggest the accommodation of blind thrust faulting of strata underlying this outcrop. All photographs are facing roughly north.

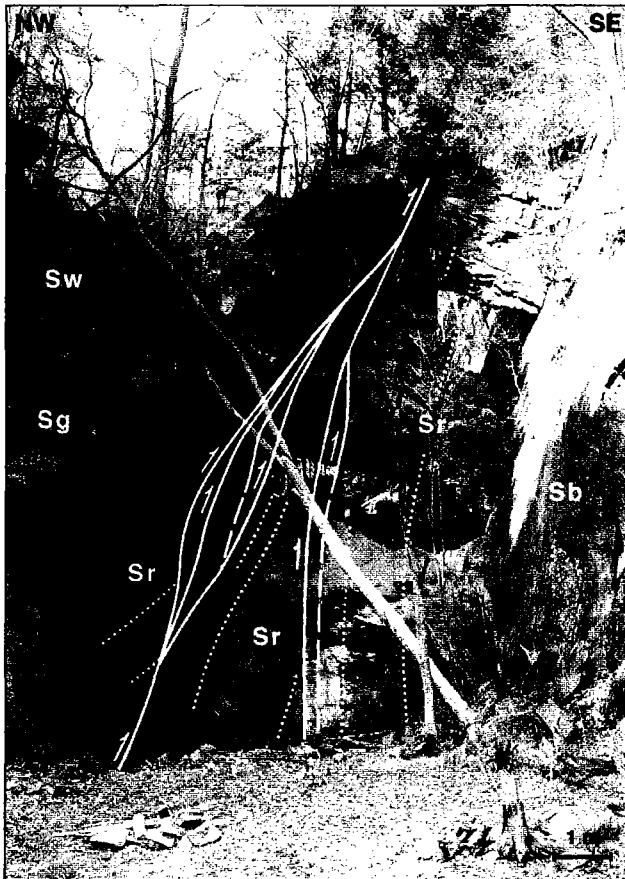
### Stop 9: Cleavage duplexes in Bakoven Shale, road cut along State Route 28, west of Kingston

Our excursion has taken us full circle, returning to the foreland of the fold-thrust belt. Here we can see evidence of deformation associated with the upper detachment of the fold-thrust belt as deformation dies out westwards. The steep escarpment and road cut along City View Terrace, beneath the Skytop Motel, contains an exposure of the two lower members of the Union Springs Formation, which is the basal unit of the Hamilton Group. Here, the upper 10 m of the fissile black shale of the Bakoven Member grade upwards into the lower 23 m of the Stony Hollow Member, a sequence of the buff-weathering shale and siltstone (Ver Straeten and Brett, 1995). The structural relationships of this location have been described by Bosworth (1984) and Nickelson (1986). Numerous examples of mesoscopic-scale deformation are visible in the Bakoven Member along the road cut. These cleavage duplexes and shear zones (Figure 14), often less than 20 cm thick, suggest that blind thrusts cut the underlying strata.

### OPTIONAL STOP LOCATIONS

#### Stop A: Road cut exposure of folds along Route 32, north of Rosendale

The road cut along the west side of State Route 32 at this locality exposes a tight, northwest verging anticline in the Kalkberg Formations. This structure is a continuation of the same series of folds exposed in the mines along Quarry Hill (Figure 7). The dip of the strata in the western limb of this fold continues to increase westward. Exposures of Becraft through Port Ewen Formations are steeply dipping to overturned in road cuts along the New York State Thruway. **End.**



**FIGURE 15 (STOP B):** Photograph looking northeast of abandoned quarry in Rosendale (Sr) and Whiteport (Sw) members of the Rondout Formation at south end of Quarry Hill. Large excavation at center is quarried out of the Rosendale Member, which is duplicated by a series of northwest dipping, out-of-the-syncline thrust faults. Although, not continuous, these faults are similar to those seen at Stop 4. Binnewater Sandstone (Sb) forms the eastern quarry wall. This surface contains a beautiful set of non-coaxial slip fibers. A folded sequence of the Glasco (Sg) and Whiteport members of the Rondout Formation from the western quarry face. Heavy lines denote fault surfaces. Dotted lines highlight bedding.

### **Stop B: South end of Quarry Hill, Bloomington**

*This stop is on private property and permission to visit must be obtained in advance. Trespassers will be prosecuted.* Walk north along the small dirt road located across from the house, keeping to the right as the road descends into the mouth of an abandoned cement quarry. The quarry at this locality is the southernmost of the Quarry Hill mines. Both the Rosendale and Whiteport Members of the Rondout Formation are extensively quarried at this location (Figure 15). Steeply dipping beds of the base of the Rosendale Member and the Binnewater sandstone form the eastern quarry wall. Moderately dipping beds of the Rondout through Manlius Formations form the western quarry face. The large, central quarry area exposes strata of the Rosendale Member that are repeated by two major thrust faults and extensively cut by a number of smaller thrusts. The easternmost of the two major thrust faults is steeply west dipping and exhibits a hanging wall ramp on footwall flat geometry. The western major thrust fault is moderately west dipping and also has a hanging wall ramp on footwall flat geometry. The western fault contains a horse of Rosendale Member. The two major thrust faults merge near the top of the quarry. The hanging wall of these faults is involved in a northwest-verging syncline, but is apparently not faulted. Some amount of bedding-parallel slip has also occurred along the eastern quarry wall. A close examination of this surface reveals a well-developed set of non-coaxial slip fibers. The faults exposed at this location are not continuous with those exposed at the north end of Quarry Hill (Stop 4). **End.**

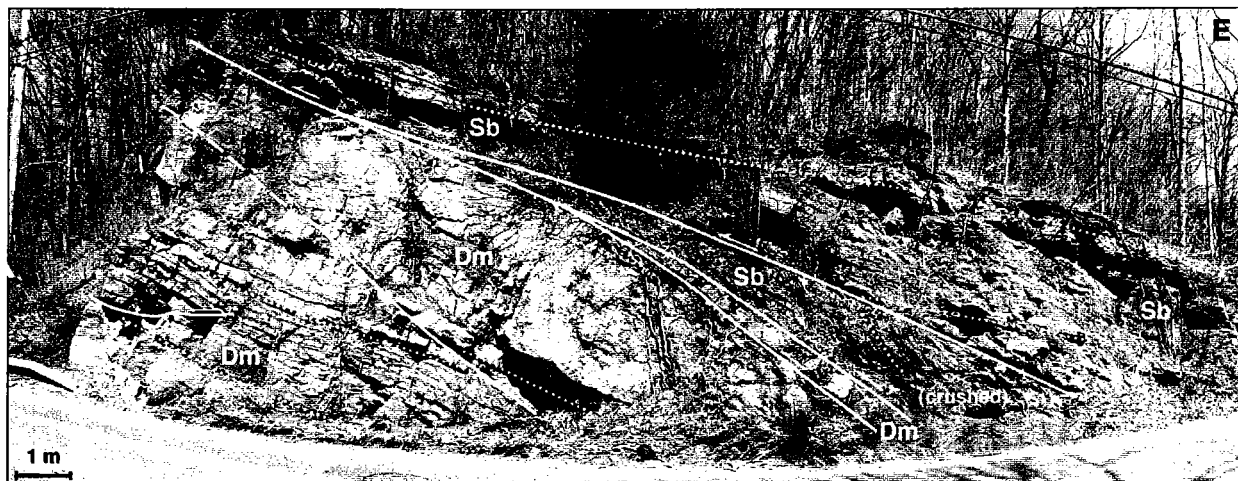
### **Stop C: Century thrust fault, road cut along State Route 213 east of Rosendale**

*Portions of this stop are on private property and permission to visit must be obtained in advance. Trespassers will be prosecuted.* Park vehicles out of the way of traffic in the lot and walk westward along the north side of State Route 213, keeping the guardrail between you and the traffic. The large quarried face to the north is the Butler's lock Quarry and includes a continuous sequence of Rondout through Kalkberg Formations. As you proceed along the road, note how the strata in the Butler's lock Quarry are folded into a small syncline. Once at the western end of the guardrail, very carefully cross to the south side of State Route 213 and quickly climb over the guardrail. Continue westward along the south side of State Route 213, again keeping the guardrail between you and traffic, until you reach the center of the large road cut along the north side of the road (Figure 16).

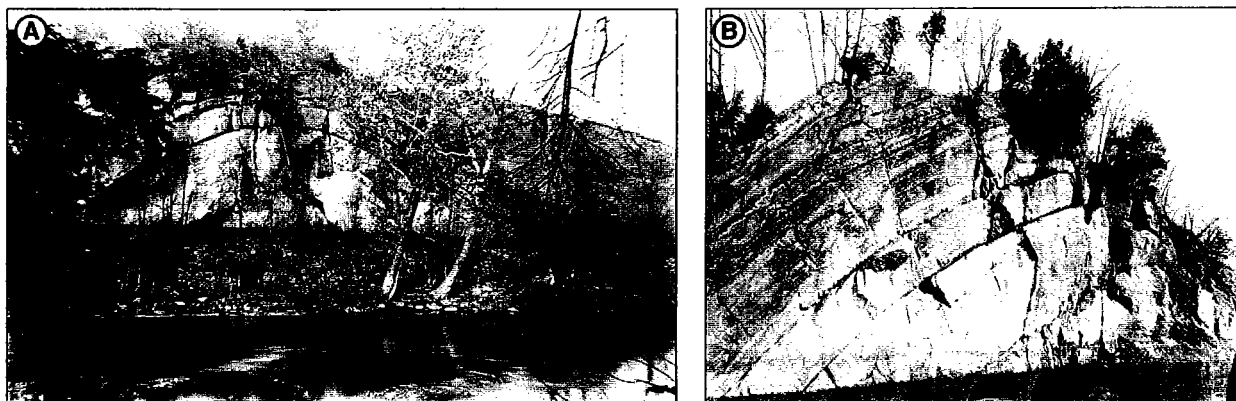
The road cut exposes the Century thrust fault, which places Binnewater Sandstone over Manlius Formation. The hanging wall contains a low-angle ramp which transitions into a hanging wall flat. Similar stair-step geometry of shallow ramps and flats in the Binnewater Sandstone can also be observed in a faulted sequence southwest of Tillson. The fault zone contains what is likely a horse of crushed Binnewater Sandstone and a sliver of Manlius Limestone. The footwall is a flat in the Manlius Limestone. The footwall is also cut by several minor thrust faults, the westernmost of which appears to be a wedge fault. Thus, the hanging wall of the Century Thrust contains the Butler's Lock Quarry. The footwall contains the Van Tassel's Quarry, which we will visit in Stop 6. Carefully return to the vehicles, and continue west on State Route 213. **End.**

### **Stop D: Lawrenceville anticline**

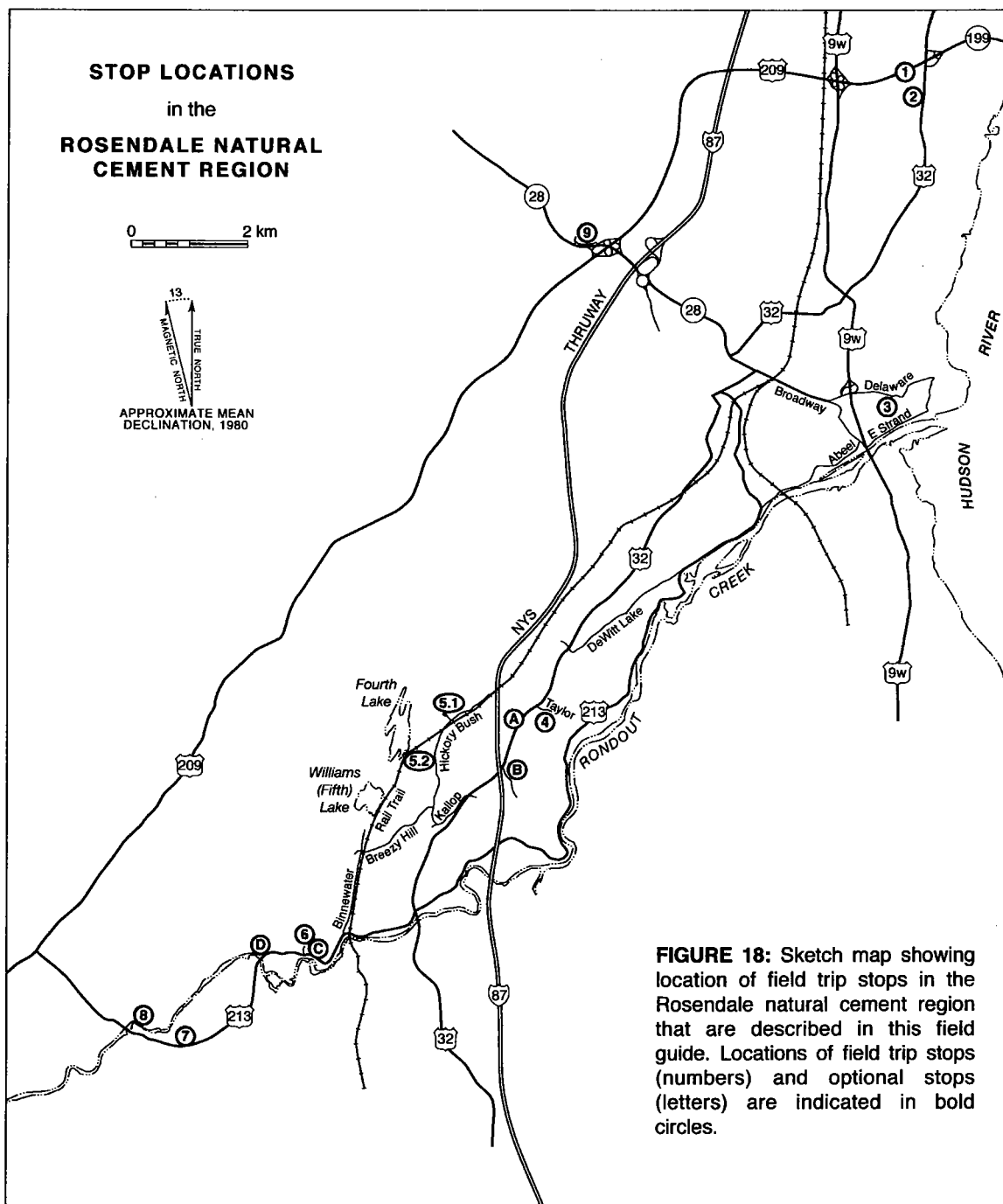
Leave the vehicles and cross to the north side of State Route 213. Walk to the center of the bridge over the Rondout Creek and look to the west. Although it may be difficult to see during the spring and summer months, this position affords a good view of the Lawrenceville anticline. The anticline is the westernmost major, map-scale fold exposed in the Rosendale region. The Lawrenceville anticline appears in numerous photographs taken of the Delaware & Hudson Canal in this region (Figure 17). **End.**



**FIGURE 16 (STOP C):** Photograph, looking north, of Century thrust fault exposed in a road cut along north side of State Route 213 between Butler's Lock Quarry (presently Turco Bros. Water Service) and Van Tassels Quarry on the Snyder Estate. Here, a hanging wall ramp and flat in Binnewater Sandstone (Sb) is thrust over a footwall flat of Manlius Limestone (Dm). Note horse of crushed Binnewater Sandstone and sliver of Manlius Limestone. Heavy lines denote fault surfaces. Dotted lines highlight bedding



**FIGURE 17 (STOP D):** Two photographs of the best-exposed anticline at Lawrenceville, along the Rondout Creek, taken during the early 20th century. Today, the outcrop is badly overgrown and often difficult to see from State Route 213 except during the spring and summer months. This anticline involves Kalkberg and New Scotland formations, and was exposed by a cut made during the construction of the Delaware & Hudson Canal. Photograph (A) shows a partially obscured anticline when viewed looking from the south bank of Rondout Creek. Photograph (B) is also taken looking north, but from the towpath atop the canal retaining wall. Photograph (A) by Cairnes (1920), photograph (B) by G. van Ingen and reprinted from Osborne (1921).



## ROAD LOG (Figure 18)

| Cum   | Part |  |
|-------|------|--|
| 0.0   | 0.0  | Road log mileage begins at the base of the exit ramp from State Route 28 East. Merge onto State Route 209 North heading towards Rhinecliff Bridge  |
| 0.1   | 0.1  | <i>Road cut</i> exposures of shallowly west-dipping beds of the Union Springs Formation along the left (northwest) side of Route 209 North   |
| 2.4   | 2.3  | <i>Bridge</i> over New York State Thruway  |
| 2.8   | 0.4  | <i>Bridge</i> crossing Esopus Creek  |
| 3.6   | 0.8  | <i>Underpass</i> for 9W, continue on State Route 209 N   |
| 3.8   | 0.2  | <i>Road cut</i> : Broad, open anticline in Schoharie Formation and Onondaga Limestone.   |
| <hr/> |      |  |
| 4.3   | 0.5  | <b>STOP 1:</b> <i>Pull off onto the shoulder of the road</i> and park in the adjacent grassy area.   |
| 4.6   | 0.3  | <i>Bridge</i> over State Route 32, stay in right-hand lane.  |
| 4.7   | 0.1  | <i>Exit for State Route 32</i> , turn right and proceed to the end of the ramp.  |
| 4.9   | 0.2  | <i>Stop sign, turn left onto State Route 32 South</i> and proceed for a short distance. If your group is large, you may want to park in the small, unmarked gravel parking lot directly across from the base of the exit ramp and walk to Stop 2.  |
| <hr/> |      |  |
| 5.2   | 0.3  | <b>STOP 2:</b> <i>Pull off onto the shoulder of the road</i> as far as possible, but use caution as the shoulder is low.   |
| 7.5   | 2.3  | <i>Turn left</i> onto entrance ramp for Route 9W South (Frank Koenig Boulevard.)   |
| 8.5   | 1.0  | <i>Turn right onto exit for Delaware Avenue.</i> Stay to the left as you proceed to end of ramp.   |
| 8.8   | 0.3  | <i>Traffic light</i> , intersection of exit ramp and Delaware Avenue. Turn left onto Delaware Avenue, keeping to the right-hand side of the road as it crosses back over Route 9W  |
| 8.9   | 0.1  | <i>Follow Delaware Avenue as it veers</i> to right, then back to the left before entering the historic district (caution 15 mph speed zone)  |
| 9.1   | 0.2  | <i>Traffic light</i> , intersection with Murray Street   |
| 9.2   | 0.1  | <i>Traffic light</i> , intersection with Corporate Drive   |
| <hr/> |      |  |
| 9.3   | 0.1  | <b>STOP 3:</b> <i>Small, hidden lane between last two houses on right-hand (south) side</i> of road. Turn right onto entrance road to Hasbrouck Park just before Delaware Avenue begins to descend steeply over the Helderberg Escarpment. Park in small gravel lot on left (east) side of lane, across from schoolyard. |
| 9.4   | 0.1  | <i>Turn right at intersection onto Delaware Avenue</i> , and follow the road down the hill (caution - tight turn)  |
| 9.6   | 0.2  | <i>Intersection with Abruyn Street</i> at base of hill, continue straight on Delaware Avenue. Note entrances to the abandoned Delaware Mine on left (west) side of road. This mine was most recently used by the Knaust Brothers to grow 'Fallout Proof' mushrooms during the 1950s.                                     |
| 9.7   | 0.1  | <i>Stop sign</i> , intersection with Lindsley Avenue, continue Delaware Avenue   |
| 9.9   | 0.2  | <i>Traffic light, turn right onto North Street.</i>  |

| <b>Cum</b> | <b>Part</b> |   |
|------------|-------------|---|
| 10.1       | 0.2         | <i>Traffic light</i> at intersection with East Union Street. Note abandoned canal barge in tidal flats off to left (east).  |
| 10.3       | 0.2         | <i>North Street bends to right and becomes East Strand Street.</i> Continue on East Strand Street.  |
| 11.0       | 0.7         | <i>Pass under Route 9W Bridge.</i> East Strand Street bends sharply to the right and becomes Broadway. The Mansion House located on this corner was built in 1833 by Jervis McEntee, one of the engineers that pointed to Rosendale as a local source for hydraulic cement for construction of Delaware & Hudson Canal (Werner, personal communication, 2003).  |
| 11.1       | 0.1         | <i>Intersection, turn left</i> onto Abeel Street  |
| 11.3       | 0.2         | <i>Traffic light</i> , intersection with Wurtz Street, continue on Abeel Street   |
| 11.5       | 0.2         | <i>Traffic light</i> , intersection with Dock/Ravine Streets, continue on Abeel Street  |
| 11.6       | 0.1         | <i>Traffic light</i> , intersection with Hunter Street, continue on Abeel Street  |
| 12.2       | 0.6         | <i>Pass under West Shore Railroad bridge.</i>   |
| 12.4       | 0.2         | <i>Intersection</i> with Wilbur Avenue  |
| 12.5       | 0.1         | <i>Traffic light</i> , intersection with State Route 213, continue on Abeel Street/ State Route 213 South. Note various ruins of cement kilns and associated buildings along next the road over the next kilometer.   |
| 13.6       | 1.1         | <i>Veer right onto DeWitt Lake Road</i> (County Route 28), and head up the east flank of Vly Mountain. This road follows the path of what was originally a plank road and later the roadbed of the horse-drawn Hickory Bush-Eddyville railroad. A consortium of cement companies constructed and maintained this thruway to avoid the shipping fees charged by the Delaware & Hudson Canal for moving their product from Hickory Bush to Kingston (Werner, personal communication, 2003). |
| 14.4       | 0.8         | <i>Stop sign</i> , intersection with the road to Eddyville, continue on DeWitt Lake Road  |
| 14.8       | 0.4         | Abandoned roadbed for Hickory Bush-Eddyville railroad diverges from course of DeWitt Lake Road and is visible in trees of to right (west)   |
| 15.1       | 0.3         | <i>Intersection</i> , turn left onto State Route 32 South   |
| 15.9       | 0.8         | <b>STOP 4:</b> <i>Turn left onto Taylor Street</i> , then immediately turn right into the parking lot of Bloomington Fire Company and proceed to back (southwest), gravel end of lot and park.  |
| 16.4       | 0.5         | <i>Intersection, turn left</i> onto State Route 32 South  |
| 16.6       | 0.2         | <b>OPTIONAL STOP A:</b> <i>Pull off highway and park in paved area along the road.</i> Please be sure to not block access to road that continues off to southwest.  |
| 16.8       | 0.2         | Road cut exposures of the Manlius Formation containing karst dissolution features along left (east) side of road.   |
| 16.8       | 0.2         | <b>OPTIONAL STOP B:</b> <i>Turn left onto private drive</i> just north of the bridge over the New York State Thruway. Proceed slowly down gravel road to the first house and park vehicles out of the way.  |
| 16.9       | 0.1         | Bridge over New York State Thruway  |



| <b>Cum</b> | <b>Part</b> |   |
|------------|-------------|---|
| 17.7       | 0.8         | <i>Turn right</i> onto Kallop (Corners) Road  |
| 17.9       | 0.2         | <i>Veer right onto Hickory Bush Road</i> at three-way intersection  |
| 18.0       | 0.1         | Intersection with Breezy Hill Road, continue straight (north) on Hickory Bush Road  |
| 18.5       | 0.5         | Well-preserved cement kilns on left-hand (northwest) side of road   |
| 18.6       | 0.1         | Entrance to the mines on right-hand side of road were once operated by the Lawrence Cement Company (Werner, personal communication, 2003), and are presently owed by Iron Mountain Incorporated as a future archival site.  |
| 18.8       | 0.2         | Cement kiln ruins on right-hand (southeast) side of road  |
| <hr/>      |             |   |
| 18.8       | 0.0         | <b>STOPS 5.1 &amp; 5.2:</b> Cross roadbed for Wallkill Valley Railroad (presently a rail trail maintained by John Rahl of Rosendale) and pull into dirt parking lot on left side of road  |
| 19.7       | 0.9         | <i>Exit parking lot and turn right</i> , heading south on Hickory Bush Road.  |
| 20.1       | 0.4         | <i>Stop sign</i> , turn right onto Breezy Hill Road.  |
| 20.6       | 0.5         | <i>Stop sign</i> , turn left (south) onto Binnewater Road (County Route 7) heading towards Rosendale. Binnewater Road follows the trace of the axis of the Binnewater anticline.  |
| 21.1       | 0.5         | The south quarries of the F.O. Norton Company in east-dipping units of the Rondout Formation located in the escarpment through trees on left (east) side of road.   |
| 21.2       | 0.1         | The Beach Mine of the Lawrenceville Cement Company (Werner, personal communication, 2003) in west-dipping strata of the Rondout Formation along right (west) side of road. Iron Mountain Incorporated now uses these mines as a document storage facility.  |
| 21.4       | 0.2         | The Hoffman Quarry of the Lawrence Cement Company located east of small Wallkill Valley RR trestle on left (east) side of road. A battery of abandoned cement kilns built by the Lawrenceville Cement Company is visible on the right (west) side of road (Werner, personal communication, 2003).   |
| 21.6       | 0.2         | Stone reservoir along right (west) side of road. This pond was created in the late 1820's to ensure a constant reserve supply of water for a feeder stream that was used to replenish water levels in the Delaware & Hudson Canal (Werner, personal communication, 2003).   |
| 22.0       | 0.4         | <i>Intersection, turn right (west) onto State Route 213.</i> Joppenbergh Hill is located immediately to the left (east) of this intersection. The hill is named after Jacob Rutsen, who became the first westerner to settle in the Rosendale vicinity in 1680. The New York and Rosendale Cement Company extensively mined Joppenbergh Hill during the 1800's. Large sections of the resulting quarry faces began collapsing in 1899, and have continued to do so as recently as April 2003. Rosendale Bridge, built by the Wallkill Valley Railroad, spans Rondout Creek from the flank of Joppenbergh Hill. Locally, State Route 213 follows the northern bank of Rondout Creek, and coincides with the trace of the Delaware & Hudson Canal. A thick, east-dipping sequence of Shawangunk Conglomerate cropping out in the south banks of the creek is visible through the trees. |
| 22.1       | 0.1         | Butler's Lock Quarry once operated by the Rosendale Cement Company. This mine is presently used as a water reservoir for Turco Brothers Water Service.  |

**Cum Part**

- 
- 22.1 0.0 **OPTIONAL STOP C:** *This is private property and permission must be secured in advance to park in this lot. Many aspects of this stop are extremely dangerous and should only be attempted by very small groups. Pull off State Route 213 into the parking lot of the Butler's Lock Quarry and the Turco Brothers Water Service, and park out of the way. Beware that semi trucks hauling full-sized water tanks enter this lot at full speed and use portions of this space to turn around. Care must be taken when parking and walking about the lot.*
- 22.2 0.1 Van Tassels Quarry visible through chain link fence along the right (north) side of road
- 
- 22.3 0.1 **STOP 6:** *Turn right into the Snyder Estate, and proceed up the driveway, across a small bridge, and park in the large, grass-covered parking lot on the right.*
- 22.6 0.3 Note Shawangunk Conglomerate in road cut immediately adjacent to the driveway of the Snyder Estate along the right (north) side of road.
- 22.9 0.3 Bridge over Rondout Creek
- 
- 23.0 0.1 **OPTIONAL STOP D:** *Pull off on small turnout on left (south) side of road. It is possible to park vehicles here for a short amount of time. For longer stays, the public parking lot located just across Rondout Creek from this location should be used.*
- 23.5 0.5 Note the outcropping of Shawangunk Conglomerate in escarpment across cornfield on left (east) side of road. These rocks are in the hanging wall of the Snyder thrust fault.
- 23.7 0.2 Intersection with Mossy Brook Road. Ruins of mill foundation along the creek behind the bed and breakfast inn is supposedly the site of one of the very first cement mills in this region (Werner, personal communication, 2003).
- 24.4 0.7 High Falls Motel on left (south) side of road
- 
- 24.4 0.0 **STOP 7:** *Pull off on right-hand side of road at base of hill. Caution, the road has a low shoulder and ground is usually fairly mucky. It is also possible to park around the corner to the north or to park for short amounts of time in parking lot on the south side of the street.*
- 24.5 0.1 Intersection with Bruceville Road (north)/Mohonk Road (south). D&H Canal Museum is located about 0.1 mile south of State Route 213 on east side of Mohonk Road
- 24.6 0.1 Intersection with Second Street, Blacksmith Shop where local dolostone was first burnt for cement located a few meters north of this intersection
- 
- 24.8 0.2 **STOP 8:** *Turn right into parking lot of High Falls park/Central Hudson Power Station. Lock vehicles and proceed down footpath towards power station.*
- 24.8 0.0 *Turn right (west) on State Route 213 from parking lot at High Falls park/Central Hudson Power Station*
- 24.9 0.1 *Bridge over Rondout Creek*
- 25 0.1 *Traffic light, intersection with County Route 1*
- 26.1 1.1 Small road cut in gently west dipping strata of the Kalkberg/New Scotland Formations.
- 26.2 0.1 *Traffic light, turn right onto State Route 209 North towards Stone Ridge*

| <b>Cum</b> | <b>Part</b> |   |
|------------|-------------|---|
| 28.9       | 2.7         | Intersection with Cottekill Road, continue on State Route 209 North   |
| 30.3       | 1.4         | Old abandoned quarry in Onondaga Limestone in trees along left (west) side of road  |
| 33.8       | 3.5         | Esopus Creek visible through trees on left (west) side of road. Onondaga Limestone underlies much of this portion of the valley                                   |
| 33.9       | 0.1         | Intersection with road to Town of Hurley. It was here that the government of New York was temporarily relocated when the British sacked Kingston in 1777          |
| 34.7       | 0.8         | Underpass   |
| 36.2       | 1.5         | <i>Bridge crossing Esopus Creek</i>   |
| 36.4       | 0.2         | <i>Merge from right-hand lane onto exit ramp leading to State Route 28 West toward Pine Hill, and proceed over bridge as it crosses back over State Route 209</i> |
| 36.8       | 0.4         | <i>Turn right onto Forrest Hill Drive</i>   |
| <hr/>      |             |   |
| 36.8       | 0.0         | <b>STOP 9:</b> Turn right onto City View Terrace Road, and pull off onto shoulder to park   |
| 37.0       | 0.2         | <i>Pull back onto City View Terrace Road and continue East</i>  |
| 37.2       | 0.2         | <i>Use parking lot of Potter Bros. Ski and Patio to turn around, heading back to the west on City View Terrace Road.</i>  |
| 37.2       | 0.0         | <i>Stop sign, turn left onto Forrest Hill Drive</i>   |
| 37.2       | 0.0         | <i>Traffic light, turn right onto State Rroute 28 West towards Pine Hill and return to Oneonta</i>  |

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## ACKNOWLEDGEMENTS

This work was partially supported by funds from the United States Geological Survey EDMAP Program and the Geological Society of America. The authors wish to thank: Russell Waines for guidance, advice, and support throughout this endeavor; Dietrich Werner and Gayle Grunwald for insights into the rich regional history and much needed assistance with lodging and local contacts; M. Scott Wilkerson for assistance with cross section construction and balancing; Chuck Ver Straeten for useful discussions on Devonian stratigraphy; Robert H. Fakundiny and Robert H. Fickies of the New York State Geological Survey; and Anita Peck, Phyllis Noreen, Warren Prandoni, Ryan Dolan, Michael DeFalco, and Philip Terpening and the Town of Rosendale for access to public and private lands. The authors would like to encourage you to read more about the history of the Rosendale natural cement industry and learn how you can support the work of the Century House Historical Society. Please visit them at <http://www.centuryhouse.org>.

## **Workshop W-1**

### **Hands-on Activities for Teaching about Earthquakes - an IRIS Workshop**

Jeffrey S. Barker, Department of Geological and Environmental Sciences, Binghamton University, Binghamton NY 13902

Accurate and timely seismological data can serve as a basis for a number of hands-on, guided inquiry-based learning exercises at the college, high school, middle school and upper elementary school levels. Workshop participants will perform several activities related to earthquakes and seismology appropriate as labs in college Geology courses or as learning activities within high school Earth Science or Physics classes. Topics include earthquake location, cause, distribution, recurrence, and magnitude. This workshop is sponsored by IRIS (Incorporated Research Institutions for Seismology) which provides the data as well as a variety of supporting materials (maps, posters, handouts, etc).

Workshop will run from 8:30 – 12:00.

## **Workshop W-2**

### **Teaching Environmental Science in the Outdoors: Pine Lake Environmental Campus**

Meredith Newman Department of Geology and Environmental Science, Hartwick College, Oneonta, New York 13820

Environmental science is by its nature interdisciplinary. This workshop will focus on the use of Hartwick College's Pine Lake Environmental Campus to illustrate the interactions between geology, biology, and chemistry, with emphasis on how these factors control the water quality in Pine Lake. Although this region of New York experiences some of the most acidic rain in the US, the pH of Pine Lake water has remained nearly neutral. We will explore how geology, biology, and chemistry explain this phenomenon. Several hands on assignments previously used to teach courses such as Introduction to Environmental Science, Earth Cycles, and Environmental Geology will be distributed and discussed. We will all also experience one of these assignments for ourselves.





## Appendix A

### Trip A-6

#### LATE OTTAWAN DUCTILE SHEARING AND GRANITOID EMPLACEMENT IN THE HUDSON HIGHLANDS, NY

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#### INTRODUCTION

The purpose of this field trip will be to examine rocks exposed in the Hudson Highlands, NY that elucidate the late Ottawaan (<1030 Ma) history of high-grade ductile shearing, migmatization, and the emplacement of a variety of granitoid plutons. Most research in the Grenville Province in eastern North America has focused on the history and processes associated with Elzevirian (1350-1180 Ma) and Ottawaan (1090-1030 Ma) orogenic events and intervening periods of extension and magmatism (see McLelland et al., 1996; Rivers, 1997; Hamner, 2000 for review). In contrast, comparatively little attention has been given to late- to post-Ottawan events and their importance to the history of the Grenville Province. The period between 1030 and 960 Ma is generally characterized by a diverse set of events that include both large-scale orogenic collapse along major extensional shear zones (e.g., van der Pluijm and Carlson, 1989; Carlson et al., 1990; Culshaw et al., 1994; Ketchum et al., 1998; Streepey et al., 2000) and localized high-grade metamorphism, thrusting, and magmatism (e.g., Lumbers et al., 1990; Mezger et al., 1991; Ratcliffe et al., 1991; Gower et al., 1991; Connelly and Heaman, 1993; Owens et al., 1994; Haggart et al., 1993; Jamieson et al., 1995; Corfu and Easton, 1997). Recent geologic mapping and structural analysis in the central Adirondack Highlands and the New Jersey/Hudson Highlands has also recognized the importance of late- to post-Ottawan, high-grade, ductile transpression (Gates, 1998; Allers et al., 2001; Valentino et al., 2001; Solar et al., 2003; Gates et al., 2001a; in press). This deformation is being taken up on large transcurrent shear zones with significant amounts of displacement (Gates, 1995; 1998). In the New Jersey/Hudson Highlands, the emplacement of a chemically diverse suite of granitoid plutons is intimately associated with this crustal-scale shearing event (Gorrington et al., 2002). A Middle Proterozoic escape tectonic event (e.g., Tapponnier et al., 1982) in the central Appalachians resulting from accretion to the north is interpreted to have produced this deformation and magmatism.

#### REGIONAL GEOLOGY

The Hudson Highlands, along with the physically contiguous New Jersey Highlands and similar rocks extending into eastern Pennsylvania, are collectively called the Reading Prong, one of the largest of several Grenville-age (1300 to 1000 Ma) basement massifs within the core of the Appalachian orogenic belt of eastern North America (Figs. 1 and 2; Rankin, 1975). These basement massifs lie outboard (east) of the main Grenville Province in eastern Canada and record variable amounts of post-Mesoproterozoic metamorphic and deformational overprint (e.g., Rankin et al., 1989; Gates and Costa, 1998). The Reading Prong displays evidence of only brittle deformation concentrated along narrow, reactivated Mesoproterozoic shear zones due to late Paleozoic compression and Mesozoic rifting (Gates, 1995; 1998) and thus, have a well preserved record of Grenville-age metamorphism and deformation. Rocks of the Hudson Highlands consist of a complex assemblage of metasedimentary, metavolcanic, and quartzofeldspathic ("granitic") gneiss, and intrusive granitoid rocks that were variably deformed and metamorphosed at upper amphibolite to hornblende-granulite facies conditions during at Grenville orogenesis (Dallmeyer and Dodd, 1971; Dallmeyer, 1974; Helenek and Mose, 1984). Based on field relations and a limited amount of recent U-Pb zircon ages (Helenek and Mose, 1984; Aleinikoff and Grauch, 1990; Ratcliffe, 1992; Ratcliffe and Aleinikoff, 2001; Gates et al. 2001a), the rocks of the Hudson Highlands can be roughly divided into two groups: (1) pre-Ottawan (>1090-1030 Ma) and (2) late- to post-Ottawan (<1030 Ma). Pre-Ottawan rocks all have strong, penetrative, high-grade metamorphic fabrics related to the Elzevirian and/or the Ottawaan orogeny.

Late- to post-Ottawan rocks are variably deformed, ranging from undeformed to those that locally have strong, high-grade, ductile fabrics but lack the regional-scale, penetrative fabrics that characterize the pre-Ottawan rocks. Locally, late- to post-Ottawan rocks truncate fabric elements in pre-Ottawan rocks.

The area of the field trip is located in parts of the Sloatsburg, Thiells, Monroe, Popolopen Lake, and West Point quadrangles west of the Hudson River within the west central Hudson Highlands, New York (Figs. 2 and 3). Previous mapping in this area, divided the units by rock types (Dodd, 1965; Jaffe and Jaffe, 1973; Dallmeyer, 1974; Helenek and Mose, 1984). Considering that about 80% of the rocks are quartz-feldspar gneisses, this system is useful for geologic maps but not for purposes of tectonic reconstructions. Gundersen (1986) suggested that lithologic and stratigraphic associations and sequences should be grouped as units. This system of mapping rock types is adopted for this field guide (Fig. 3).

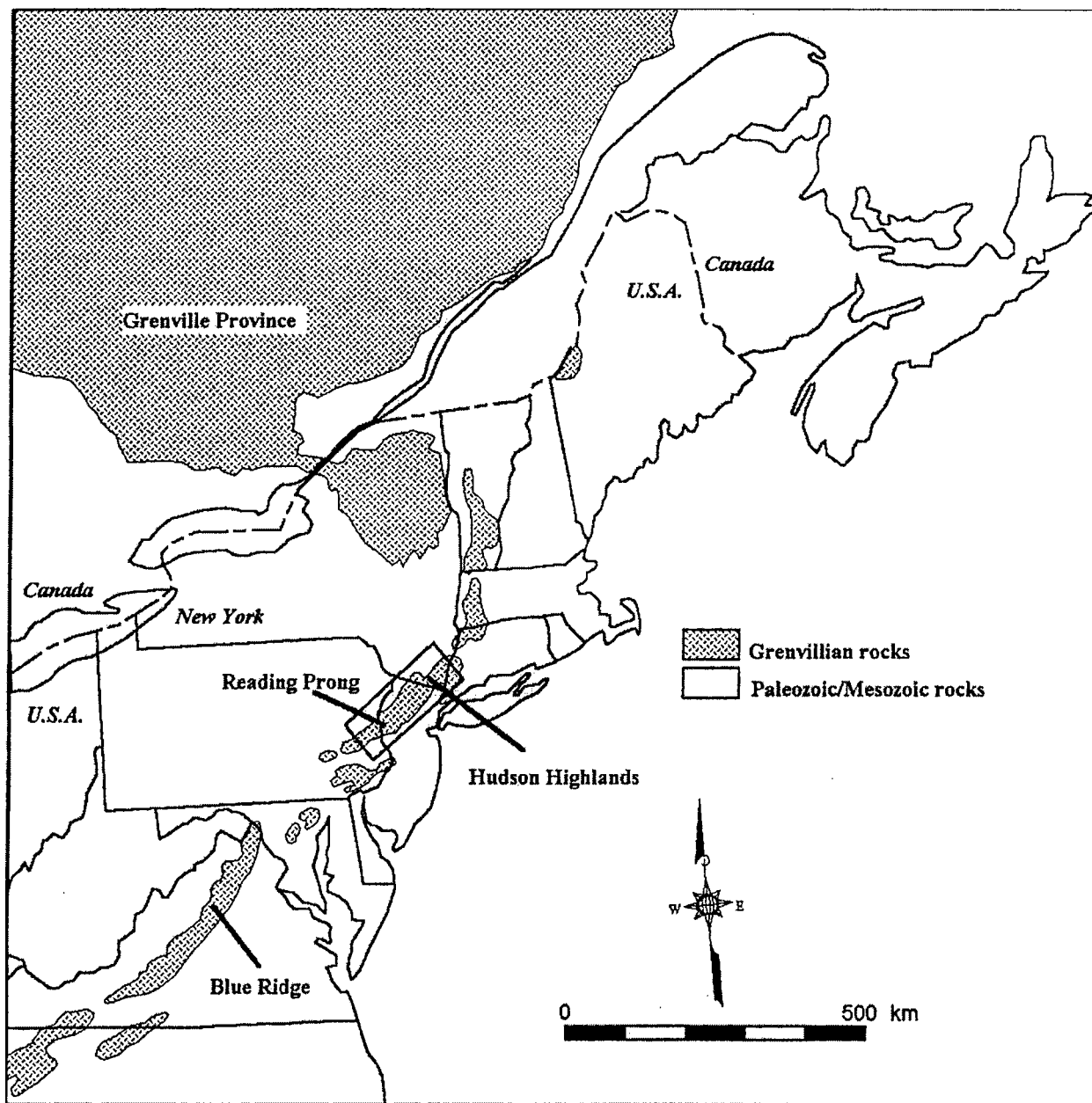


Figure 1. Regional map of eastern North America showing the geographic distribution of Grenville rocks. The area of Figure 2 is outlined by a rectangle. Map from Gates et al. (2001c).

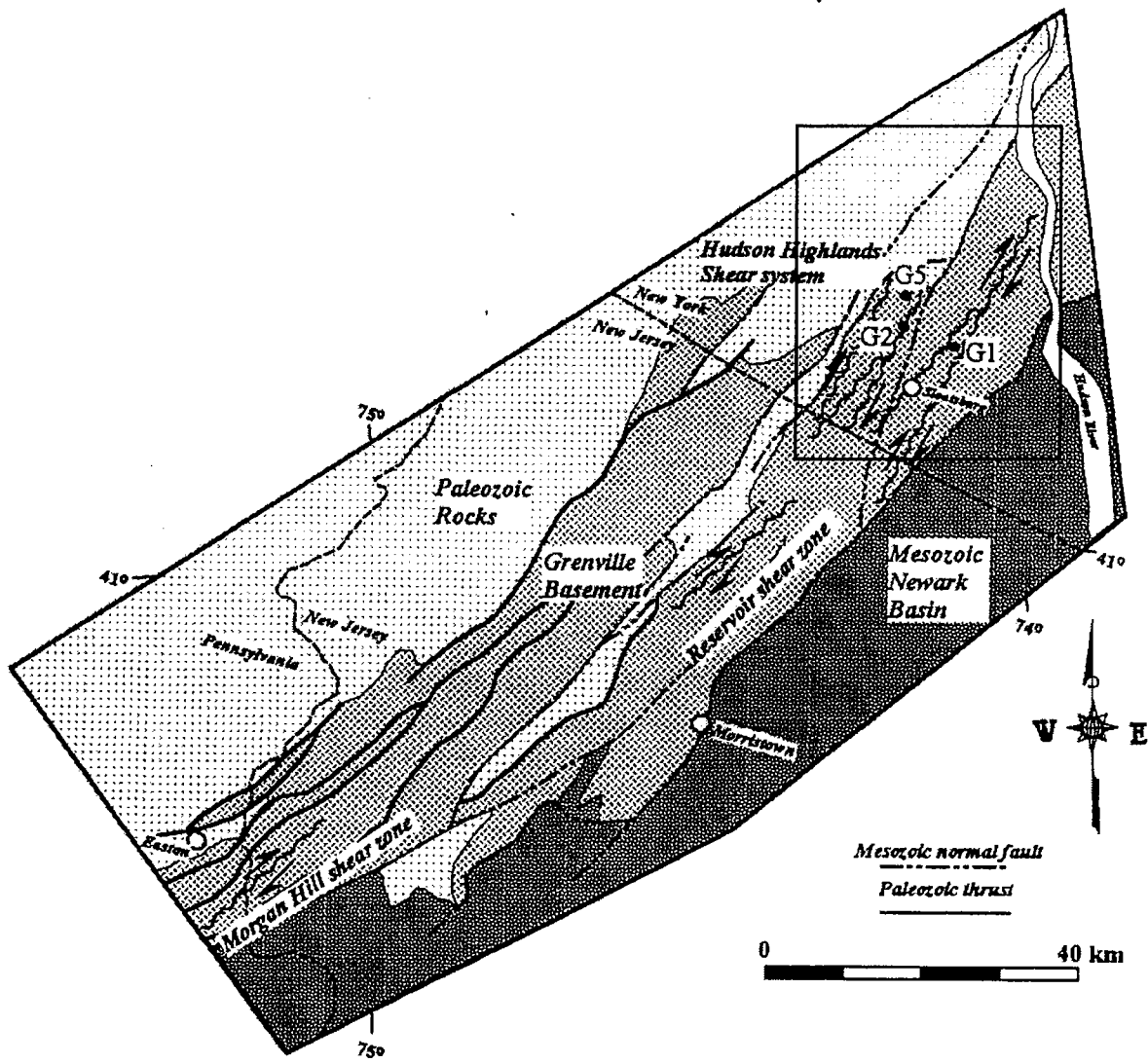


Figure 2. General geologic map of the Reading Prong and Hudson Highlands. The study area shown in Figure 3 is outlined by a box. G1, G2, and G5 are sample localities where Gates et al. (2001b) obtained SHRIMP U-Pb zircon ages. G1 is the Lake Tiorati Metadiorite at Stop #4 (see Fig. 6) and G2 is the sheared quartzofeldspathic gneiss at Stop #3 (see Fig. 4). Map modified from Gates et al. (2001c).

## PRE-OTTAWAN ROCKS

### *Metavolcanic Lithofacies*

Perhaps the oldest rocks in the region are a suite of quartzofeldspathic orthogneiss which include strongly banded, interlayered, very light colored, biotite- and/or hornblende-quartz-plagioclase gneiss, charnockitic (orthopyroxene-bearing) quartz-plagioclase gneiss, and amphibolites of mafic to intermediate compositions. Mafic, intermediate, and felsic compositional banding ranges in thickness from 5 cm to 5 m with varying proportions of each rock type. There are local interlayers of quartzite and calc-silicate gneiss. Migmatites also occur locally in this unit. These rocks are interpreted to represent a continental volcanic-plutonic arc suite of calc-alkaline rocks mixed with minor sediments (Ratcliffe, 1992; Gates et al., 2001c). Rocks equivalent to the metavolcanic lithofacies extends southwestward into the New Jersey Highlands where it is called the Losee Metamorphic Suite (Volkert and Drake, 1999) and northeastward into the eastern Hudson Highlands (Ratcliffe, 1992). Rocks of the metavolcanic

lithofacies are lithologically and chemically very similar to tonalitic and charnockitic gneisses found in the southern Adirondacks (McLelland and Chiarenzelli, 1990) and to the Mount Holly Complex in the Green Mountain Massif in Vermont (Ratcliffe et al., 1991), which have been dated at ~1350 to 1300 Ma.

### ***Metasedimentary Lithofacies***

Throughout the western Hudson Highlands there are belts of rock considered to have sedimentary protoliths including pelitic-, psammitic-, calcsilicate-gneisses, quartzite, and marble. Belts of rock upward of a few kilometers wide may contain all or some of these rock types, interlayered at the scale of meters to 100's of meters. These rocks have been included in the metasedimentary lithofacies that is portrayed on the geologic map (Fig. 3). The metapelite consists of interlayered biotite-garnet gneiss with medium to coarse quartz, plagioclase, K-feldspar and local sillimanite, and cordierite with quartzofeldspathic layers. Within the metapelite are zones of graphite-pyrite-garnet gneiss with biotite, quartz, K-feldspar, plagioclase, and sillimanite locally. Quartzite layers of 10-50 cm thickness also occur within this unit as do rare and discontinuous layers of diopside and diopside-garnet marble to calcsilicate of 10 cm to 2 m thickness. The calc-silicate is quartzofeldspathic with salite, apatite, sphene, scapolite, and hornblende, and is commonly migmatitic. Metaturbidite rocks are interlayered metapelite and metapsammite at the cm-scale, and are ubiquitously migmatites of metatexite (stromatic) and diatexite (structurally disrupted) types with variable degrees of disruption of the rock structure. These rocks are typically residual in mineral content suggesting melt loss.

Based on the abundance of graphite-sulfide rocks and the presence of minor interlayers of amphibolite of probably volcanic origin, Gates et al. (2001b) interprets this sequence as most likely a suite of continental- to oceanic-arc extensional basin deposits. Volkert and Drake (1999) came to a similar conclusion based on field relations and whole-rock geochemical data on correlative rocks in New Jersey Highlands. Similar packages of metasedimentary rocks are common in the Adirondacks (e.g., McLelland et al., 1996) and in the Central Metasedimentary Belt in Canadian Grenville (e.g., Rivers, 1997, Carr et al., 2000).

The contacts with the quartzofeldspathic gneiss (see below) and rocks of the metavolcanic lithofacies are usually gradational such that age relations based on field relations are ambiguous due to transposition of original stratigraphic and/or cross-cutting relations. However, in the New Jersey Highlands, Volkert and Drake (1999) have recognized field evidence that indicates that the correlative metasedimentary sequence unconformably overlies the equivalent of the metavolcanic lithofacies (Losee Metamorphic Suite). Demonstrably unconformable relations between these units in the Hudson Highlands has yet to be recognized in the study area of this field trip, however, Ratcliffe (1992) also places the equivalent of the metavolcanic lithofacies in the eastern Hudson Highlands at the base of the "stratigraphy", below metasedimentary rocks. Dodd (1965) and Helenek and Mose (1984) document a few localities where metasedimentary lithofacies rocks are crosscut by metaplutonic quartzofeldspathic gneisses (e.g. Storm King granite gneiss) in the Popolopen Lake quadrangle near Bear Mountain. Thus, the age of the metasedimentary lithofacies rocks is roughly constrained to be possibly younger (or at least contemporaneous) than the metavolcanic lithofacies (<1300-1350 Ma), but clearly older than quartzofeldspathic gneiss unit (see below) which has recently been dated at ~1175 Ma by SHRIMP U-Pb zircon techniques (Ratcliffe and Aleinikoff, 2001).

### ***Quartzofeldspathic Gneiss***

The quartzofeldspathic gneiss ranges from massive to layered quartz-plagioclase gneiss and quartz-K-feldspar-plagioclase gneiss with minor amounts of clinopyroxene, hypersthene, hornblende and/or biotite. Locally, this unit contains magnetite or garnet in trace amounts. Compositional layers are defined by the proportion and type of the ferromagnesian mineral component. Locally, this unit contains apparent textural gradation across the fabrics by an increase in the amount of mica and decrease in layer spacing with sharp contacts between, suggesting a relict sequence. However, such relict sequences in granulite terranes are difficult to interpret. Quartzofeldspathic gneiss is locally interlayered with quartzite and with mafic gneiss at the contact with the metavolcanic lithofacies. The gradational contacts with the metavolcanic and metasedimentary lithofacies, and the internal compositional layers suggest that parts of the quartzofeldspathic units could represent a volcanoclastic sequence.

However, in the more massive rocks, this unit is strongly lineated (L>S) defined by stretched hornblende prisms and rodded quartz-feldspar aggregates interspersed with large (2-4 cm) plagioclase and/or K-feldspar augen and locally contains mafic gneiss xenoliths. These features clearly support a metaplutonic origin for at least part of quartzofeldspathic gneiss. These more massive textured rocks are interpreted here to be correlative to hornblende

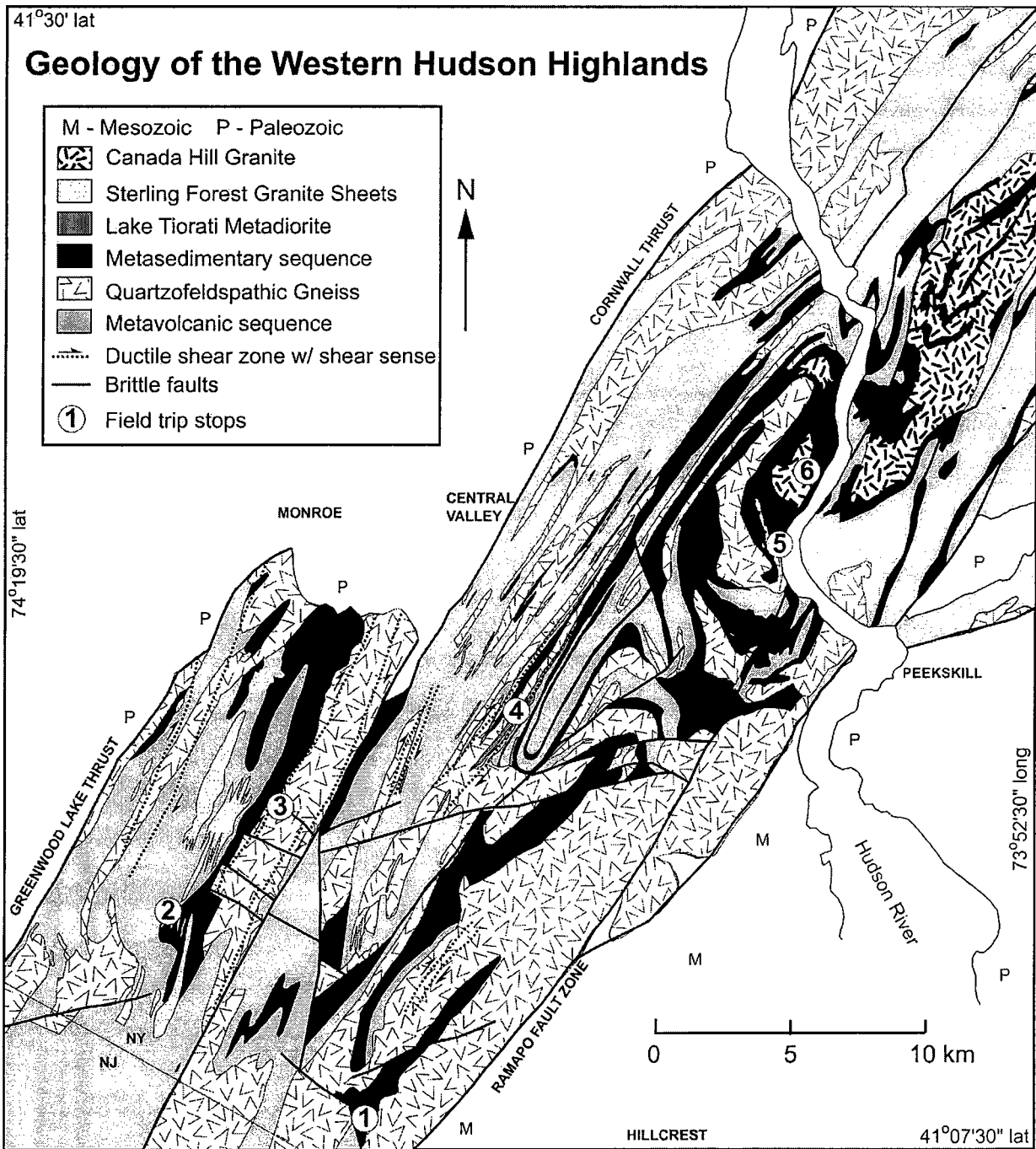


Figure 3. Generalized geologic map of the west central Hudson Highlands showing field trip stops 1-6. Based on geologic mapping of Dodd (1965), Offield (1967), Helenek (1971), Ratcliffe (1992), Valentino (2001); Gates et al., (2001c).

granite gneiss mapped in the Popolopen Lake (Dodd, 1965) and West Point quadrangles (Helenek and Mose, 1984), and Oscawana Lake quadrangles (Ratcliffe, 1992) which has been historically referred to as the Storm King granite (Berkey, 1907; Lowe, 1950). These rocks also correlate with similar metaplutonic hornblende granite gneiss lithologies within the Byram Intrusive Suite in the New Jersey Highlands (Volkert et al., 2000). Preliminary whole-rock geochemical data shows that massive-textured quartzofeldspathic gneiss, Storm King granite gneiss, and the Byram Intrusive Suite all have nearly identical geochemistry, consistent with a protolith that was a strongly A-type granitoid, which further supports their regional correlation (Gorring, et al. 2001; Verrengia and Gorring, 2002).

Current ages constraints on the crystallization age of these metaplutonic rocks based on recent SHRIMP U-Pb ages on igneous cores of zircons range from 1160 to 1230 Ma (Ratcliffe and Aleinikoff, 2001; Gates et al., 2001b) (Fig. 4). The quartzofeldspathic gneisses are chemically similar to other metaplutonic rocks of A-type chemical affinity with ages of 1170 to 1210 Ma from other parts of the Grenville orogen including southeastern Canada (Easton, 1986; Lumbers et al., 1990; Davidson 1995) and the Adirondack Lowlands (Wasteneys et al., 1999) as well as the slightly younger AMCG suite (~1145-1155 Ma) in the Adirondack Highlands (McLelland and Whitney, 1990).

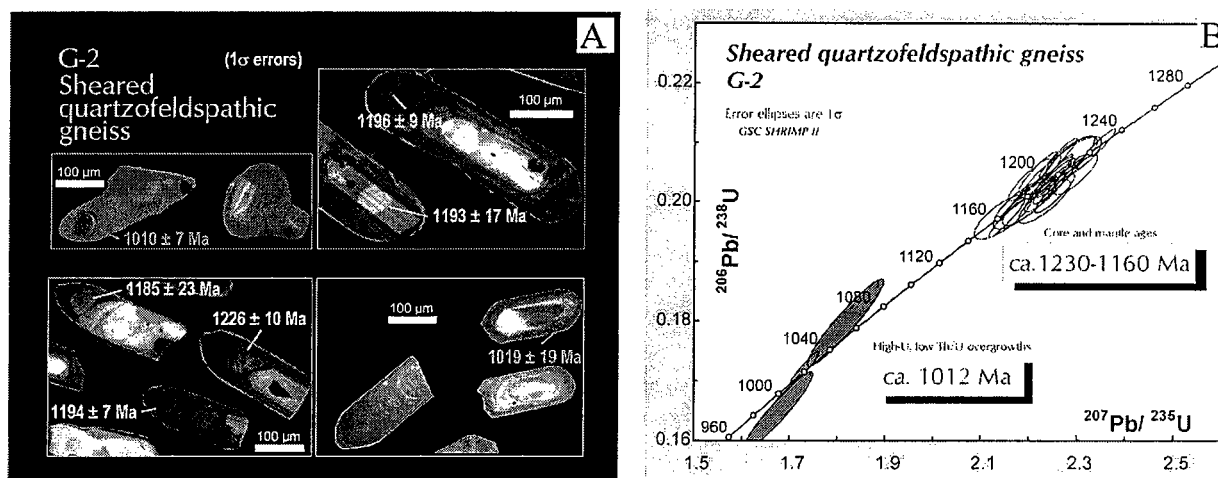


Figure 4. Representative cathodoluminescence images (A) and U-Pb concordia plot (B) from zircons extracted from sheared quartzofeldspathic gneiss within the Indian Hill Shear Zone at Stop 3 analyzed using the SHRIMP II instrument at the Geological Survey of Canada, Ottawa (Gates et al., 2001b).

## PRE- TO SYN-OTTAWAN METAMORPHISM AND DEFORMATION

There are at least two (and perhaps three) major deformational events recorded in the crystalline rocks of the Hudson Highlands. The dominant deformational structure of the older event(s) is a penetrative gneissosity that occurs in every unit except the Lake Tiorati Metadiorite, Sterling Forest Granite Sheets, Canada Hill Granite, and late pegmatites (see below). This gneissosity is defined by virtually all minerals but especially by platy and elongate minerals. Biotite, amphibole, sillimanite, and pyroxene are aligned in the strongly foliated quartz-feldspar matrix. Additionally, aggregates of quartz and feldspar define layering in some lithologies. Amphibole and pyroxene clots show similar rotation textures forming  $\delta$ -porphyroclasts (Passchier and Simpson, 1986). Some pelitic rocks contain garnet-fish structures, and locally, some rocks contain intrafolial asymmetric isoclinal folds 5 to 20 cm thick. The vergence of these folds is consistent in some areas and appears to indicate westward transport. Mesoscopic and megascopic folds produced during this event are recumbent to shallowly reclined. They are tight to isoclinal and commonly asymmetric with the lower limbs sheared out. This asymmetry consistently indicates northward transport. The weak and sparse kinematic indicators described above support this shear sense. Thinner layers in these folds contain mesoscopic parasitic folds that are especially well developed on the upper limb. Metamorphism associated with these structures is of hornblende granulite facies and maximum P-T estimates are on the order of 700-750°C and 4±1 kilobar based on mineral assemblages in metapelitic and mafic metavolcanic units (Dallmeyer and Dodd, 1971). Deformation and metamorphism associated with this event is most likely of Ottawa age and is interpreted to have been the result of a Himalayan-type continent-continent collision (Gates et al., in press). However, this does not preclude the possibility of pre-Ottawan deformation events (e.g., Elzeverian, Shawinigan as defined by Rivers, 1997) that could have been obliterated by the Ottawa event. In fact, this is likely the case. Dallmeyer, (1972), Helenek and Mose (1984) and Ratcliffe (1992) report structural evidence (e.g. refolded foliation) for multiple deformation events (e.g. refolded foliation) in the Bear Mountain area and in the eastern Hudson Highlands. Further evidence for a pre-Ottawan deformational history comes from a recent U-Pb SHRIMP zircon crystallization age of 1144 ±13 Ma on the Canopus Pluton in the eastern Highlands that lacks the older fabric elements (Ratcliffe and Aleinikoff, 2001). Structural and geochronologic evidence for multiple, penetrative fabric elements that can be assigned to distinct pre- and syn-Ottawan deformation events has yet to be recognized in the western Hudson Highlands.

## EMPLACEMENT OF LATE TO POST-OTTAWAN GRANITOIDS

Granitoids of late- to post-Ottawan age (<1030 Ma) are volumetrically minor in the Hudson Highlands compared to pre-Ottawan rocks, but are important for constraining the late geologic history of the area. They consist of a suite of leucogranite sheets in the Sterling Forest (here called the Sterling Forest granite sheets), Canada Hill Granite (1010 ± 4 Ma; Aleinikoff and Grauch, 1990), Lake Tiorati Diorite (1008 ± 4 Ma; Gates et al., 2001b) and a suite of late, crosscutting pegmatite dikes and mineralized zones (ca. 1000-925 Ma; Gates and Krol, 1998). The Mount Eve Granite (1020 ± 4 Ma; Drake et al., 1991, Gorrington et al., in press), located in the far western New Jersey and Hudson Highlands, is part of this suite of late- to post-Ottawan granitoids but will not be visited on this field trip. Similar plutonic granitoid activity of late- to post-Ottawan age has been documented elsewhere in the Adirondacks (ca. 1035 Ma; Lyonsdale Bridge pegmatite; 935 Ma Cathead Mt leucogranite; McLelland et al., 2001) and in the Green Mountain Massif (ca. 960 Ma; Stamford Hill rapakivi granite; Ratcliffe et al., 1991).

**Table 1: Representative Major and trace element analyses of Sterling Forest/Harriman State Park granitoids**

| Unit                               | Sterling Forest granite sheets |       |       |       |       | Lake Tiorati Metadiorite |       |       |       |        |       | Canada Hill Granite |                   |                     |                    |                    |                     |      |
|------------------------------------|--------------------------------|-------|-------|-------|-------|--------------------------|-------|-------|-------|--------|-------|---------------------|-------------------|---------------------|--------------------|--------------------|---------------------|------|
| Sample                             | SF-28                          | SF-29 | SF-31 | SF-35 | SF-36 | LT-1                     | LT-2  | LT-3  | LT-4  | LT-6a  | LT-6b | C-5 <sup>1</sup>    | 81-3 <sup>2</sup> | 13346b <sup>3</sup> | 13349 <sup>3</sup> | 13350 <sup>3</sup> | 13351c <sup>3</sup> |      |
| SiO <sub>2</sub> (wt%)             | 75.57                          | 76.01 | 74.97 | 75.25 | 74.83 | 50.66                    | 51.02 | 49.71 | 50.94 | 51.78  | 49.09 | 74.30               | 71.30             | 74.30               | 74.50              | 75.70              | 75.40               |      |
| TiO <sub>2</sub>                   | 0.04                           | 0.03  | 0.02  | 0.05  | 0.10  | 0.54                     | 0.49  | 0.53  | 0.80  | 0.57   | 0.64  | 0.23                | 0.14              | 0.07                | 0.03               | 0.02               | 0.02                |      |
| Al <sub>2</sub> O <sub>3</sub>     | 14.18                          | 13.73 | 13.41 | 13.55 | 13.57 | 15.63                    | 15.34 | 14.63 | 18.54 | 13.74  | 12.92 | 14.00               | 14.70             | 14.20               | 14.60              | 14.30              | 13.70               |      |
| Fe <sub>2</sub> O <sub>3</sub> (T) | 0.41                           | 0.51  | 0.17  | 0.47  | 1.18  | 9.09                     | 8.36  | 9.34  | 7.99  | 10.12  | 11.48 | 0.96                | 2.10              | 0.65                | 0.28               | 0.22               | 0.36                |      |
| MnO                                | 0.02                           |       |       |       | 0.02  | 0.14                     | 0.16  | 0.15  | 0.10  | 0.16   | 0.18  |                     | 0.05              | 0.02                | 0.02               | 0.02               | 0.02                |      |
| MgO                                | 0.10                           | 0.07  | 0.05  | 0.07  | 0.20  | 6.90                     | 7.40  | 7.91  | 5.42  | 8.59   | 9.90  | 0.29                | 0.57              | 0.28                | 0.16               | 0.12               | 0.10                |      |
| CaO                                | 1.06                           | 1.32  | 0.80  | 0.87  | 1.39  | 10.61                    | 12.27 | 12.41 | 10.00 | 11.77  | 11.86 | 1.08                | 0.86              | 0.95                | 0.81               | 1.57               | 1.13                |      |
| Na <sub>2</sub> O                  | 4.61                           | 3.64  | 2.73  | 3.53  | 3.43  | 3.66                     | 3.52  | 3.09  | 4.38  | 2.96   | 2.48  | 3.15                | 2.83              | 3.51                | 2.76               | 3.98               | 3.58                |      |
| K <sub>2</sub> O                   | 3.85                           | 3.94  | 7.48  | 5.34  | 4.89  | 1.39                     | 0.93  | 1.19  | 1.05  | 0.77   | 0.80  | 5.04                | 6.16              | 4.68                | 5.59               | 3.07               | 4.53                |      |
| P <sub>2</sub> O <sub>5</sub>      | 0.01                           | 0.01  | 0.02  | 0.02  | 0.02  | 0.03                     | 0.02  | 0.05  | 0.04  | 0.03   | 0.03  | 0.12                | 0.10              | 0.13                | 0.10               | 0.08               | 0.07                |      |
| Total                              | 99.86                          | 99.26 | 99.65 | 99.15 | 99.63 | 98.65                    | 99.51 | 99.01 | 99.26 | 100.48 | 99.38 | 99.17               | 98.81             | 98.79               | 98.85              | 99.08              | 98.91               |      |
| La (ppm)                           | 7.2                            | 7.5   | 9.2   | 5.9   | 35.0  | 14.7                     | 4.5   | 34.0  | 5.8   | 13.7   | 8.6   |                     |                   |                     |                    |                    | 51                  |      |
| Ce                                 | 15.5                           | 10.5  | 13.2  | 9.9   | 67.1  | 28.6                     | 9.6   | 54.1  | 15.3  | 34.3   | 22.5  |                     |                   |                     |                    |                    |                     | 110  |
| Pr                                 | 2.1                            | 1.3   | 1.6   | 1.3   | 7.3   | 3.5                      | 1.7   | 5.1   | 2.4   | 5.1    | 3.1   |                     |                   |                     |                    |                    |                     | 20   |
| Nd                                 | 7.0                            | 3.0   | 4.1   | 3.4   | 24.8  | 11.6                     | 5.8   | 15.5  | 10.3  | 21.0   | 12.4  |                     |                   |                     |                    |                    |                     | 54   |
| Sm                                 | 1.94                           | 0.35  | 0.54  | 0.64  | 4.19  | 2.31                     | 1.51  | 2.47  | 2.87  | 5.22   | 3.48  |                     |                   |                     |                    |                    |                     | <10  |
| Eu                                 | 0.23                           | 0.45  | 0.67  | 0.35  | 0.64  | 0.75                     | 0.57  | 0.62  | 0.81  | 1.03   | 0.86  |                     |                   |                     |                    |                    |                     | <2   |
| Gd                                 | 2.43                           | 0.51  | 0.63  | 1.20  | 3.94  | 2.72                     | 1.93  | 3.14  | 3.35  | 5.47   | 4.09  |                     |                   |                     |                    |                    |                     | 10   |
| Tb                                 | 0.56                           | 0.08  | 0.09  | 0.30  | 0.46  | 0.49                     | 0.39  | 0.50  | 0.65  | 1.00   | 0.79  |                     |                   |                     |                    |                    |                     | <20  |
| Dy                                 | 4.22                           |       |       | 2.32  | 1.82  | 3.16                     | 3.04  | 2.79  | 4.13  | 6.06   | 4.75  |                     |                   |                     |                    |                    |                     | 14   |
| Ho                                 | 1.08                           | 0.06  | 0.06  | 0.61  | 0.36  | 0.64                     | 0.54  | 0.60  | 0.89  | 1.40   | 1.19  |                     |                   |                     |                    |                    |                     | <4   |
| Er                                 | 3.48                           | 0.47  | 0.41  | 1.87  | 1.13  | 1.80                     | 1.51  | 1.72  | 2.28  | 3.44   | 3.02  |                     |                   |                     |                    |                    |                     | 8    |
| Tm                                 | 0.67                           | 0.07  | 0.05  | 0.33  | 0.15  | 0.27                     | 0.24  | 0.26  | 0.42  | 0.70   | 0.62  |                     |                   |                     |                    |                    |                     |      |
| Yb                                 | 5.66                           | 0.40  | 0.25  | 2.22  | 0.92  | 1.88                     | 1.53  | 1.59  | 2.41  | 4.30   | 4.03  |                     |                   |                     |                    |                    |                     | 7    |
| Lu                                 | 1.00                           | 0.06  | 0.03  | 0.36  | 0.15  | 0.30                     | 0.23  | 0.27  | 0.39  | 0.74   | 0.72  |                     |                   |                     |                    |                    |                     |      |
| Sr                                 | 21                             | 53    | 74    | 30    | 60    | 195                      | 150   | 158   | 181   | 147    | 106   |                     |                   |                     |                    |                    |                     | 150  |
| Ba                                 | 22                             | 113   | 334   | 97    | 224   | 247                      | 71    | 170   | 79    | 107    | 75    |                     |                   |                     |                    |                    |                     | 630  |
| Cs                                 | 0.15                           | 0.20  | 0.71  | 0.38  | 0.36  | 0.17                     | 0.15  | 0.22  | 0.09  | 0.04   |       |                     |                   |                     |                    |                    |                     |      |
| Rb                                 | 133                            | 74    | 139   | 125   | 163   | 27.5                     | 10.6  | 22.6  | 13.9  | 8.8    | 8.0   |                     |                   |                     |                    |                    |                     |      |
| U                                  | 4.88                           | 0.96  | 0.67  | 2.45  | 6.42  | 0.23                     | 0.41  | 0.14  | 0.14  | 0.26   | 0.35  |                     |                   |                     |                    |                    |                     | 7.79 |
| Th                                 | 3.12                           | 1.06  | 2.20  | 7.99  | 30.3  | 0.88                     | 0.81  | 1.42  | 1.42  | 0.83   | 0.95  |                     |                   |                     |                    |                    |                     | 41   |
| Y                                  | 33.7                           | 1.6   | 3.0   | 20.6  | 11.6  | 18.2                     | 14.4  | 16.5  | 24    | 40.2   | 34.7  |                     |                   |                     |                    |                    |                     | 48   |
| Zr                                 | 22                             | 50    | 19    |       | 134   | 29                       | 22    | 25    | 43    | 40     | 28    |                     |                   |                     |                    |                    |                     |      |
| Hf                                 | 0.29                           | 1.72  | 0.19  | 1.14  | 3.67  | 0.69                     | 0.73  | 0.58  | 1.17  | 0.98   | 0.75  |                     |                   |                     |                    |                    |                     |      |
| Nb                                 | 15.1                           | 0.6   | 0.2   | 0.5   | 0.4   | 1.7                      | 1.0   | 1.1   | 3.0   | 3.5    | 3.5   |                     |                   |                     |                    |                    |                     | 4    |
| Ta                                 | 0.98                           | 0.19  | 0.06  | 0.01  | 0.01  | 0.14                     | 0.13  | 0.09  | 0.38  | 0.33   | 0.36  |                     |                   |                     |                    |                    |                     |      |
| Sc                                 | 3.6                            | 0.4   | 2.2   | 5.4   | 1.9   | 45.0                     | 50.6  | 51.7  | 32.1  | 52.4   | 53.6  |                     |                   |                     |                    |                    |                     | 8    |
| Cr                                 | 1.8                            | 1.9   | 2.1   | 2.2   | 3.2   | 41                       | 91    | 135   | 198   | 137    | 133   |                     |                   |                     |                    |                    |                     | 12   |
| Ni                                 | 4.4                            | 2.5   | 2.2   | 1.4   | 2.3   | 56                       | 65    | 160   | 64    | 73     | 99    |                     |                   |                     |                    |                    |                     | 8    |
| Co                                 | 1.4                            | 1.2   | 0.4   | 0.8   | 1.2   | 34                       | 36    | 49    | 31    | 40     | 50    |                     |                   |                     |                    |                    |                     | 3    |
| V                                  | 5.3                            | 7.5   | 5.2   | 5.1   | 9.4   | 136                      |       | 220   | 271   | 190    | 208   |                     |                   |                     |                    |                    |                     | 12   |

Major elements, Sr, Ba, Zr, Y, and Sc by ICP-OES at Montclair State University; all other elements by ICP-MS at Binghamton University or INAA at Cornell University

<sup>1</sup> data from Helenek and Mose (1984); <sup>2</sup> Aleinikoff and Grauch (1990); and <sup>3</sup> Ratcliffe (1992)

### Sterling Forest Granite Sheets

Recent mapping in the Monroe and Sloatsburg quadrangles by Valentino et al. (2001) has identified a series of leucocratic granite sheets, primarily occurring west of the NY Thruway in the Sterling Forest section of Harriman State Park, that intrude pre-Ottawan rocks of the metavolanic and metasedimentary lithofacies and the quartzofeldspathic gneiss unit. Jaffe and Jaffe (1973) recognized these rocks in the Monroe quadrangle and mapped them as alaskite bodies. Offield (1967) also mapped isolated occurrences of this lithology immediately to the east in the Greenwood Lake quadrangle. The granite is typically medium to coarse grained, locally megacrystic, and lacks penetrative deformational fabrics. However, locally granite sheets are foliated where intersected by later

ductile shear zones and thus, clearly are slightly pre- or syn-kinematic to this event. Currently there are no radiometric age constraints on this important unit.

The granite sheets are leucocratic, with K-feldspar, quartz, plagioclase, only minor (<5%) hornblende and/or biotite, and accessory apatite, zircon, and titanite. The texture is equigranular with subhedral to anhedral interlocking grains, and locally they contain xenoliths of metasedimentary and metavolcanic lithofacies rocks. Where the granite sheets are mylonitic, the contact with the quartzofeldspathic gneiss is difficult to determine. The sheets range in thickness from 5 to 200 m and typically strike northeast-southwest and dip moderately to steeply to the southeast parallel to foliation in the surrounding gneiss. Most are laterally continuous for several kilometers. In the central part of Sterling Forest, the granite sheets extend out from two large tabular shaped bodies of granite. These two bodies of granite and sheet appendages occur within core of shallowly plunging, northeast trending, parallel, open antiformal structures that can be traced from the Monroe quadrangle southward.

The Sterling Forest granite sheets are high SiO<sub>2</sub> (~75%), leucocratic granites with <5% modal mafic minerals (Table 1; Fig. 5A). They are metaluminous to slightly peraluminous (ASI = 0.95 to 1.1) and have highly variable K<sub>2</sub>O/Na<sub>2</sub>O (0.3 to 3.3) reflecting variability in the modal abundance of K-feldspar or Na-plagioclase as the dominant feldspar. These rocks are divided into three chemically distinct groups based on REE patterns (Fig. 5B and C). The first group is characterized by LREE-enriched, HREE-depleted patterns with moderately negative Eu anomalies (Eu/Eu\* = 0.50 to 0.7) (Fig. 5B). The second group has a distinctive concave upward, "dished" MREE-depleted, HREE-enriched pattern with moderately negative to negligible Eu anomalies (Eu/Eu\* = 0.35 to 1) (Fig. 5B). The third group is defined by very low total REE's, strong LREE enrichment, depleted and flat MREE to HREE, extremely positive Eu anomalies (Eu/Eu\* up to 3.5) (Fig. 5C), and relatively high Sr and Ba concentrations (Table 1) relative to the other two groups. Group 1 granite sheets are best interpreted as partial melts of plagioclase-free source rocks with abundant residual amphibole ± garnet coupled with fractional crystallization of quartz + feldspars ± trace element-rich accessory phases (e.g., zircon, apatite, monazite, allanite). The garnet-bearing, plagioclase-free source mineralogy implies melt generation probably occurred at deep crustal levels probably involving source rocks of mafic to intermediate compositions. In comparison, Group 2 granite sheets clearly were generated by partial melting of garnet-free source rocks and hence melt generation probably occurred at shallower crustal levels. Group 3 granite sheets most likely represent rocks that accumulated feldspar, perhaps by some sort of filter pressing mechanism that extracted granitic melts during emplacement. These chemically distinctive groups of granite sheets have similar field relations and appear to be part of the same magmatic event, thus crustal melting apparently occurred at various crustal levels. On tectonic discrimination diagrams, the Sterling Forest granite sheets plot scattered along the boundary between fields for syn-collisional and volcanic arc granitoids (Figs. 5D-E).

### ***Canada Hill Granite***

The Canada Hill Granite (Berkey and Rice, 1919; Lowe, 1950; Helenek and Mose, 1984) is a distinctively white to blue-gray, coarse-grained leucogranite that occurs as small plutons, sheets, pods, and stringers almost exclusively within metapelitic gneisses of the metasedimentary lithofacies in the northeastern part of the Hudson Highlands (Fig. 3). The largest masses of this unit occur on the eastern side of the Hudson River in the vicinity of Canada Hill in the West Point quadrangle, where it was originally mapped and defined by Berkey and Rice (1919) and again formalized as a distinct, mappable unit by Helenek (1971). Lowe (1950) and Dodd (1965) also recognized this unit to the southwest in the Popolopen Lake quadrangle as well as by Ratcliffe (1992) to the east in the Oscawana Lake quadrangle. The Canada Hill granite is almost always associated with migmatitic host rocks.

The Canada Hill Granite is composed of quartz, white K-feldspar, and white to gray plagioclase in roughly equal proportions. Biotite is ubiquitous as the mafic phase with accessory amounts of sphene, apatite, and zircon. Garnet is locally abundant, especially near contacts with the enclosing migmatites and is interpreted to represent undigested xenocrysts derived from the metapelites. It is predominantly massive textured with only local development of a weak foliation. The orientation of the sheets and pods of Canada Hill Granite parallel the foliation in the surrounding metapelites and contacts are generally gradational and migmatitic, except locally where the granite clearly truncates foliation in the migmatitic metapelites. Aleinikoff and Grauch (1990) obtained a conventional, multigrain, TIMS U-Pb zircon crystallization age of 1010±6 Ma for the Canada Hill Granite and 1010±4 Ma for associated thin leucosome from surrounding migmatitic metapelitic gneiss from the same locality.

The Canada Hill Granite has long been interpreted as a late, synmetamorphic (anatectic) granite derived from partial melting of the surrounding metapelitic layers in the metaturbidites of the metasedimentary lithofacies (Lowe, 1950; Helenek and Mose, 1984; Ratcliffe, 1992). The intimate association of Canada Hill Granite and the



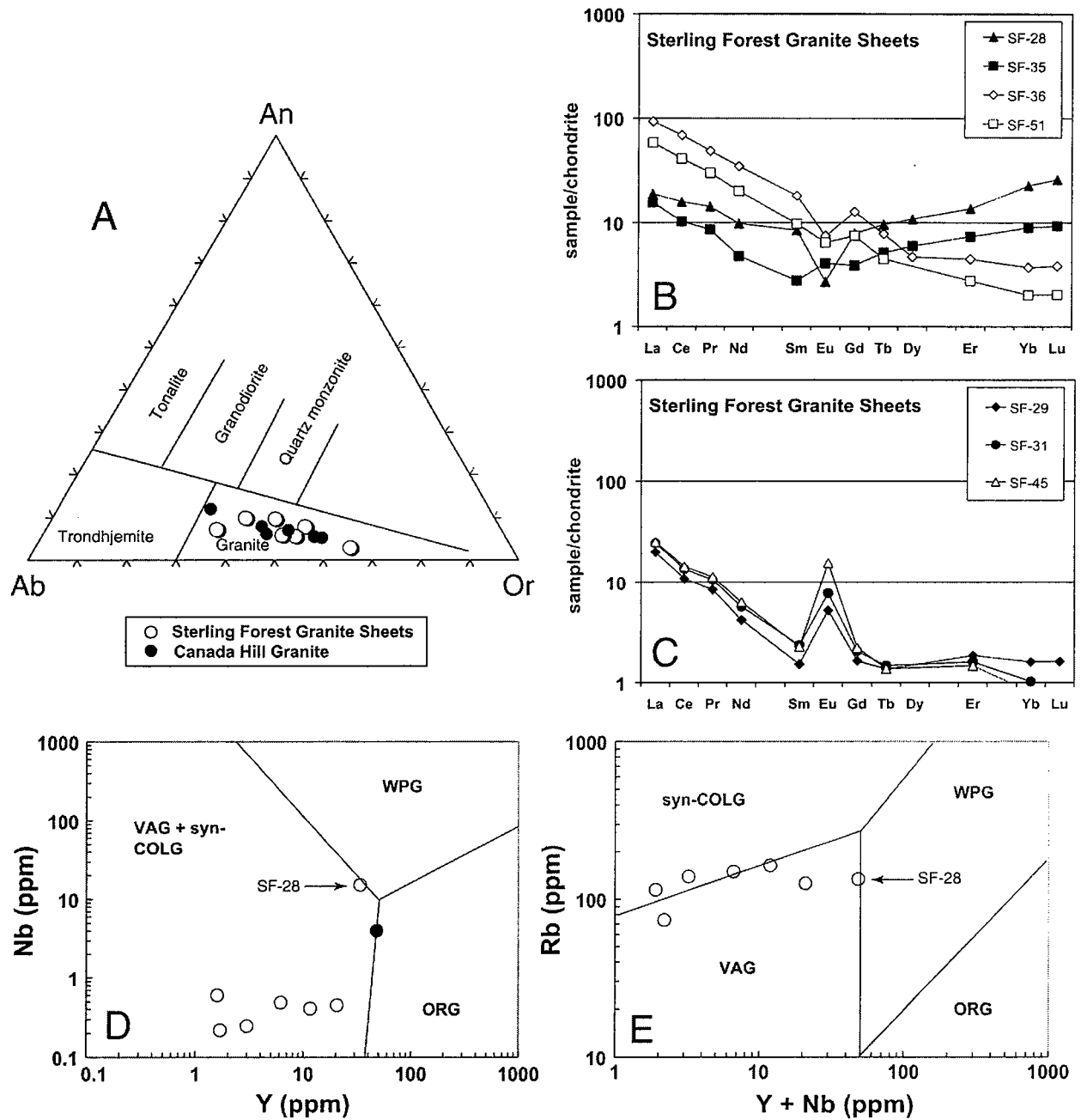


Figure 5. Geochemical plots for Sterling Forest granite sheets and the Canada Hill Granite. (A) Normative Ab-An-Or classification diagram (O'Connor, 1965). (B) and (C) REE plots showing three distinctive REE patterns within the Sterling Forest suite. Sample SF-28 is from field trip Stop #2. (D) and (E) are granitoid tectonic discrimination diagrams (Pearce et al., 1984). VAG = volcanic arc granitoid; syn-COLG = syn-collisional granitoid; WPG = within-plate granitoid; and ORG = ocean ridge granitoid. Chondrite normalization factors are from Masuda et al., (1973). Canada Hill data from Helenek and Mose (1984) and Ratcliffe (1992).

migmatites on a regional scale and the similarity of leucosome compositions and U-Pb zircon ages in migmatites with the Canada Hill support a petrogenetic link between them. Available geochemical data on the Canada Hill Granite consists of several analyses reported in Helenek and Mose, (1984), Aleinikoff and Grauch (1990); and Ratcliffe (1992) that are compiled in Table 1. The Canada Hill Granite is a high SiO<sub>2</sub> (71 to 75 wt%), very low

FeO<sub>T</sub> (generally <1 wt%), peraluminous, corundum-normative, biotite granite. These major-element chemical characteristics along with a high initial <sup>87</sup>Sr/<sup>86</sup>Sr ratio of 0.7186±0.0017 reported by Helenek and Mose (1984) provides additional support for a metapelitic source rock (e.g., S-type granite) for the Canada Hill Granite. We are currently conducting a more detailed geochemical study of the Canada Hill Granite that will elucidate more precisely the petrogenesis of this unit and its bearing on the late Ottawa history of the area.

### Lake Tiorati Metadiorite

The Lake Tiorati Metadiorite is a coarse- to very coarse-grained black and white speckled rock composed of sodic-plagioclase, pyroxene (both clinopyroxene and orthopyroxene), hornblende, and biotite locally. The type locality and location of the largest body (~0.5 km wide and ~5 km long) is along the western shores of Lake Tiorati in the central part of Harriman State Park within the Popolopen Lake quadrangle (Fig. 3) and was originally described by Lowe (1950) and mapped by Dodd (1965) as a coarse grained amphibolite (“Amphibolite II”). A few smaller, lens-shaped bodies occur a few kilometers to the southwest in the northeast corner of the Sloatsburg quadrangle recently mapped by Valentino et al. (2001). The metadiorite grades to lower pyroxene, gabbroic anorthosite compositions locally. Texture ranges from granoblastic to foliated and mylonitic with S-C fabric and rotated porphyroclasts. The metadiorite also locally contains xenoliths of mostly metasedimentary lithofacies rocks. Recent SHRIMP U-Pb dating of small, subhedral zircons with minimal zoning obtained from undeformed metadiorite from the type locality yielded a cluster of concordant ages averaging 1008 ± 4 Ma (Fig. 6). This is interpreted to be the crystallization age of the Lake Tiorati Metadiorite

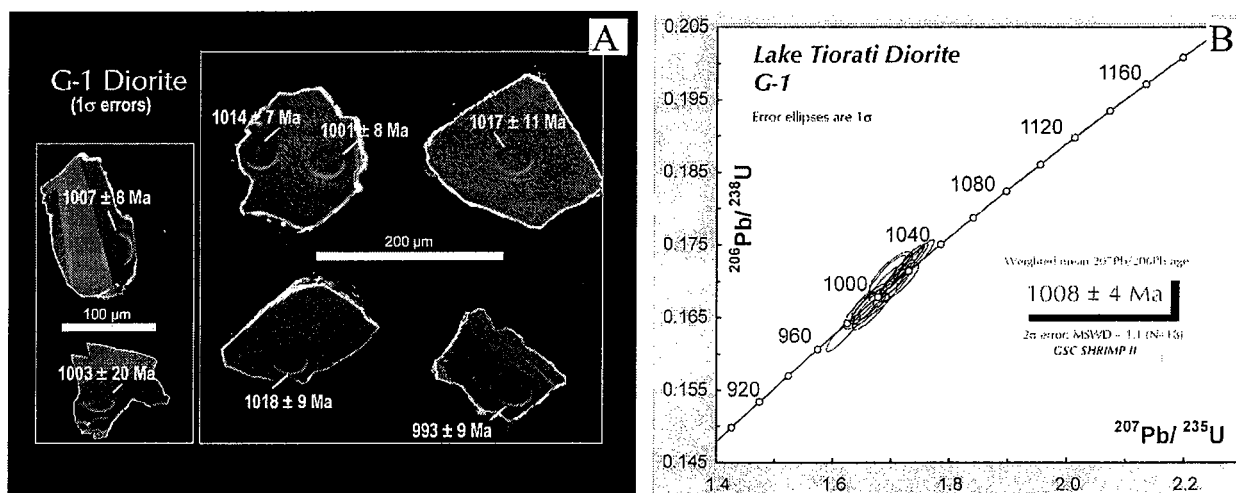


Figure 6. Representative cathodoluminescence images (A) and U-Pb concordia plot (B) from zircons extracted from the Lake Tiorati Metadiorite from Stop 4c analyzed using the SHRIMP II instrument at the Geological Survey of Canada, Ottawa (Gates et al., 2001b).

Major element chemistry of coarse-grained, relatively undeformed samples of the Lake Tiorati Metadiorite indicate they are uniformly mafic plutonic rocks that have moderate to strong calc-alkaline geochemical signatures (Table 1; Fig. 7A). REE patterns of most samples are weak to moderately LREE-enriched ( $La/Yb_N = 1.5$  to 5) and slightly concave upward or “dished”, MREE-depleted patterns with little to no HREE-depletion (Fig. 7B). They also have variable negative Eu anomalies ( $Eu/Eu^* = 0.6$  to 1.0). The mafic, calc-alkaline composition, relative strong negative Eu anomalies and slight MREE depletions in some samples suggests that significant plagioclase ± hornblende crystallization was important in the petrogenesis of these rocks before final emplacement. The lack of strong HREE and Y depletions relative to other trace elements indicates mantle melting occurred at relatively shallow depths above the garnet stability field (e.g. <65 km). All samples have very strong HFSE depletions and on plot well within volcanic arc fields on tectonic discrimination diagrams characteristic of calc-alkaline rocks associated with subduction zones (Figs. 7C). We interpret the arc signature in these rocks to have been inherited from lithospheric mantle sources that had been metasomatized by prior subduction events and/or extensive crustal contamination during emplacement in the crust.

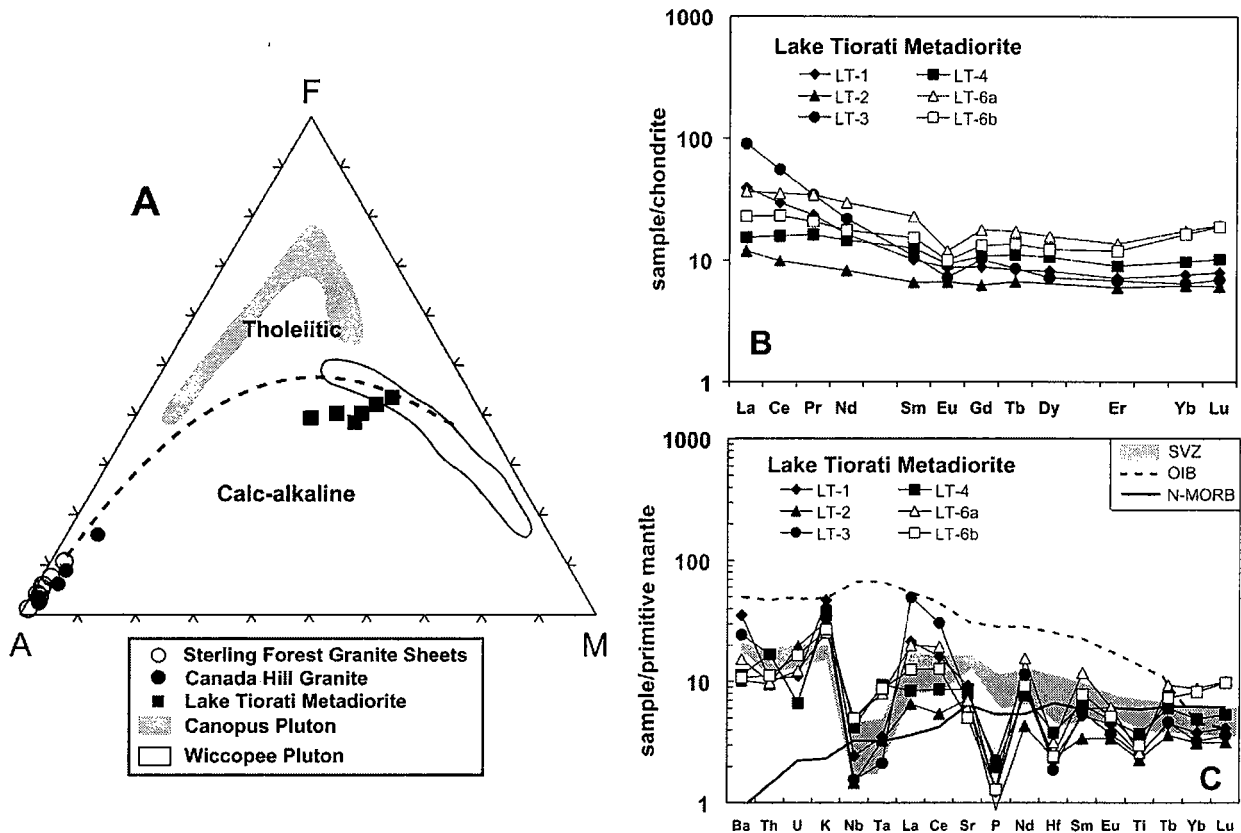


Figure 7. (A) AFM diagram showing the calc-alkaline affinity of the Lake Tiorati Metadiorite in comparison with other Mesoproterozoic mafic plutonic rocks (Canopus and Wicopee Plutons) from the eastern Hudson Highlands (Ratcliffe, 1992). (B) REE patterns and (C) multi-element diagrams showing the strong volcanic arc chemical signature (LREE-enriched, HFSE-depleted) of the Lake Tiorati Metadiorite. OIB is average oceanic island basalt; N-MORB is average normal mid-ocean ridge basalt; and SVZ is field for Quaternary mafic lavas from the Andean Southern Volcanic Zone (Hickey et al., 1986; López-Escobar et al., 1993). OIB, N-MORB, and normalizing factors are from Sun and McDonough (1989) and Masuda et al., (1973).

### Late Pegmatite Dikes

Gates et al. (2001b) recognized two generations of pegmatite dikes in this part of the Hudson Highlands. Early dikes are white and contain K-spar, quartz, muscovite, and garnet locally. They are largely parallel to subparallel gneissic foliation (concordant), commonly boudinaged, and contain internal foliation and deformed grains. Thickness ranges from 10cm to 1m. Many are associated with the Sterling Forest granite sheets and thus, are probably pre- or syn-deformational with the late ductile shearing event. The late dikes are pink, and very coarse grained with K-spar, quartz, and locally muscovite, hornblende, magnetite, pyroxene, titanite, and/or garnet depending upon the rock intruded. They are highly discordant, commonly within brittle faults, and contain xenoliths of fault rocks. They exhibit no deformational fabric. Thickness ranges from 1m to 10m. They are locally associated with small granite bodies. Dodd (1965), Offield (1967), Jaffe and Jaffe (1973), and Ratcliffe (1992) have also described similar late, cross-cutting pegmatites throughout the Hudson Highlands. Existing radiometric age dates for late pegmatite dikes include a hornblende  $^{40}\text{Ar}/^{39}\text{Ar}$  ages of  $923 \pm 2.8$  Ma reported by Gates and Krol (1998) and conventional TIMS U-Pb zircon ages of 964 to 1023 Ma obtained by (Aleinikoff et al., 1982).

## LATE TO POST-OTTAWAN STRUCTURES

Recent geologic mapping and structural analysis in both the New Jersey and Hudson Highlands has recognized the importance of a late to post-Ottawan, high-grade, ductile shearing event (Gates, 1998; Allers et al., 2001). Similar structures have been recognized recently in the central Adirondacks that may also be of late Ottawan age (Valentino et al., 2001; Solar et al., 2003). This deformation event ( $D_2$ ) is characterized by a group of anastomosing shear zones across the area (Figs. 2 and 3). These shear zones overprint the structural features associated with the first (and/or second) deformational event(s) ( $D_1$ ) that is characteristic of the pre-Ottawan rocks (metasedimentary and metavolcanic lithofacies and the quartzofeldspathic gneiss unit). The shear zones strike northeast and are either vertical or steeply northwest to southeast dipping. They range from 0.5 to 2 km in thickness though the boundaries are diffuse and difficult to determine in some areas. The shear zones are marked by well-developed type II S-C mylonite (Lister and Snoke, 1986) with shallowly northeast plunging mineral lineations. The dominant lithology within the mylonite is quartzofeldspathic gneiss but some rocks of the metavolcanic and metasedimentary lithofacies are also sheared. Metadiorite also locally forms S-C mylonite constraining the time of emplacement to pre-kinematic with regard to this event. Kinematic indicators within the mylonite include C/S fabric, rotated porphyroclasts, shear bands, asymmetric boudins and flattened asymmetric intrafolial folds. There are well-developed mesoscopic sheath folds with shallow northeast plunge, and megascopic drag folds adjacent to the main shear zone. All kinematic indicators show a consistent dextral strike-slip sense of shear. Minerals within the sheared rocks include amphibole and biotite as well as quartz and feldspar, all of which show plastic deformation but with full recovery. By texture and mechanical response of the minerals (Simpson, 1985), metamorphic conditions must have been upper amphibolite to granulite facies.

Adjacent to the zones are sheath folds and boudins within deformed layered sequences. They are strongly lineated and the structures are extended parallel to the lineations. Mesoscopic gentle to open upright folds also occur adjacent to the shear zones locally. These folds plunge gently from due north to north-northeast. The folds occur in well-layered metavolcanic sequences and within 150 m of the shear zone boundary. They locally appear en echelon. A pervasive steeply northwest-dipping crenulation cleavage occurs throughout the area. It is best developed in the metapelitic and thin layered metavolcanic units. Intersection lineations with the gneissic foliation produced in the early event are generally parallel to the stretching lineations in the mylonite. Late in the movement history, the shear zones became dilational. One to 5 km long mineralized veins occur along the shear zones. The veins parallel the zones but clearly cut the mylonitic foliation with ragged to planar contacts. The veins were progressively filled with salite and scapolite locally followed by magnetite. These zones are also commonly intruded by late pegmatites, some of which contain xenoliths of mineralized rock.

Timing of this ductile shearing event is constrained by recent SHRIMP U-Pb zircon ages on mylonitic rocks within and adjacent to the shear zones (Gates et al., 2001b). The Lake Tiorati Metadiorite is cut by the Fingerboard Mountain Shear Zone and has age of  $1008 \pm 4$  Ma (Fig. 6) and thus provides an upper age limit to the deformation in this area. SHRIMP data on zircon rims from the sample of sheared quartzofeldspathic gneiss from within the Indian Hill Shear Zone are not well clustered (range from 1000 to 1060 Ma), but the most concordant data yield ages around 1010 to 1020 Ma (Fig. 4). Late pegmatites that crosscut mylonitic fabric in these shear zones would provide lower age limits, unfortunately there are currently no age constraints on these specific occurrences. As mentioned above, similar late pegmatites have radiometric ages in the range of about 925 to 1000 Ma. Therefore, based on the geochronologic data presently available, this ductile shearing event is constrained to have occurred between ~1010 and 925 Ma, with the lower limit being relatively uncertain.

## IMPLICATIONS FOR GRENVILLE TECTONICS

One of the principal tectonomagmatic events in the Grenville orogen during the late Mesoproterozoic was the Ottawan Orogeny (~1090–1030 Ma; e.g., McLelland et al., 1996; 2001). The Ottawan Orogeny is thought to have been a Himalayan-style continental collision event with associated crustal thickening, high-grade metamorphism, ductile nappe-style folding in the southeast (e.g., in the Central Granulite Terrane, Adirondack Highlands, and Appalachian massifs) and brittle northwest-directed thrusting farther west (e.g., Grenville Front Tectonic Zone and Central Metasedimentary Belt) in the orogen. This collisional event is thought to be related to the accretion of various continental masses to form the supercontinent Rodinia (Hoffman, 1988; Dalziel, 1991; Borg and DePaolo, 1994). Although the timing of peak Ottawan orogenesis varies spatially, this event severely affected most rocks older than ~1060 Ma throughout much of the Grenville orogen (e.g., McLelland, 1996, 2001; Aleinikoff et al., 2000). The age and field relations of the Lake Tiorati Metadiorite and the Canada Hill Granite suggests that

penetrative deformation assigned to the Ottawa Orogeny as classically defined was finished prior to ~1010 Ma in the Hudson Highlands. Supporting evidence for this statement also comes from the undeformed Mount Eve Granite suite (Gorring et al., in press) which places a lower limit of ~1020 Ma for penetrative Ottawa metamorphism and deformation in the far western portion of the New Jersey/Hudson Highlands.

The presence of large-scale, vertical strike-slip ductile shear zones that cross-cut rocks of ~1010 Ma indicates that the late- to post-Ottawan history in the Hudson Highlands is characterized by a high-grade, dextral transpressional shearing event that occurred between ~1010 to 925 Ma. This event likely represents final adjustments of the amalgamated Rodinian supercontinent and/or another accretionary event that must have occurred far to the north of the Hudson Highlands. A collision in the area of the Canadian Appalachians and Scandinavia may have generated tectonic escape (e.g., Tapponnier et al., 1982; Burke and Sengor, 1986) of eastern Laurentia to the south along large dextral strike-slip faults (Fig. 8A). The strike-slip environment could explain the temporal association of the late- to post-Ottawan granitoid suite described here (Canada Hill Granite, Lake Tiorati Metadiorite, Mount Eve Granite). Collectively these granitoids consist of small, dispersed plutonic bodies that form a volumetrically minor, chemically diverse group that ranges from A- and S-type granites to calc-alkaline, I-type diorite. Localized crustal heating due to upwelling asthenosphere associated with localized extension and/or transtension in the overall dextral transpressional regime could explain the small volumes and limited areal extent of these granitoids (Fig. 8B). A similar transpressional tectonic model has been proposed by Speer et al. (1994) to explain the occurrence of small volume, chemically diverse plutonic suites of Alleghenian age in the southern Appalachians. This type of model may explain similar small volume occurrences of late- to post-Ottawan (ca 1030–930 Ma) granitoids elsewhere in the Grenville orogen, particularly in the Adirondacks (e.g., Lyonsdale Bridge and Cathhead Mountain pegmatites; McLelland et al., 2001) and the Green Mountain massif (e.g., Stamford Hill rapakivi granite; Ratcliffe et al., 1991).

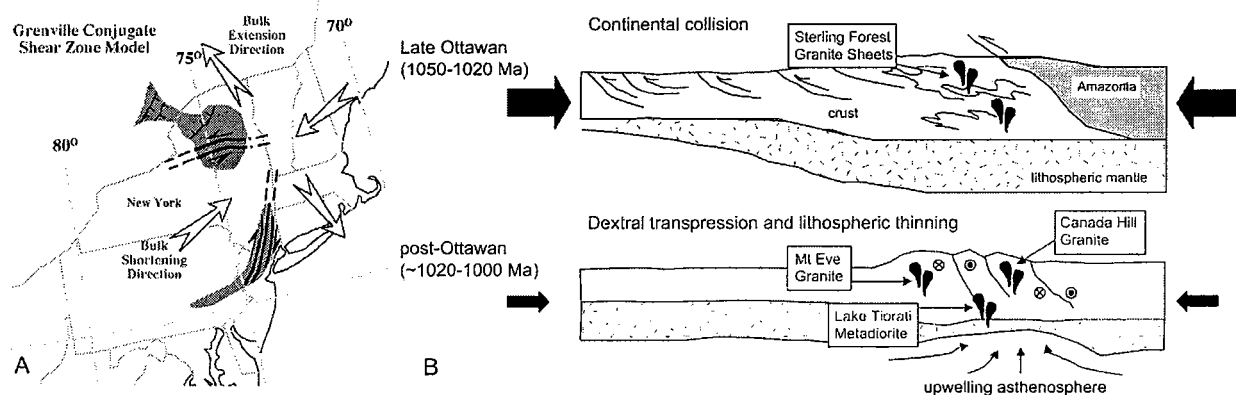


Figure 8. (A) Regional map view and (B) cross-sectional view of model for late to post-Ottawan tectonics in the Hudson Highlands. Diagrams modified from McLelland et al., (1996) and Gates et al., 2001a; in press.

## CONCLUSION

The late- to post-Ottawan (<1030 to 925 Ma) geologic history of the west central Hudson Highlands is characterized by large-scale, high-grade ductile shearing, migmatization, and the emplacement of a chemically diverse suite of granitoid plutons. The ductile shear zones are large (0.5 to 2 km wide, 2-10 km long), subvertical to vertical, and all kinematic indicators (e.g., S-C fabrics, rotated porphyroclasts) are consistent dextral strike-slip deformation. Timing of deformation on these shear zones is currently broadly constrained to an upper limit of ~1010 Ma based on the SHRIMP U-Pb zircon ages on the Lake Tiorati Metadiorite and on metamorphic rims on zircons from a sample of sheared quartzofeldspathic gneiss and a rough lower limit of 925 Ma based on hornblende  $^{40}\text{Ar}/^{39}\text{Ar}$  age obtained from an undeformed, late pegmatite. A suite of pre- and syn-deformational granitoid plutons with diverse geochemical characteristics we also emplaced associated with this shearing event. They consist of the Sterling Forest Granite Sheets, the Canada Hill Granite (1010 ± 4 Ma; Aleinikoff and Grauch, 1990), the Lake Tiorati Metadiorite (1008 ± 4 Ma; Gates et al., 2001b) and late, crosscutting pegmatite dikes and mineralized zones (ca. 1000-925 Ma; Gates and Krol, 1998). The Mount Eve Granite (1020 ± 4 Ma; Drake et al., 1991, Gorring et al., in press) is also part of this suite of late- to post-Ottawan granitoids. Dextral strike-slip shearing and granitoid

emplacement is interpreted here to have resulted from a tectonic escape mechanism due to late- to post-Ottawan adjustments within the newly amalgamated Rodinian supercontinent and/or a separate, late accretionary event to the north of the study area.

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## ROAD LOG

### MILEAGE

- 0.0 From Reeves Meadows Visitor Center parking lot, Harriman State Park, turn left out of the parking area onto Seven Lakes Drive.
- 1.5 Turn **LEFT** onto NY Rt. 17 at the T-intersection and traffic light in Sloatsburg. Continue on NY Rt. 17 through the town of Sloatsburg.
- 3.7 Stay in left lane and follow signs for NY Rte 59.
- 4.0 Continue straight ahead on NY Rt. 59 East. Large roadcuts off to the right (southbound side of NY Thruway) are hornblende granite gneiss and mafic to intermediate metavolcanic gneiss.
- 4.7 Turn **LEFT** on Torne Valley Road (County Road 95) and travel north. Large outcrops of hornblende granite gneiss on right (east) side of road (see Fig. 9A).
- 5.4 Pull over on left (west) side of road into gravel parking lot of the "H. Pierson Mapes/Flat Rock County Park" along the bank of the Ramapo River.

### STOP 1: RESIDUAL MIGMATITIC METATURBIDITE, DIATEXITE AND GRANITE SHEETS

**Stop 1a:** The first of two substops is a spectacular riverbed exposure of typical migmatitic gneiss found in the Hudson Highlands. **No hammers please.** Pavement exposure of migmatitic metaturbidite composed of steeply SE-dipping 5-10 cm-thick alternating layers of metapelite and metapsammite (Fig. 9B). Metapelitic layers are 3-5 cm thick with cm-scale, mesoscopically poikilitic garnet porphyroblasts draped by matrix minerals that form sinistral strain shadow tails defined by aggregates of Qtz+Bt+Pl+Sil. The mineral composition of the metapelite layers are consistently residual suggesting significant melt production and loss from these rocks.

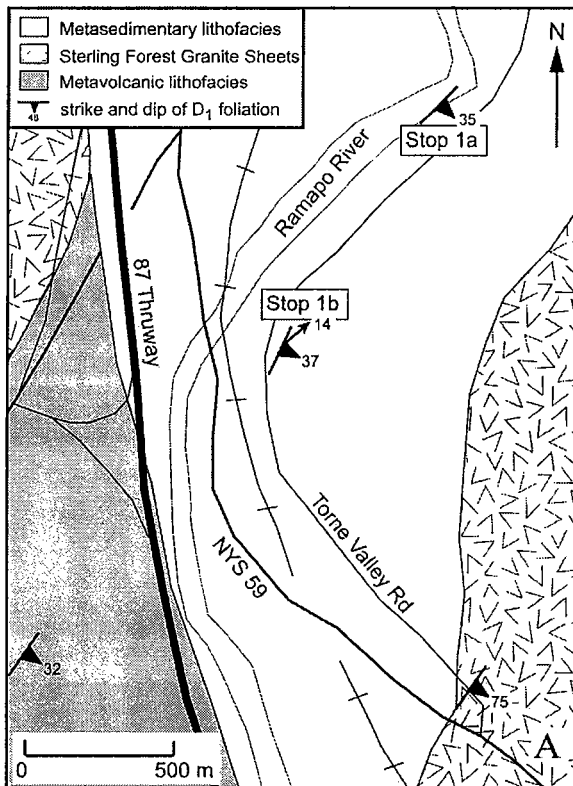


Figure 9. (A) Geologic map showing location of Stop 1a and 1b. (B) Pavement exposure of migmatitic metaturbidite along the Ramapo River at Stop 1a. D. Valentino's legs for scale.

**Stop 1b:** Walk south along road to pavement surfaces (~0.1 miles) and vertical roadcut (~0.2 miles) exposed on the left (east) side of the road (Fig. 9A). These outcrops are the along strike continuation of the migmatitic metaturbidite of Stop 1a. These rocks are similar to those of the stream pavement, except for the presence of residual diatexite (Grt+Bt+Pl+Qtz+Sil) and concordant- to weakly discordant cm-scale granite sheets (best seen in the

pavement exposures). Where present, diatexite shows local disruption of the protolith and stromatic migmatite structure (seen at Stop 1a) that apparently formed during syntectonic melting and melt movement within the diatexite. Note the local pinch-and-swell of granite sheets and the fabric boudinage of diatexite gneiss (quartzite in the interboudin partition). Residual diatexite is disrupted locally in which the diatexite has formed schollen with concordant granitic composition tails and Grt+Bt-rich melanosomatic drapes at the edges of their long dimension.

- 6.1 Return back down (south) on Torne Valley Road and turn **RIGHT** onto NY Rte 59 (Orange Ave.).
- 6.5 Continue traveling West on NY Rte 59 straight through traffic light and then follow signs for NY Rte 17 North.
- 8.1 Follow exit ramp right for County Road 72, follow Rt. 72 West.
- 11.2 Turn **RIGHT** at traffic light onto Long Meadow Road. You will see signs for Sterling Forest Park.
- 14.6 Small roadcut on left (west) side of metavolcanic amphibolite.
- 15.4 Old Forge Road enters at very sharp angle from left (west) side.
- 15.5 Pull over on right (east) side of road on decent gravel pad opposite roadcut of prominent pink granite.

## STOP 2: STERLING FOREST GRANITE SHEETS

This roadcut along northwest side of Long Meadow Road near the Sterling Forest Park headquarters exposes a fine example of the late- to post-Ottawan suite of pink, granite sheets found in the Hudson Highlands (Fig. 10). These

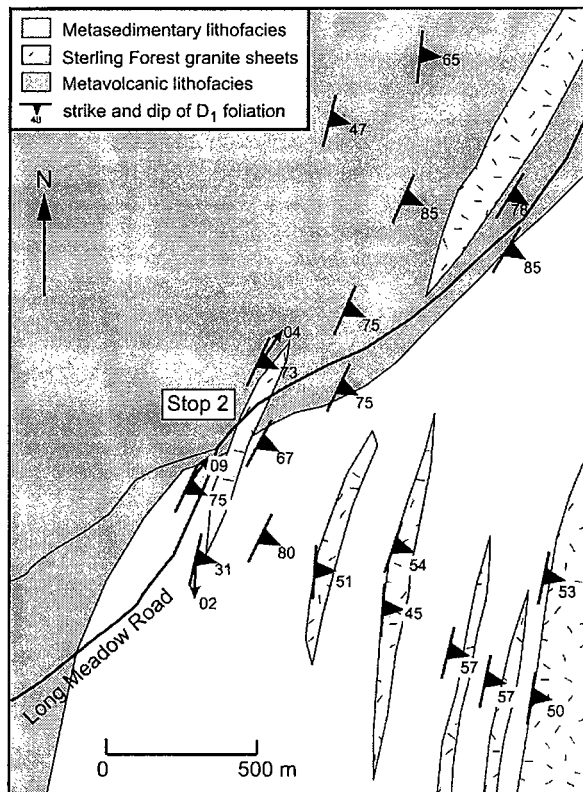


Figure 10. Geologic map showing location of Stop 2.

This roadcut along northwest side of Long Meadow Road near the Sterling Forest Park headquarters exposes a fine example of the late- to post-Ottawan suite of pink, granite sheets found in the Hudson Highlands (Fig. 10). These sheets are very leucocratic (~75 wt% SiO<sub>2</sub>), composed of almost entirely quartz, K-feldspar, and plagioclase, with only small amounts (<5%) of hornblende ± biotite with accessory apatite, titanite, and zircon. At this locality, the sheet has a medium-grained, equigranular igneous texture and shows no evidence of deformational fabric. However, locally these granite sheets are foliated where intersected by late-stage, ductile shear zones and thus, are clearly pre- or syntectonic with the shearing event. This particular sheet is one of several, 10-100 m thick, finger-like bodies that extend both southward and northward several kilometers from the main plutonic masses at Hogback Mountain (to the north) and Bill White Mountain (to the south). Here, the sheet is about 20 m thick, strikes N20E and dips ~65° to the southeast and has intruded parallel to foliation in the surrounding gneisses. It intrudes amphibolites and intermediate gneisses of the metavolcanic unit (contact observed at the north end of the outcrop) and metapelitic gneisses of the metasedimentary lithofacies (exposed across the road a short distance into the woods on the other side of the road). The granite sheet exposed here (sample SF-28) has a distinctly “dished” MREE pattern, a moderately strong negative Eu anomaly (Eu/Eu\* = 0.35), high HREE contents (~20x chondritic) and very low Ba and Sr concentrations (~20 ppm) (Fig. 5B and Table 1). These chemical characteristics indicate that amphiboles and feldspars were important phases in the petrogenesis of this particular granite. The data is consistent with partial melting of garnet-free, mafic amphibolitic source rocks and/or fractional crystallization of amphibole + plagioclase ± K-feldspar at shallower crustal levels.

- 19.1 **RIGHT** onto NY Rte 17A heading east (two-lane divided highway).
- 19.2 Large roadcuts of calc-silicate gneisses on opposite (north) side of 17A.
- 19.6 Make U-turn at Sylvan Way and go back on 17A West. Roadcut on right (north) side and in median is rocks seen at Stop 3.

19.9 Drive past west end of roadcut ~0.1 mile and pull over on wide shoulder. **BE CAREFUL!! Fast moving traffic and limited sight distance.**

### STOP 3: INDIAN HILL SHEAR ZONE

Roadcuts on north side and in median of NY Rte 17A afford an excellent view of highly sheared quartzfeldspathic gneisses within the Indian Hill Shear Zone (Fig. 11). The Indian Hill Shear Zone is one of several deformation zones in an anastomosing system of ductile shear zones that occur in the Hudson Highlands (Fig. 2). Timing of ductile deformation on the Indian Hill Shear Zone is currently constrained (albeit loosely) by recent SHRIMP U-Pb zircon ages from one sample of sheared quartzofeldspathic gneiss within the Indian Hill Shear Zone. The sample yielded zircons with rhythmically zoned igneous cores and clear, unzoned metamorphic rims (Fig. 4a). Analyses of the cores and rims produced two clusters of concordant ages (Fig. 4b). The cores ranged from 1160 to 1230 Ma. The most concordant rims ranged from 1000 to 1060 Ma, yielding an imprecise average of ~1010-1020 Ma. We interpret the zoned cores and associated ages to represent the original igneous history for this rock, and the rims to represent the regional metamorphic overprint, which includes high-grade ductile shearing on the Indian Hill Shear Zone. More rigorous constraints on the timing of ductile shearing in this area await additional U-Pb SHRIMP zircon geochronology on rocks within the Indian Hill Shear Zone as well as the Sterling Forest Granite sheets (upper limit), and cross-cutting, late pegmatites (lower limit).

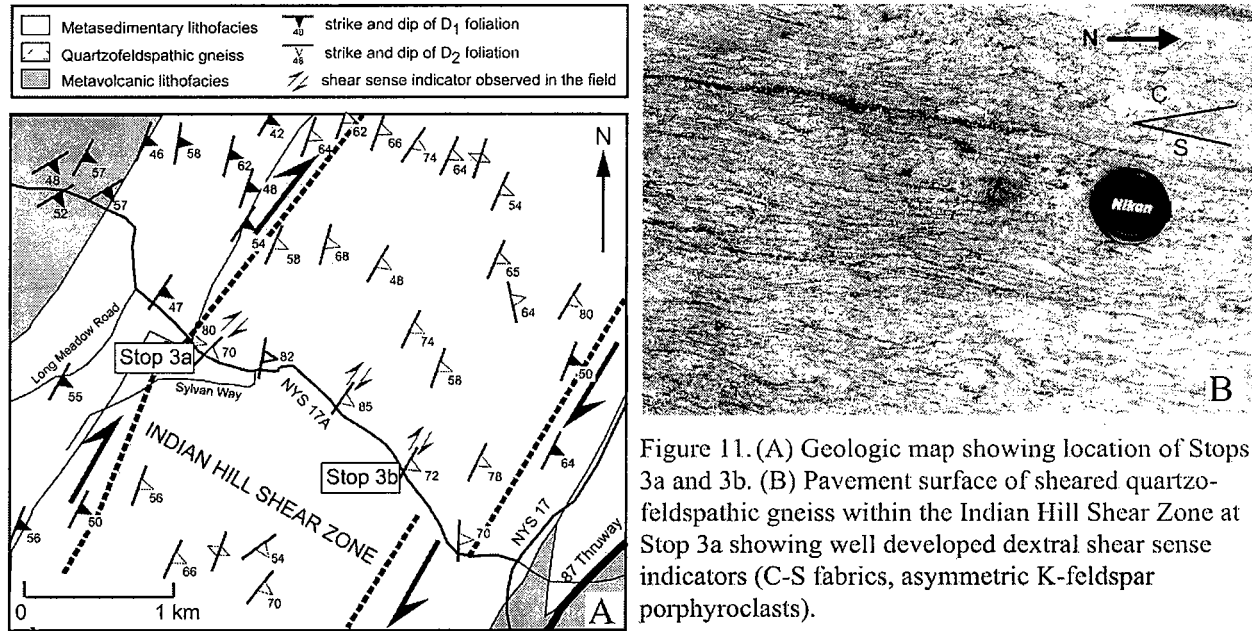


Figure 11. (A) Geologic map showing location of Stops 3a and 3b. (B) Pavement surface of sheared quartzofeldspathic gneiss within the Indian Hill Shear Zone at Stop 3a showing well developed dextral shear sense indicators (C-S fabrics, asymmetric K-feldspar porphyroclasts).

**Stop 3a:** Spectacular ductile shear zone structures are found on the pavement surface on top part of the outcrop in the median strip. The rock is a quartzofeldspathic mylonitic gneiss with thin (10-50 cm) interlayers of amphibolite and garnet-biotite-quartz-plagioclase gneiss locally. The quartzofeldspathic mylonite is well foliated and lineated and composed of plagioclase, quartz, K-feldspar, and biotite. This mylonite exhibits well-developed kinematic indicators including S-C fabric, reverse shear cleavage (RSC), rotated porphyroclasts, and shear bands. These kinematic indicators show a consistent dextral shear sense. The width of the zone and low S-C angle indicate significant offset. Locally there are small sinistral shear zones that crosscut the main foliation and are interpreted to be conjugate. The mylonitic foliation is commonly folded into open to tight shallowly northeast-plunging asymmetric folds. There are pegmatite dikes that are parallel to mylonitic foliation and which commonly displays pinch and swell. There are also late pegmatites that form in "gaps" in the mylonite. Cutting the mylonite at this locality is a prominent late pegmatitic dike (~2-3 m thick) composed of coarse-grained granite with large crystals of hornblende that form radiating and linear aggregates. It exhibits no deformation fabric and clearly postdates the ductile deformation of the quartzofeldspathic gneisses. Unfortunately, there are no radiometric age constraints for the pegmatite.

- 20.2 Proceed west on NY Rte. 17A and immediately get in left lane. **Make U-turn** at Long Meadow Road and head east on 17A. Small roadcuts of mylonitic quartzofeldspathic gneiss within the Indian Hill Shear Zone are seen along the north side of the road east of Stop 3b.
- 21.6 Carefully make **U-turn** at the bottom of hill at intersection with NY Rte. 17. Go back up hill on NY Rte 17A West.
- 21.8 Roadcut of quartzofeldspathic gneiss.
- 22.1 Pull over on shoulder of highway just past (north) of roadcut.

**Stop 3b:** This small roadcut again displays highly-sheared quartzofeldspathic gneisses and thin (10-50 cm) layers of amphibolite and garnet-biotite-quartz-plagioclase gneiss on both pavement surfaces and vertical cuts.

- 22.5 Continuing west on NY Rte 17A, make a **U-turn** at Sylvan Way and then proceed back on 17A east.
- 23.4 Intersection with NY Rte 17 again, this time continue straight on through intersection. NY Rte 17A becomes County Road 106 and enter Harriman State Park.
- 28.5 Enter Kanawake Circle and go ¾ around and follow Seven Lakes Drive toward Lake Tiorati and Bear Mountain.
- 30.4 Entrance to Camp Thendara on right. Pull over on right side as much as possible (careful of large rocks that line the road) (see Fig. 12).

**Note:** If you do this trip yourself, you will have to park about 1 mile to the north at Lake Tiorati Circle or gain permission from Harriman State Park to park along the road.

Walk north ~0.2 miles along Seven Lakes Drive to the blue blaze trailhead on the left (west) side (Fig. 12). Follow the blue trail up the east flank of Fingerboard Mountain for ~0.5 mile to the lean-to shelter near the top. Stop for lunch and enjoy the spectacular mylonite outcrops!!

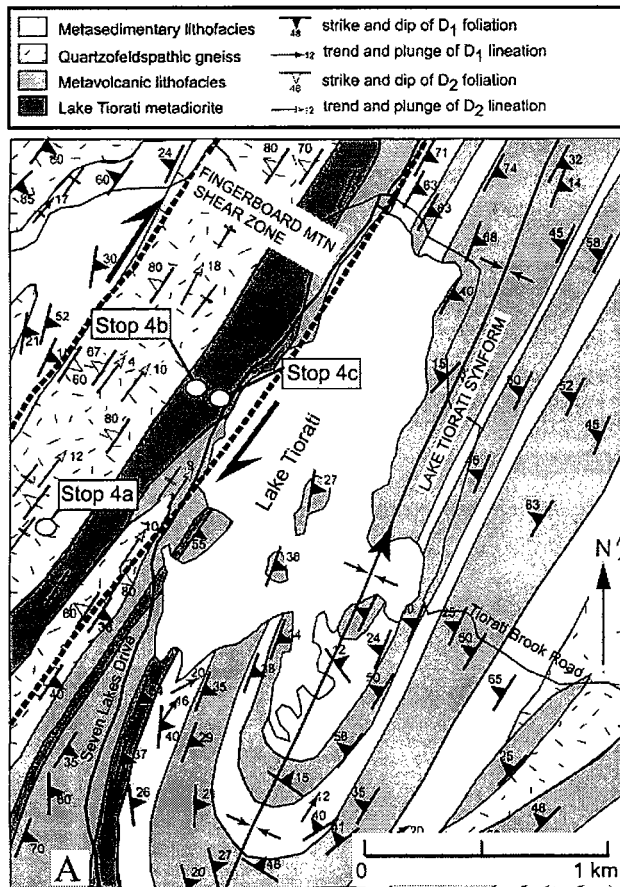
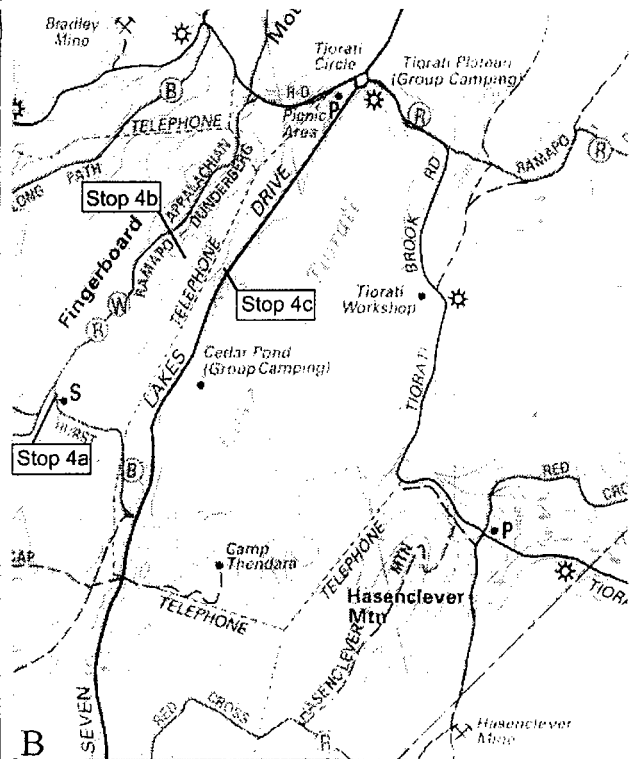


Figure 12. (A) Geologic and (B) topographic map showing the locations of Stops 4a-c. Geologic map modified from Dodd, (1965).



#### STOP 4: FINGERBOARD MTN SHEAR ZONE AND LAKE TIORATI METADIORITE

The next series of substops will require a moderate walk (~ 2 miles and 300 vertical feet) along hiking trails and finally through some open forest. We will examine another spectacularly exposed ductile shear zone within quartzofeldspathic gneiss along the Appalachian Trail on the crest of Fingerboard Mountain and then look at field relations and chemistry of a sheared mafic metaplutonic rock along the shore of Lake Tiorati. Bring food and water, as we will stop near the top of Fingerboard Mountain to eat lunch.

**Stop 4a:** Here at the shelter, and for the next ½ mile northward along the Appalachian Trail on the crest of Fingerboard Mountain, a series of nearly continuous pavement surfaces exposes mylonites of another 1- to 2-m wide, 10- to 20-km long ductile shear zone. The rock is quartzofeldspathic mylonitic gneiss with mineralogy and structure very similar to that observed in the Indian Hill Shear Zone at Stop 3. Foliation is either vertical or very steeply dipping to the northwest or southeast with shallow plunging (<20° NE) lineation indicating dominantly strike-slip ductile shearing (Fig. 12A). Like the Indian Hill Shear Zone at Stop 3, kinematic indicators (e.g., S-C fabric, rotated porphyroclasts) show consistent dextral shear sense and the very low S-C angle indicates significant offset.

Follow the blue trail above the shelter to the junction with Appalachian Trail. Turn right (north) and walk along the crest of Fingerboard Mountain for about ~0.5 miles (rocks of Stop 4a above are here). Descend about 80 vertical feet into small col and then make a right (east) turn into the woods following the drainage for ~0.1 mile. Ascend slightly up the north flank of the drainage through a series of closely spaced outcrops near the contact of quartzofeldspathic mylonite and amphibolites of the metavolcanic unit (not well exposed). Traverse northeastward for ~0.2 miles along the upper eastern flank of Fingerboard Mountain and stop and examine one of the larger exposures of sheared metadiorite.

**Stop 4b:** This outcrop is still within the Fingerboard Mountain Shear Zone, but we have crossed over the contact with the quartzofeldspathic gneiss and into a distinctively coarse-grained, sheared mafic metaplutonic rock, here referred to as the Lake Tiorati Metadiorite (after Gates et al., 2001c). This particular outcrop is part of a relatively large body that is ~200-300 m thick and is ~5-6 km in length located on the west side of Lake Tiorati (originally mapped by Dodd, 1965). The mineralogy of the rock at this locality is essentially all hornblende and plagioclase, with minor clinopyroxene and biotite. Foliation, defined by compositional layering, strikes northeast and dips steeply (~65°) to the northwest. Interestingly, mineral lineations plunge steeply to the northwest, indicating local thrust or normal motions on this part of the shear zone. The metadiorite is also cut by an undeformed granite pegmatite on the top of the outcrop.

Walk east-southeast, directly downhill toward Lake Tiorati for about ~0.2 miles. Reach the base of the hill and Seven Lakes Drive and then walk a few hundred meters northward to a small roadcut marked by a stick placed into a drill hole near the top of the outcrop.

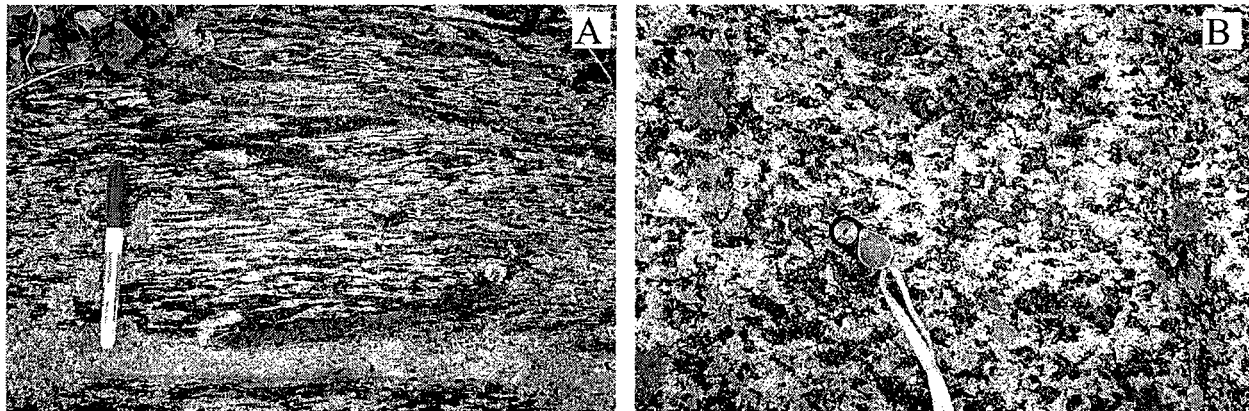


Figure 13. (A) Pavement surface of sheared Lake Tiorati Metadiorite from the summit of Blackrock Mountain ~5 km southeast of Stop 4c. Pen is oriented NW-SE. (B) Undeformed metadiorite from Stop 4c.

**Stop 4c:** This outcrop is the type locality for the Lake Tiorati Metadiorite and is located near the eastern margin of the Fingerboard Mountain Shear Zone. It is part of the same large body as mentioned in Stop 4b (Fig. 12A). At the southern end of the outcrop, the rock is essentially undeformed with good igneous textures developed (Fig. 13B). The diorite is composed mostly of plagioclase, hornblende, and clinopyroxene with minor orthopyroxene, magnetite and ilmenite. The orthopyroxene occurs as brown cores surrounded by coronas of clinopyroxene and/or hornblende. There is also a large xenolith of well-foliated biotite-plagioclase-quartz (metasedimentary) gneiss in the upper part of the outcrop. In the central and northern parts of the exposure, the diorite is cut by several anastomosing mylonite bands. The mylonite strikes northeast and is near vertical. Lineations plunge shallowly to the northeast. Kinematic indicators include rotated porphyroclasts and S-C fabrics (Fig. 13A). Where it can be determined, shear sense is consistently dextral. Recent SHRIMP U-Pb dating of small, subhedral zircons with minimal zoning obtained from undeformed metadiorite from this outcrop yielded a cluster of concordant ages averaging  $1008 \pm 4$  Ma (Fig. 6). Since this body is clearly cut by right-lateral ductile shear zones, this age provides an upper limit on the ductile deformation event that produced the Fingerboard Mountain Shear Zone.

Whole-rock major and trace element chemistry of samples from this outcrop indicate a mafic (~50% SiO<sub>2</sub>, see Table 1) calc-alkaline, arc-like affinity for these rocks (Figs. 7A and C). We interpret the arc signature in these rocks to have been inherited from the continental lithosphere during magma generation and emplacement, and thus they do not indicate that there was active subduction zone in this part of the Grenville at the time. Therefore, based on field relations, geochronology, and geochemistry, the Lake Tiorati Metadiorite is best interpreted as a syntectonic, mafic plutonic rock that was emplaced just prior to or synchronously with a major right-lateral, ductile shearing event.

- 32.0 Proceed ½ of the way around Tiorati Circle, continue to follow Seven Lakes Drive and signs for Bear Mountain.
- 35.6 Again proceed ½ of the way around major traffic circle and follow NY Rte 6 East.
- 36.2 Large roadcuts of metapelitic gneisses on right (southeast) side of highway.
- 37.6 Very large roadcuts on the right (southern) side of highway of classic Storm King Granite gneiss on the northern flank of the Bear Mountain pluton. Nearly continuous outcrop for next 0.5 miles to the east.
- 38.6 Bear Mountain Circle. Proceed ¼ of the way around circle, following signs for NY Rte 9W South and US 202 West. Immediately turn right on Hessian Drive. Park in nearest lots for park lodges. Walk north ~0.1 mile, through middle of traffic circle, to prominent, but low, roadcut on the northern side of the circle and on the east side of NY Rte 9W.

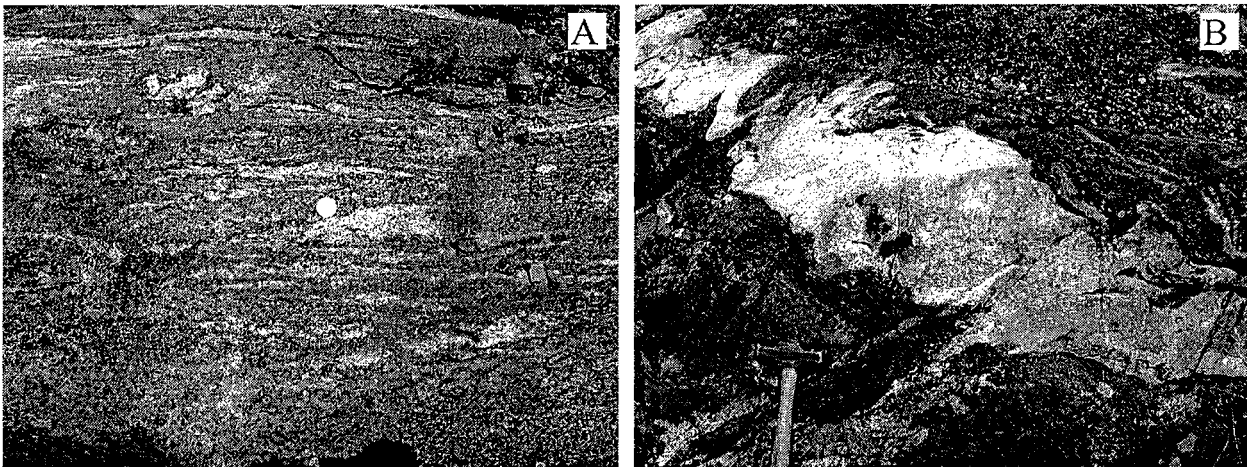


Figure 14. (A) Stromatic migmatite at Stop 5 (view is to the NE, down-plunge of the mineral lineation). Paleosome (gray rock) is Grt+Bt+Qtz+Pl+Sil schist with discrete and sub-concordant lenses and elongate pods of granitic rock in leucosomes (light gray) bound by thin concordant melanosomes composed of Bt+Grt+Sil (dark gray). The residual composition and texture of the paleosome suggests melt loss after partial melting of a pelitic protolith. (B) Sub-concordant granite sheet (Canada Hill granite?) in residual composition diatexite (Bt+Grt+Sil schist). The composition and textures of the diatexite are consistent with extensive melt loss from these rocks after advanced partial melting of a pelitic protolith.



## STOP 5: STROMATIC MIGMATITE, RESIDUAL DIATEXITE AND SHEETS OF CANADA HILL GRANITE.

In general, this exposure is typified by stromatic migmatite with local zones of residual diatexite in metaturbidite (Fig. 14A and B), presumably similar in protolith to that of rocks at Stop 1. The layered structure of this rock is moderately NE-dipping with a penetrative, yet variably oriented, moderately NE-plunging mineral lineation. Sub-concordant leucogranite sheets are penetrative, with one ~0.5 m-thick sub-concordant sheet at the structural bottom, and one at the structural top of the exposure. In views perpendicular to the main fabric (best at the top and south-facing parts of the exposure, the typical tripartite structure of the stromatic migmatite is seen where the rock is dominated by the Bt+Grt-rich host (melt-depleted) to cm-scale concordant lenses and sub-concordant elongate pods of Pl+Qtz leucosome bound and draped by mm-scale concordant Bt+Grt+Sil melanosomes. Locally, these rocks have distinctive cm-scale layers of Gr+Py+Grt with subordinate Bt+Qtz+Kfs+Pl+Sil lenses. The residual diatexite in this outcrop is defined by Grt+Bt+Pl+Sil (Grt is cm-scale) in apparently melt-depleted zones of the outcrop, typically associated with cm-scale granite sheets and pods. The diatexite shows ubiquitous apparent disruption of the host-rock structure, typically marked by an abundance of granitic material. Locally, the diatexite is apparently melt-enhanced (relatively more Pl+Qtz in the matrix) coincident with the most disrupted zones of the exposure, consistent with melt-enhanced granular flow of this rock during deformation. Each of these zones are delimited by a aggregate of cm-scale Grt and Bt melanosomes consistent with melt flow at the grain scale. Meter-scale dikes and sills of similar leucogranite to that found in this exposure are found in other exposures in the vicinity of West Point to the north may represent melt-escape structures to the melt lost from this rock.

- 39.1 Return to vehicles and turn **RIGHT** onto NY 9W South. (Unless you want to make an illegal move by cutting across dirt strip directly into the northbound side traffic circle!). **U-TURN** at “false entrance” to the Bear Mtn Park at the large brown sign just before traffic light. Proceed back on NY 9W North toward Bear Mtn Circle.
- 39.5 Bear Mountain Circle. Proceed ½ of the way around and follow NY 9W North.
- 39.6 Bridge over deep canyon of the Popolopen Creek (“The Hell Hole”) and views of Anthony’s Nose and the Hudson River.
- 41.3 Beginning of ~1 mile long series of roadcuts on both sides of the highway in migmatite with prominent lenses and pods of leucogranite sheets (Canada Hill Granite).
- 42.2 Exit right following signs for NY 218 (West Point, Highland Falls). At top of exit ramp, make a very hard **LEFT** on 218. Cross the bridge over NY 9W.
- 42.3 Park on the right side where a very wide shoulder exists at the entrance to James I. O’Neill High School. Walk down entrance ramp to large roadcuts on both sides. Mind the traffic at this stop.

## STOP 6: STROMATIC MIGMATITE AND CONCORDANT SHEETS OF CANADA HILL GRANITE

Large roadcuts along exit/entrance ramps on the west side of the NY Rte 9W expose excellent examples of migmatitic metapelites and several, thin (5-10 m) interlayered sheets and pods of distinctively blue-gray to white, leucocratic Canada Hill Granite at the northeastern margin of the Crystal Lake Pluton (Figs. 15 and 16A) (Heleneck and Mose, 1976; 1984). This outcrop was part of a Hudson Highlands fieldtrip (Trip B-1, Stop #5) run during the 48<sup>th</sup> NYSGA in 1976 (Heleneck and Mose, 1976). It is also from this locality that Aleinikoff and Grauch (1990) obtained a conventional, multigrain, TIMS U-Pb zircon crystallization age of 1010±6 Ma for the Canada Hill Granite and 1010±4 Ma for thin leucosome from the migmatitic metapelite.

The Canada Hill Granite at this locality is a coarse- to very coarse-grained equigranular rock that is composed of quartz, white K-feldspar, and white to gray plagioclase in roughly equal proportions (Fig. 16B). Locally, Pl dominates this rock. Biotite is ubiquitous as the mafic phase with accessory amounts of sphene, apatite, and zircon. Garnet is locally abundant, especially near contacts with the enclosing migmatites and is interpreted to represent undigested xenocrysts from those rocks after prolonged anatectic erosion of migmatite schollen. The host rock to the granite is stromatic migmatite with a strongly residual Bt+Grt+Pl+Qtz+Sil composition. Locally, the migmatite is schollen in the granite sheets with apparent anatectic erosion at the edges. The structure is regional (N30E, moderate SE-dip). Contacts between the migmatite and granite are typified by irregular structure and aggregates of Grt+Bt (cm-scale garnet), 1-3 cm thick. This is best seen in the exposure on the outside of the ramp (Figs 16A and B). Exposures in the median of the ramp show folded stromatic migmatite. The granite is predominantly massive textured with only local development of a weak foliation associated most commonly at contacts with migmatite host



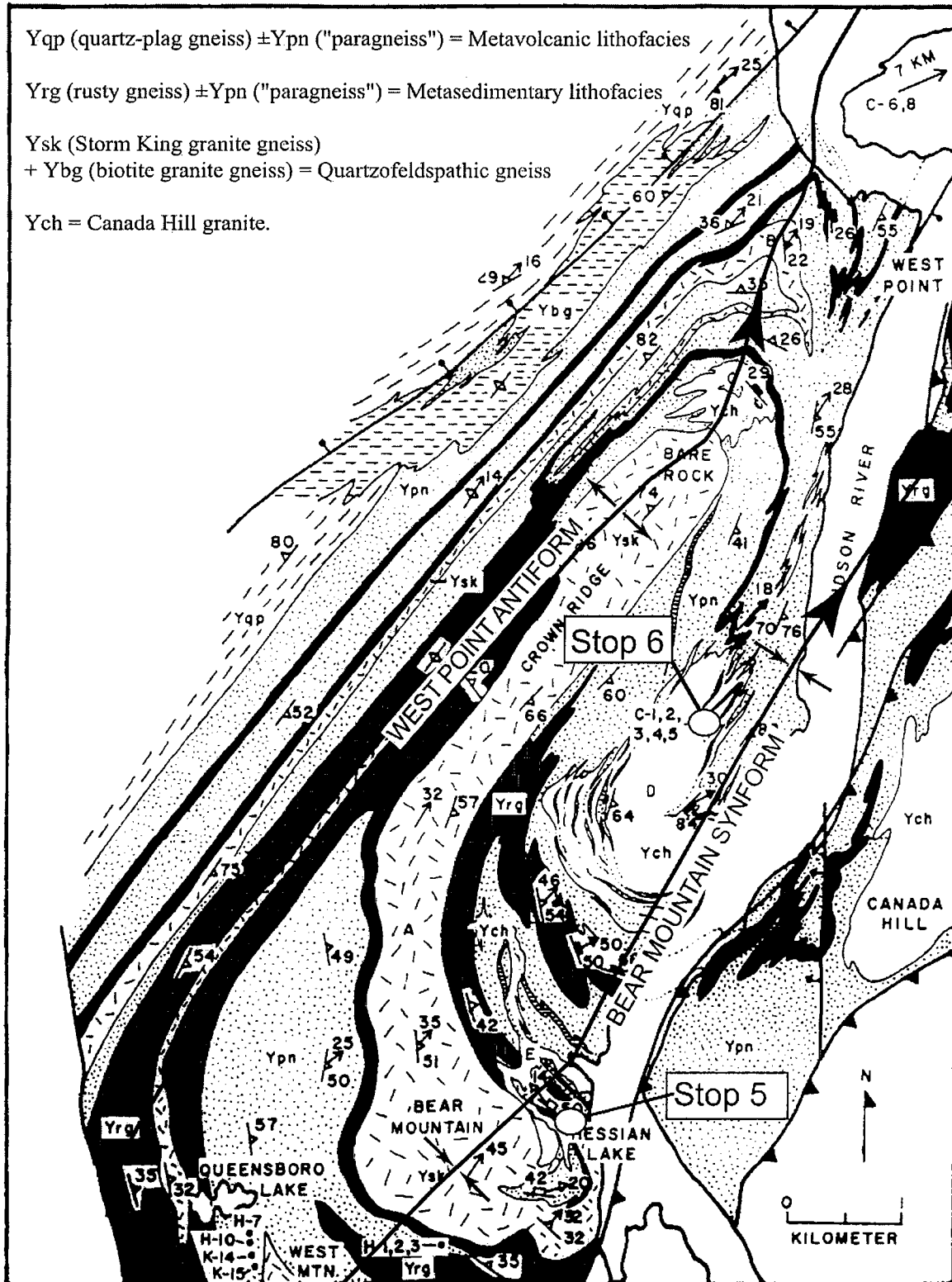


Figure 15. Geologic map of West Point - Bear Mountain area modified from Helenek and Mose (1984). Stop 6 is on the northeastern margin of the Crystal Lake pluton (labeled D) of the Canada Hill granite. Explanation and correlation of rocks units with the nomenclature presented here is given above.

rocks. The Canada Hill Granite here is in sheet and pod form, concordant with the foliation in the surrounding moderately SE-dipping migmatite (~N30E, 60SE). Contacts are generally gradational and diatexitic, except locally where the granite truncates foliation in the migmatite. This is best seen at south end of the northern median outcrop where granite sheets cut foliation near the hinge zone of complexly folded stromatic migmatite. Similar crosscutting relations have been documented elsewhere in the Hudson Highlands. For example, Helenek and Mose (1984) and Ratcliffe (1992) observed the Canada Hill Granite cutting gneissic fabric elements within the Storm King granite gneiss, metapelitic, and metavolcanic rocks in the Popolopen Lake and Oscawana Lake quadrangles to the south and east. In the Bear Mountain area, Helenek and Mose (1984) found xenoliths of Storm King granite gneiss enclosed in Canada Hill Granite along the southern margin of the Brooks Lake Pluton. Helenek and Mose (1984) also noted a weak axial planar foliation and the location of Canada Hill Granite in the hinges of large-scale, broad, open, upright, plunging folds (e.g., Bear Mountain synform, West Point antiform, Fig. 15) and interpreted this to indicate that the Canada Hill Granite was syntectonic to and slightly deformed during a late-stage Grenville orogenic event. This event is constrained to be syn- or post-1010 ±6 Ma, based on the current U-Pb zircon ages of the Canada Hill Granite (Aleinikoff and Grauch, 1990).

END OF TRIP - Return to vehicles and head back to Reeves Meadow Visitor Center

- 45.2 Proceed down entrance ramp onto NY 9W South. At Bear Mountain Circle follow signs for NY Rte 6 West.
- 47.6 Take Exit 18 and continue to follow NY Rte West.
- 47.9 Proceed ½ of the way around traffic circle and follow signs for Seven Lakes Drive.
- 51.6 Lake Tiorati Circle. Proceed ½ way around and continue to follow Seven Lakes Drive.
- 54.9 Kanawauke Circle. Proceed ½ way around and continue to follow Seven Lakes Drive.
- 60.5 Turn LEFT into parking lot at Reeves Meadow Visitor Center.

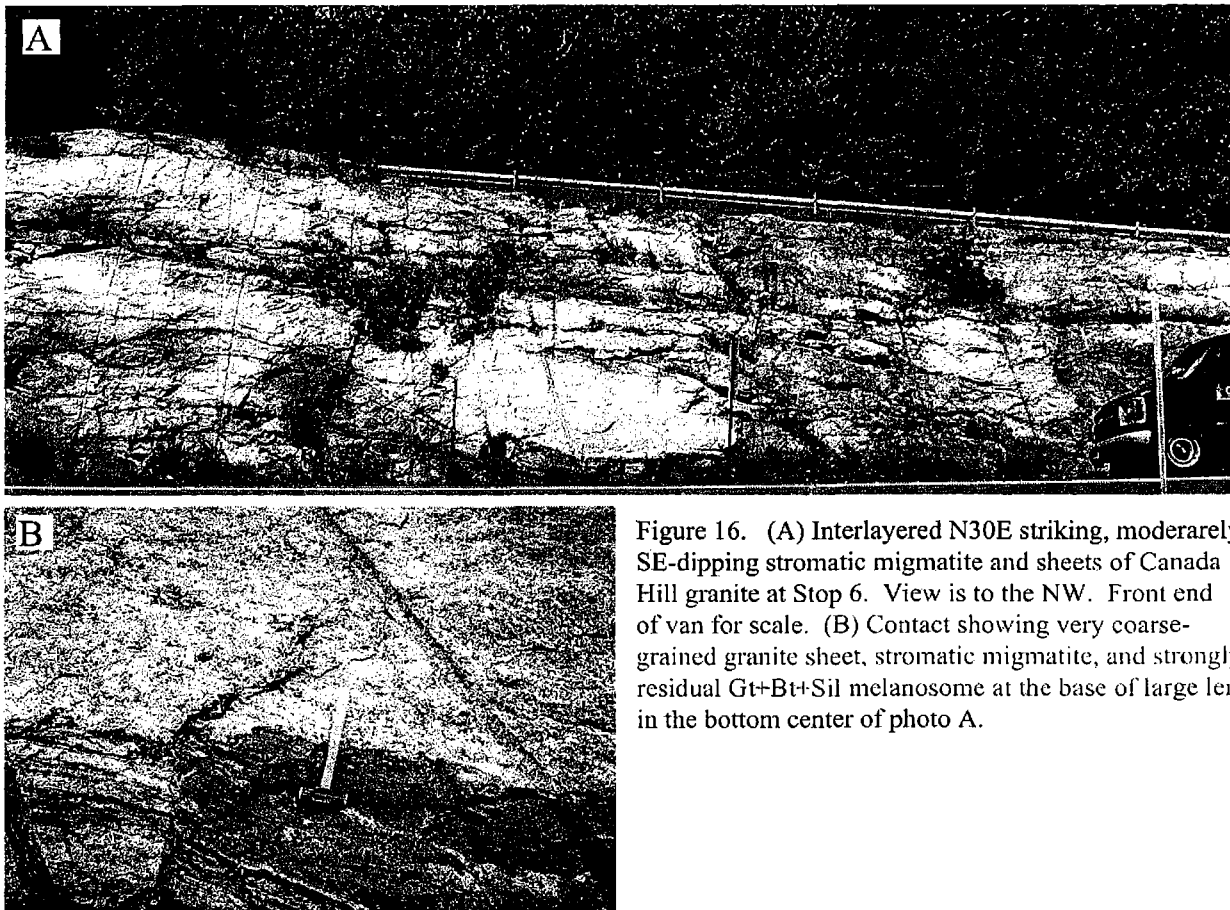


Figure 16. (A) Interlayered N30E striking, moderately SE-dipping stromatic migmatite and sheets of Canada Hill granite at Stop 6. View is to the NW. Front end of van for scale. (B) Contact showing very coarse-grained granite sheet, stromatic migmatite, and strongly residual Gt+Bt+Sil melanosome at the base of large lens in the bottom center of photo A.

